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Effects of wildfire and permafrost on soil organic matter and soil climate in interior Alaska

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Abstract

The influence of discontinuous permafrost on ground-fuel storage, combustion losses, and postfire soil climates was examined after a wildfire near Delta Junction, AK in July 1999. At this site, we sampled soils from a four-way site comparison of burning (burned and unburned) and permafrost (permafrost and nonpermafrost). Soil organic layers (which comprise ground-fuel storage) were thicker in permafrost than nonpermafrost soils both in burned and unburned sites. While we expected fire severity to be greater in the drier site (without permafrost), combustion losses were not significantly different between the two burned sites. Overall, permafrost and burning had significant effects on physical soil variables. Most notably, unburned permafrost sites with the thickest organic mats consistently had the coldest temperatures and wettest mineral soil, while soils in the burned nonpermafrost sites were warmer and drier than the other soils. For every centimeter of organic mat thickness, temperature at 5 cm depth was about 0.5 °C cooler during summer months. We propose that organic soil layers determine to a large extent the physical and thermal setting for variations in vegetation, decomposition, and carbon balance across these landscapes. In particular, the deep organic layers maintain the legacies of thermal and nutrient cycling governed by fire and revegetation. We further propose that the thermal influence of deep organic soil layers may be an underlying mechanism responsible for large regional patterns of burning and regrowth, detected in fractal analyses of burn frequency and area. Thus, fractal geometry can potentially be used to analyze changes in state of these fire prone systems.

Keywords: black spruce, boreal, carbon, combustion, fire emission, fire, fractal analysis, ground fuel, moss cover, organic mat, organic soil, peat, permafrost, soil, surface temperature

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Introduction

Boreal forests store significant amounts of organic carbon, the majority of which resides in the 'belowground' moss, litter, and organic soil layers (FAO, 1988). The boreal forest ecoregion also is underlain by discontinuous permafrost (Van Cleve *et al*., 1986), which has shifted during the late Holocene (Zoltai, 1995). More recent changes in air temperatures (Krupnik & Jolly, 2002), soil temperatures (Osterkamp & Romanovsky, 1999; Clow & Urban, 2002), and water balance (Oechel *et al*., 2000) in northern latitudes raise the issue as to whether such large carbon stocks may become vulnerable to fire and decomposition in ways that are unprecedented for the recent past. The fate of organic soil layers in a changing environment is of particular importance because organic soils (1) often reside at the interface between the water table and vegetation and, therefore, influence plant-available water, (2) provide thermal insulation between the atmosphere and deeper soils and therefore exert some control on seasonal thaw, (3) provide nutrients during the thaw season, whether directly through nutrient cycling within the organic soils or indirectly through thawing and providing access to underlying mineral soils, and (4) provide as ground fuels a significant portion of fuel to the large

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wildfires that typify boreal forests around the globe (French *et al*., 2004).

Within the discontinuous permafrost zone of Alaska, ecosystem types range widely from forest stands dominated by Aspen (*Populous tremulodes*), White Spruce (*Picea glauca*), and Black Spruce (*P. mariana*) to wetlands and peatlands dominated by terrestrial, semiaquatic, or aquatic herbaceous plants and mosses (e.g. *Sphagnum*, *Hylocomnium*, *Pleurozium*). Soil drainage and its topographic and geologic controls (Reiger *et al*., 1979) provide a broad regional control on forest and ecosystem type (Harden *et al*., 2003).

The association of permafrost with vegetation and soil type is dynamic and complex. For example, recovery from fire can be relatively rapid with relatively small changes in the active layer or vegetation type (Harmon *et al*., 1986 for poorly drained soils) or can take decades, with deep thaw of permafrost and a complex trajectory of revegetation (see for e.g. Harmon *et al*., 1986; Viereck *et al*., 1986; O'Neill *et al*., 2002). Furthermore, carbon stocks across boreal landscapes and in organic soils in particular are distributed across a mosaic of soil drainage, fire history, permafrost history, and vegetation recruitment (Turetsky *et al*., 2005). Therefore, it is difficult to forecast the net effects of a changing climate without investigating interactions between physiognomy, biological activity, and fire behavior. Developing associations among soil drainage (which depends upon permafrost distributions, available water, and sediment and soil characteristics), fire occurrence and severity, and vegetation may help us understand the origin and potential fate of organic soils in these regions. Such associations between ecosystem processes and physical factors in the past have led to the accumulation of large carbon stocks in boreal ecosystems. It is important, however, to understand how these relationships are changing with the rapid climate warming currently occurring in the boreal region, and how these changes will influence organic matter accumulation in the future.

Here, we examine soil organic matter storage, combustion losses, temperature, and moisture across sites varying in burning (burned in 1999 vs. unburned) and the presence of permafrost (shallow permafrost vs. no permafrost). Our goal is to explore potential interactions among permafrost and burning in controlling the amount and chemistry of organic soils, and in controlling postfire recovery of near-surface soil temperature and moisture.

Methods

The sites included in this study are located on Fort Greeley Military Base near the town of Delta Junction, AK (Fig. 1). In July 1999, the Donnelly Flats wildfire burned over 8000 hectares (20 000 acres) of black spruce forest. At the time of the fire, wind velocities were 15–20 miles h^{-1} and daytime temperatures were 24 °C (75°F). The buildup index moisture code for ground fuels was 115 (Kasischke & Johnstone, 2005; Turetsky *et al*., 2005) during the days preceding the fire, a value that is ranked as 'extreme' by the Alaska Fire Service. The year 1999 was an average fire year for Alaska with respect to burn area, with 426 000 hectares burned as compared with the long-term average of 392 000 ha yr^{-1} for the period 1950–2005. This fire was one of 77 fires that burned in Alaska that year, and the Donnelly Flats fire was one of 12 fires that burned more than 8000 ha. By comparison, 123 fires ($>$100 acres) burned in 2004 in Alaska, 40 of which were larger than 8000 ha (Alaska Fire History database, available at http://agdc.usgs.gov/data/blm/fire/). The Donnelly Flats fire burned areas with and without subsurface permafrost. Four sampling sites were chosen to compare burned (B) and unburned (U) areas on both permafrost (P) and nonpermafrost (NP) regimes (Fig. 1).

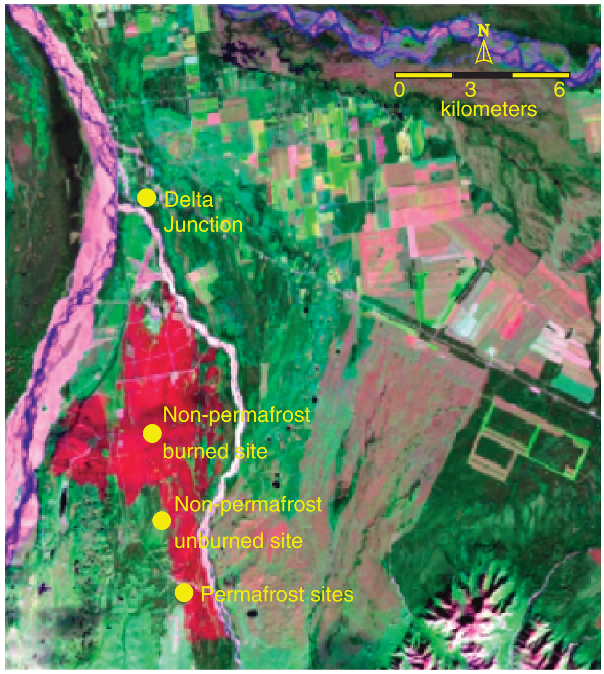

Fig. 1 A Landsat 7 ETM image showing the location of the study sites. About 10 plots plus 5–8 soil pits were studied within or near each dot. The area in red represents the location of the 1999 Donnelly Flats fire.

Using unburned sites as controls for wildland burns is an imperfect experiment because of the natural tendency for burns to terminate at wetter sites that are genetically different from the burn (O'Neill *et al*., 2003). Therefore, in both permafrost and nonpermafrost areas, site selection for soil and plot characterization was used to maximize similarities in soil drainage and vegetation between the burned and unburned site pairs. Sites were selected to maximize the distance from the base of the hill (permafrost site) or stream (permafrost site) (Fig. 1). Sites were assessed for tree density and, where recognizable, for moss type and understory in order to maximize similarities between burned and unburned pairs.

In the nonpermafrost areas, organic soils at both burned and unburned sites were underlain by gravelly glacial till and outwash that lacked permafrost in the upper 2 m of the surface within weeks of the fire (S. Chapin, S. Chambers, personal communication, 2000). These well drained sites are referred to as the nonpermafrost sites (NP) and correspond with the eddy covariance tower sites (DFTC and DFTB) used to measure continuous CO_2 (Liu *et al*., 2005) and a number of environmental and isotopic parameters (Manies *et al*., 2004; Neff *et al*., 2005). The unburned nonpermafrost site is vegetated by black spruce and a number of upland mosses (*Hylocomnium*; *Aulocomnium*) and lichen. In 2003, the burned, nonpermafrost site was vegetated by grasses (*Calamagrastis*) and colonizing mosses (*Polytrichum*; *Ceratadon*; see Harden, unpublished data). The unburned (NP-U) site (stand age >60 years) (Fig. 1) is located about 4 km south of the burned site (NP-B).

In the sites underlain by permafrost, moderately drained soils were typified by 0.3–1.0 m thick loess overlying glacial till and outwash (Pewe & Holmes, 1964; Neff *et al*., 2005). In August of 2000, permafrost was noted within the upper meter of the surface, but in subsequent years the active layer thickened to >1 m. The unburned permafrost site contained black spruce and a variety of upland mosses (*Hylocomnium*; *Sphagnum*; *Rhytidium*) and the burned site in 2003 contained colonizing mosses (*Polytrichum*; *Ceratadon*) and sparse grass (*Calamagrastis)*. These moderately drained sites are referred to as permafrost sites (P). The unburned site (PU; stand age $\sim$115 years) exists as a linear patch at the base of a hilly moraine and is separated from the burn [permafrost burned (PB)] by a small, ephemeral creek. The sampling transect for the unburned (permafrost) site was located halfway between the base of the hill and the ephemeral creek in order to avoid any naturally occurring wet areas.

Soil properties were measured along a 200 m long transect at each of the four sites (PB, NP-B, PU, NP-U). Five to eight pits were excavated at 40 m intervals for soil sampling using a variety of coring and volumetric harvesting techniques (Manies *et al*., 2004). Soils were described according to methods of both Soil Survey Staff (1951) and Canadian Agricultural Services Coordinating Committee (1998). Subsequently we coded soil horizons according to a simple nomenclature following the approach of Manies *et al*. (2004). Common terms include:

LN lichen
L live moss
D dead moss (more moss than roots)
F fibric organics (more roots and amorphous material than moss detritus)
M mesic or moderately decomposed organics (amorphous organics)
H humic or highly decomposed organics (amorphous, unrecognizable organics).

Samples were collected in depth increments of 5–20 cm depending on horizon characteristics and homogeneity of material. In general, depth increments for sampling were smaller (e.g. 5 cm or less) in surface organic layers and larger (20 cm) at depth. Chemical analyses for soils include total carbon and nitrogen using a Carlo Erba NA1500 elemental analyzer, which in the absence of inorganic carbon represents total organic C and total N. Stable isotopes of carbon were analyzed by infrared mass spectrometry. Concentrations of several other elements were determined after acid digestion of whole sample and subsequent analysis by inductively coupled plasma-atomic emission spectrometry. All methods are described in more detail by Manies *et al*. (2004).

Adjacent to our sampling transect which was destructive of vegetation, we established 10 nondestructive plots of $\sim 1\,m^2$ size which were spaced at 10–20 m intervals along transects. Using a 2 cm diameter core at the corner of each plot to minimize destruction, depths of soil layers were noted at each plot. In 2003, soil temperatures were recorded in half-hour increments using HOBO temperature probes (ONSET computers, Hobo professional series, external probe) that were placed at 5 cm depth from the surface (either ground-layer moss, litter or organic soil). We also measured soil moisture content and photosynthetically active radiation (PAR) at each plot weekly or biweekly. Moisture content of the mineral ($<50\%$ organic matter) soil was measured using a HYDROSENSE probe (Campbell Scientific 615, 10 cm length, HydroSense CS620 probe Campbell Scientific, Logan, UT, USA) for volumetric moisture content by placing the probe downward vertically from the top of the mineral soil. A small block of organic mat was replaced after sampling and the same spot was re-measured each time. Based on time domain reflectrometry, these probes were calibrated for a variety of mineral soils by Campbell Scientific, and we used the factory calibration for our sites. PAR was measured at ground level just above the moss using a 60 cm long probe (Apogee Instruments

Inc., Logan, UT, USA) with five sensors mounted in line and leveled; three measurements were made at each plot facing the sun. Relative shading of each plot was estimated by normalizing the data from each plot to $1240\,\mu mol\,m^{-2}/s^{-1}$, the minimum ground-level PAR measured anywhere in 2003.

We calculated organic matter, carbon and nitrogen stocks using concentration data, oven-dried soil bulk density, and the depth of layer or horizon thickness, the products of which are summed to the base of the organic layers. We estimated combustion losses following Harden *et al*. (2004) using aluminum (Al) as the stable constituent, although we found that ash content (quantified by loss on ignition at 550 °C for 5 h) produced similar results. Calculations of fuel include a composite averaging in which replicate unburned soil profiles were combined by multiplying the concentration of each soil constituent (C, bulk density, N, etc.) by the thickness of each soil layer within each unburned profile, summing these values across all soil layers within the profile (successively, starting with Live moss (L) + dead moss (D) + Fibric (F) + $\cdots$), and dividing by the profile's total thickness (for more information, see Harden *et al*., 2004).

We explored the effects of permafrost on fire-related parameters (including ground-fuel storage, combustion loss and chemistry, and postburn fuel and chemistry) using two-sided *t* tests. We also examined the influence of permafrost, burning, and permafrost × burning interactions on a selection of physical measurements indicative of vegetation status (moss cover and shading at ground level) and of soil thermal properties (volumetric moisture content of mineral soil; organic mat thickness; mean annual temperature at 5 cm; July and February average and ranges of temperature) using analysis of variance models and Tukey post hoc comparison of means tests. We used principal components analysis to explore how well these physical variables explained site differences in permafrost and burning. All analyses were performed in Statistica (v. 6, Statsoft Inc.).

As our study was not replicated at the stand level, for example, we did not examine multiple PU stands, we used spatial data of burn area to explore variations in interannual and intraannual burn patterns. We examined relationships between the frequency (e.g. count) and burn area according to fractal geometry relationships (Turner *et al*., 2001). Specifically, we examined frequency–area relationships of the Donnelly Flats fire to other Alaskan fires in 1999 and we compared fires in 1999 to Alaskan fires in 2004 and in all years from 1988 to 2004. Area–frequency plots have been used to assess landscape patterns and to extrapolate those patterns through time or across various landscape scales (Turner *et al*., 2001). Fractal geometry specifically uses the analysis of slopes of a double log plot and is a method for evaluating or describing the relationships of landscape features. Differences in the fractal, or slope of log frequency vs. log area, are noted as evidence for a change. In our case, we examine fractal differences between relatively wet years such as 1999 and relatively dry years such as 2004. Resilience, or the ability of a system to maintain ecosystem structure and behavior in the face of repeated disturbance (Hollings, 1986), is thus the underlying phenomenon for these landscape patterns, with complex interactions among biological, chemical and physical processes being responsible for resilience (Odum, 1971; Turner *et al*., 2003).

Results

Ground-fuels, soil layering, and combustion effects

Organic soil layers, representing ground-layer fuels, occurred in various stages of decomposition (Fig. 2) and generally showed increases in organic matter storage with depth owing to changes in bulk density. Carbon concentrations decreased with depth from 45% in moss layers (L and D horizons) to as low as 10–20% in humic layers (for detailed data see Manies *et al*., 2004). Nitrogen concentrations ranged from 0.35 to $1.72\,g\,m^{-2}$ and were more variable with depth. The unburned permafrost site had thicker organic mats (17 ± 8 cm) than the nonpermafrost site (6 ± 3 cm); however, ground fuels ($g\,m^{-2}$ organic matter, Table 1) within the uppermost moss and fibric layers were similar for the two unburned sites (Fig. 2). Thus, differences in total organic matter stocks (Table 1) were due to deep soil layers (generally M and H layers; Fig. 2) that were more common and/or thicker in the permafrost site (Fig. 2). Similarly, nitrogen stocks were greater in permafrost than nonpermafrost sites owing to differences in the deepest layers (Table 1).

While the burned sites had smaller organic matter stocks than the unburned permafrost or nonpermafrost sites on average, there was large variability in organic matter stocks within sites (Table 1). Burning of ground fuels reduced organic matter and carbon stocks by 40–60% across both sites (differences not significant) and reduced nitrogen stocks by about 30 (permafrost) to 40% (nonpermafrost) (Table 1). This combustion led to carbon emissions of over $1600\,g\,C\,m^{-2}$ and over $35\,g\,N\,m^{-2}$ (Table 1) during this single fire event; these emissions did not vary between the permafrost and nonpermafrost sites (Table 1; P-value = 0.77). Organic matter stocks before and after the burn, however, were significantly different between permafrost and nonpermafrost sites (P-value = 0.02 and P-value < 0.01, respectively, Table 1).

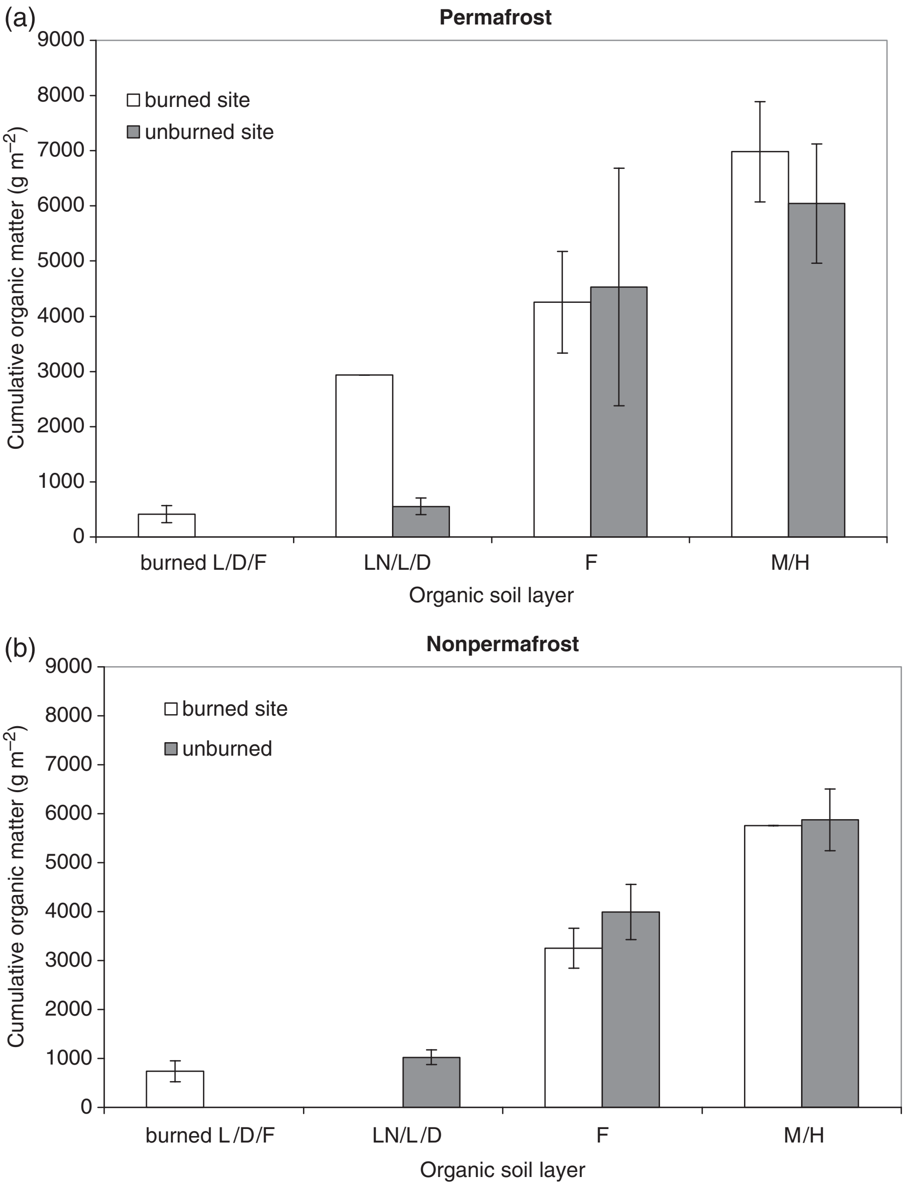

Fig. 2 Organic matter and carbon storage in organic soil layers from surface (left) to deep (right) layers (average ± one SE) for permafrost (a) and nonpermafrost (b) sites. These data were used to calculate ground fuels in the permafrost and nonpermafrost soils. In Table 1. In B, burned and charred layers occur above unburned counterparts. LN, lichen; L, live moss; D, dead moss; F, fibrous; M, moderately decomposed; H, humified. Burned layers shown together whether burned into the L, D, or F layer.

Shading and moss cover

Relative shading, as estimated from PAR measurements at the ground surface (normalized to the highest PAR value), was greater in the unburned sites than the burned sites (Table 2, significant B effect), but there were no significant differences in shading owing to permafrost (Table 2). Interactive effects for shading (Table 2) might be related to tree density differences between permafrost and nonpermafrost sites (5500 vs. 7050 trees ha^{-1}, respectively for the unburned sites; K. Manies, unpublished data). Moss cover also showed a burning effect with interactions between burning and permafrost. Post hoc tests suggest that while burning reduces moss cover, the permafrost site had significantly lower moss cover than the nonpermafrost site during this third year after burning.

Moisture content of mineral soil

Significant interactive effects exist between burning and permafrost for the average moisture content of the

Table 1 Organic matter (OM) consumption and carbon (C) and nitrogen (N) emissions from our permafrost (P) and nonpermafrost (NP) sites (average ± one SD) using tau calculations as described in Eqn (1)

	n	OM ground fuels ($g\,m^{-2}$)	Carbon ($g\,m^{-2}$)	Nitrogen ($g\,m^{-2}$)
Fuel in organic layers				
NP	8	5563 ± 908^{b}	2918 ± 1005^{b}	81 ± 24^{a}
P	5	8636 ± 3005^{a}	4626 ± 1567^{a}	56 ± 57^{b}
Combustion and emission (using Tau_{al})				
NP	6	-3475 ± 1017^{a}	-1653 ± 696^{a}	-35 ± 29^{a}
P	4	-3517 ± 1996^{a}	-1817 ± 1069^{a}	-47 ± 30^{a}
Proportion of ground fuels lost to combustion (by Tau_{Al})				
NP	6	$-62 \pm 18\%^{a}$	$-57 \pm 23.9\%^{a}$	$-43 \pm 36\%^{a}$
P	4	$-41 \pm 23\%^{a}$	$-39 \pm 23.1\%^{a}$	$-30 \pm 19\%^{b}$
Postburn storage in organic layers				
NP	6	3559 ± 1581^{a}	2048 ± 850^{a}	71 ± 30^{a}
P	4	8225 ± 1388^{b}	4476 ± 855^{b}	125 ± 13^{b}

Nonsignificant results are shown with same letter superscripts ($P > 0.05$ in two-sided t-tests of means).

mineral soil. Post hoc results suggest that moisture contents are higher in permafrost sites and lower in burned sites (Table 2), but that burning may not affect permafrost and nonpermafrost landscapes equally. Across all sites, the average moisture content was positively correlated to organic matter thickness of the plot ($n = 29$, $P < 0.05$, $r^2 = 0.84$).

Near-surface temperature

Effects of permafrost and burning on near-surface temperatures also proved to be complex, as indicated by a number of significant but interactive effects (Table 2). February average temperatures were affected by burning, but permafrost sites proved to be cooler than the nonpermafrost sites in unburned areas and warmer in the burned areas. This difference could be related to moisture conductance. For example, wet, frozen soils conduct cold air more effectively than dry soils; although we found that mineral layers are wetter in permafrost sites, the moisture content of the organic mat may be altered by burning. Alternatively, snow pack may be different in burned sites without tree canopy, which may contribute to the interactive effects. Although an interaction was found when examining July average temperatures, we found warmer temperatures at the unburned permafrost sites and cooler temperatures at the burned permafrost sites. Moisture content of the mineral and organic soils, snow pack, and shading are likely confounding the effects of permafrost on temperatures.

More clear relationships of permafrost and burning are evident for mean annual soil temperature (MAT) and diurnal range of temperature for July. The effect of burning was evident as an increase in MAT and in July average temperatures, an effect noted by a number of investigators (Dyrness & Norum, 1983; Kasischke *et al.*, 1995; O'Neill, *et al.*, 2002; Yoshikawa *et al.*, 2003). The effect of permafrost was also evident from a decrease in MAT. July temperature range is significantly greater in permafrost sites, where higher moisture contents likely contribute to high thermal conductance during summer months.

Overall, the effect of burning at the nonpermafrost sites (NU vs. NB) was to increase the mean annual and summer temperatures, with no significant effect on diurnal temperature range in either summer or winter. The annual amplitude, or difference in summer vs. winter means, appears also to have increased slightly as a result of burning (Fig. 3, dashed lines). The effect of burning at the permafrost sites (PU vs. PB) was to increase the mean annual temperature by about 2° and to increase both summer and winter temperatures. Spring and fall months, however, are less affected by burning at the permafrost site (Fig. 3).

Based on the entire data set, with all site data, July diurnal temperature range (averaged for the month and replicated across all sites) predicted organic matter thickness with a coefficient of variation of 6 cm ($P < 0.0015$; $n = 38$ plots):

$$\text{cm organic mat} = 1.38 + 3.71X, \qquad (1)$$

July diurnal temperature range predicted mineral soil moisture within about 10% ($P < 0.0016$, $n = 38$)

$$\text{Volumetric moisture content} = 2.3 + 3.4X. \qquad (2)$$

Correlation and covariance

Many physical properties were highly intercorrelated (Table 3), with one of the highest correlations occurring

Table 2 Physical and thermal surface properties the four Delta sites

	Volumetric moisture of mineral soil $^{+P,B}$ (%)	Organic matter thickness^{P} (cm)	Shade Index^{1+B} (% of maximum)	Moss cover^{+B} (%)	Mean annual Soil temperature^{P,B} (°C at 5 cm)	February average Temperature^{+B} (°C at 5 cm)	July average temperature $^{+P,B}$ (°C at 5 cm)	February temperature range (°C February)	July temp range^{P} (°C July)
Permafrost burned (PB)									
Site mean (SD)	37.55 (3.9)^{a}	13.58 (4.0)^{b}	12.51 (6) %^{a}	5 (6)%^{a}	2.03 (4.81)^{c}	−3.29 (0.72)^{a}	11.76 (1.49)^{b}	0.23 (0.10)^{a}	9.49 (2.9)^{a,b}
n	10	10	10	10	9	9	10	9	10
Permafrost unburned (PU)									
Site Mean (SD)	58.82 (3.5)^{b}	16.97 (7.7)^{b}	66.87 (7)%^{b}	91 (12)%^{b}	−0.04 (0.54)^{a}	−7.47 (1.13)^{b}	10.17 (1.48)^{a}	0.72 (0.65)^{a}	10.72 (3.79)^{b}
n	10	10	10	10	9	9	9	10	9
Nonpermafrost burned (NB)									
Site mean (SD)	17.30 (4.1)^{c}	3.94 (1.9)^{a}	23.11 (11)%^{a}	27 (11)%^{c}	2.55 (0.31)^{c}	−5.68 (0.83)^{c}	14.38 (0.78)^{c}	0.39 (0.14)^{a}	6.22 (1.76)^{a}
n	10	10	10	10	10	10	10	10	10
Nonpermafrost unburned (NU)									
Site Mean (SD)	30.37 (8.5)^{d}	5.71 (2.5)^{a}	62.59 (14)%^{b}	67 (19)%^{d}	1.05 (0.68)^{b}	−4.00 (0.83)^{a}	9.64 (1.21)^{a}	−0.056 (1.54)^{c}	7.80 (3.00)^{a,b}
n	10	10	10	10	8	9	9	9	9

Temperatures are for 5 cm depths below organic soil surface, generally in organic substrate for all but the NB site.
VMC, volumetric moisture content; O, organic soil layers above mineral substrate; diurnal, diurnal temperature range, calculated as average of daily temperature range for replicate plots within each site; *n*, number of replicate plot measurements within the site. Different letter superscripts indicate significant differences ($P < 0.05$ and commonly < 0.02) based on Tukey postoc comparison of means tests from factorial analysis of variance shown in Table 3. A plus sign (+) denotes significant interactive effects between burning and permafrost. P or B denotes significant effects for permafrost or burning based on Wilk's lambda.

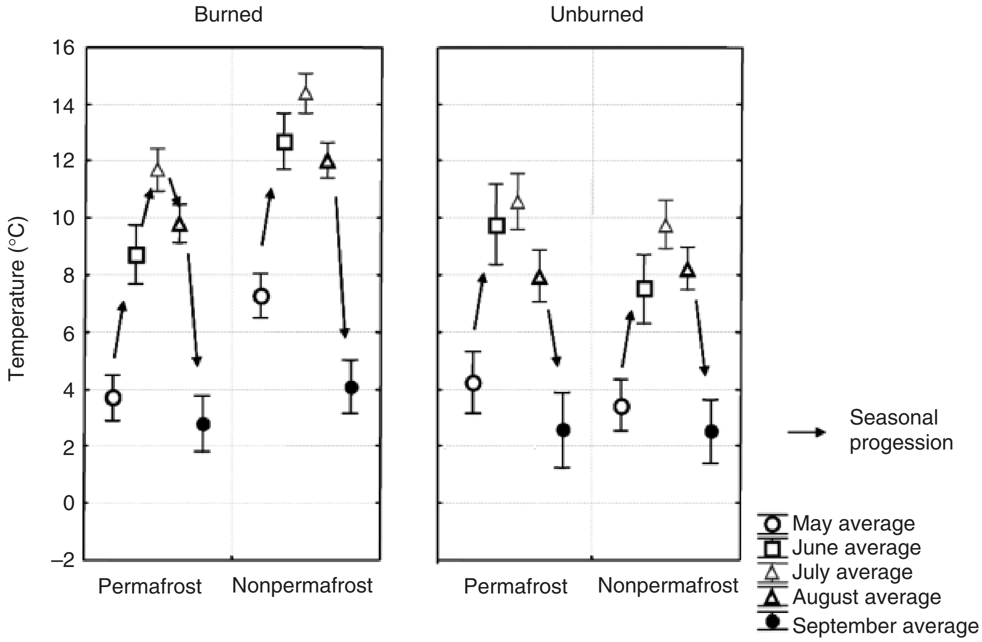

Fig. 3 Effects of burning and permafrost on surface temperature. Least square means for temperatures at 5 cm depth. Please see Table 2 for post hoc tests of differences and Table 3 for interactive effects between permafrost and burning. Vertical bars denote 95% confidence intervals.

Table 3 Correlation coefficients for all plots at all four sites

	Moss Cover	Average mineral moisture	July average temperature	July diurnal temperature	February average temperature	February diurnal temperature	Mean annual temperature	O thickness	Shade index
Moss cover	—	**0.44**	**−0.49**	0.20	**−0.50**	0.24	**−0.70**	0.15	**0.87**
Mineral moisture		—	**−0.57**	**0.50**	**−0.40**	−0.04	**−0.77**	**0.73**	**0.46**
July average temperature			—	0.03	−0.12	−0.14	**0.70**	−0.22	**−0.65**
July diurnal temperature				—	−0.24	0.13	−0.27	**0.52**	0.18
February average temperature					—	0.02	0.54	−0.22	**−0.39**
February diurnal temperature						—	−0.15	−0.12	0.28
Mean annual temperature								**−0.39**	**−0.76**
O thickness								—	0.14
Shade index									—

Significant correlations (>95%) are in bold. Sample numbers range from 29 to 36. Temperature data based on hourly measurements at 5 cm depth. Moisture data based on biweekly measurements in uppermost mineral soil. Shade index based on ground-level photosynthetically active radiation (PAR) normalized to brightest measurement 1240 $\mu mol\ s^{-1}$

between minimum daily surface temperature in July and organic layer thickness (Fig. 4). Combining all plots from all sites, correlations among many variables were significant (Fig. 4) and differences in thickness among sites explains why temperatures are different (Fig. 4, inset). Only within the NU site, where thickness varies greatly (Table 2), were temperature/thickness correlations significant. Overall, the daily minimum temperatures decreased by about 0.4–0.5 °C for every centimeter of organic mat during the summer (June–August) and by about 0.2 °C for every centimeter during May and September (Fig. 4). Daily average temperatures also decreased with organic mat thickness (data not shown), in summer months by about 0.20–0.27 $°C\ cm^{-1}$ thickness, and by about 0.12–0.17 $°C\ cm^{-1}$ during the shoulder seasons of May and September. Spatial variations in near-surface soil temperatures varied among these four sites by about 6 °C in summer (from 16 to 10 °C) and by about 2° in May and September (from 4 to 2 °C, Table 2) according to their permafrost and burn

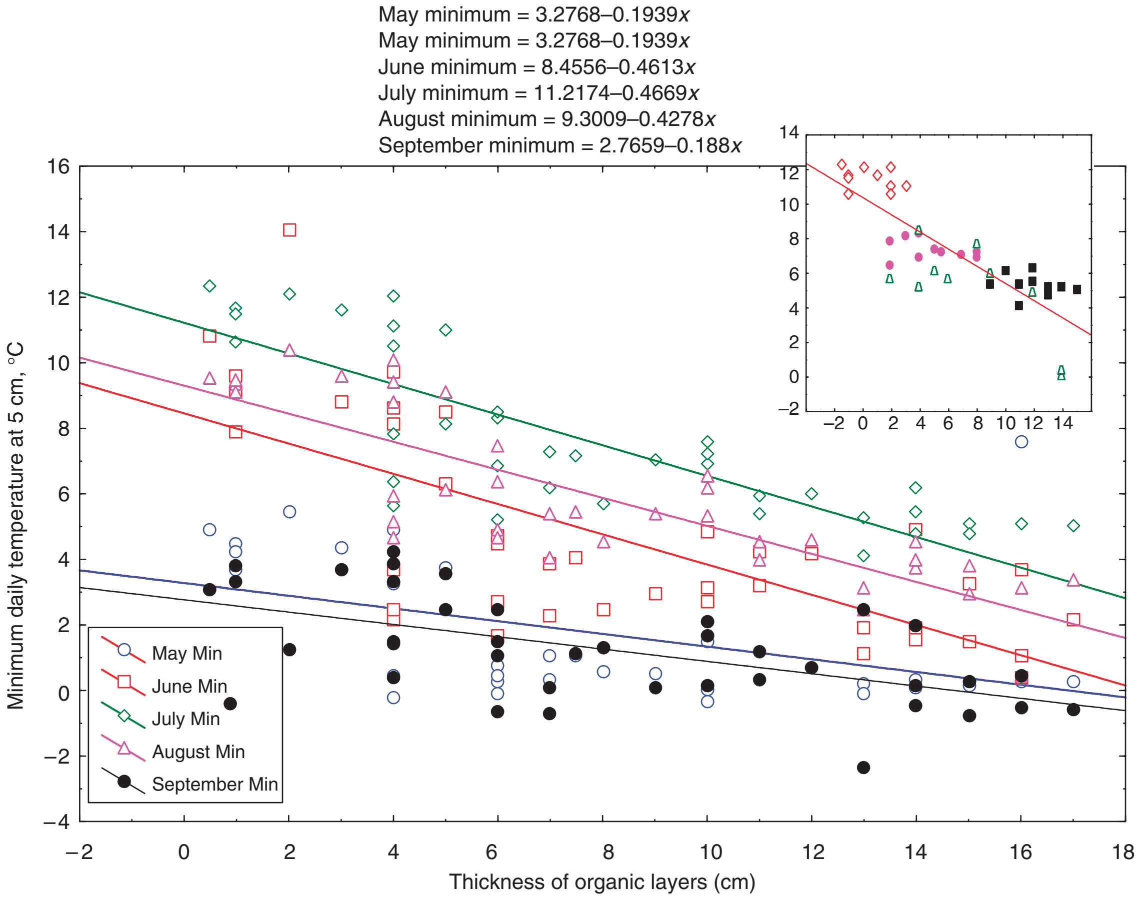

Fig. 4 Minimum daily surface (5 cm) temperatures in relation to thickness of organic layers. Data are from 10 replicate moss plots at each of the four sites during May to September 2003. Temperatures are averaged from daily minima based on hourly measurements. Equations for lines shown at top. Equation P-values ≤ 0.05. Insert only, July measurements showing sites separately – open red diamonds (NB), closed pink circles (PB), closed black squares (PU), and open green triangles (NU).

histories. We also examined the data using principal components analysis (data not shown) because so many of the physical and biological variables were correlated. We found that almost half of the variance among samples is described by Factor 1, which is comprised mainly of July, February, and mean annual temperature vs. biological attributes of shading and moss cover. The principal Factor 1 separated burned and unburned effects most effectively. Factor 2 accounted for 17% of the variance and was correlated with moisture and organic mat thickness vs. biological attributes of shading and moss cover. July and February temperature ranges were at opposite ends of the Factor 2 plane, suggesting that they help characterize differences between permafrost and nonpermafrost landscapes.

Frequency–area analysis of fires

The Donnelly Flats fire was typical in size per area for large fires that burned in 1999 in Alaska (Fig. 5, inset showing Donnelley Flats fire). Differences in fire behavior for small vs. large fires in 1999 are apparent from their difference in slopes (Fig. 5). The more pervasive burning of 2004, which was an extremely long and areally significant fire year, is evident in both area and number of fires (Fig. 5), and the less steep slope indicates that the shift is due to an increase in large fires. For the year 2004, slopes for small and large fires are more similar than they are for other years. Most historic fires in Alaska between 1988 and 2004 (Fig. 5, all years) have a more similar frequency/area relationship to the 1999 fires than they do to the 2004 fires.

Discussion

Permafrost and burning effects found in this study argue strongly for replicate studies at a number and variety of forest stands in the boreal. For example, differences between permafrost and nonpermafrost stands suggest that heterogeneity of discontinuous per-

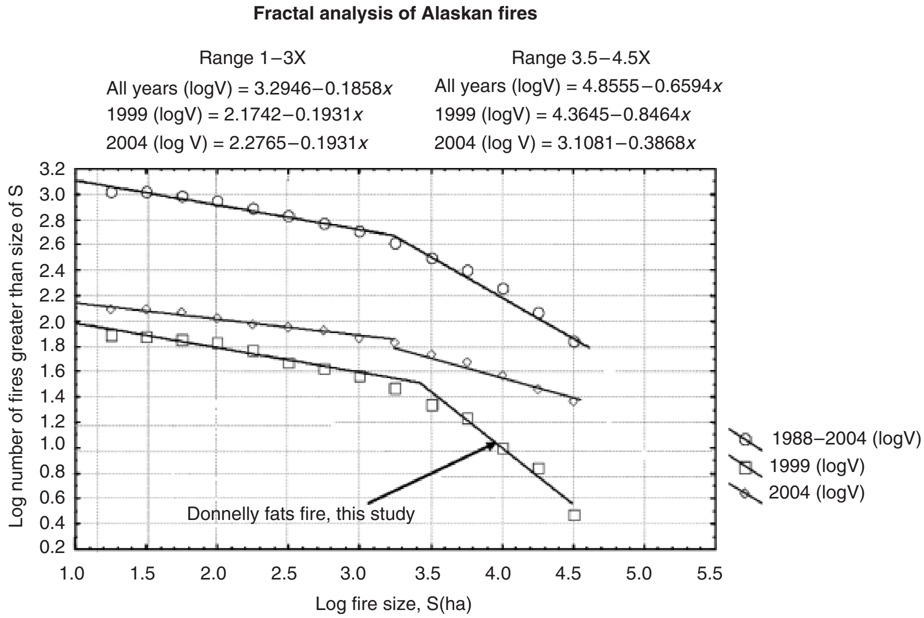

Fig. 5 Frequency–area plot of Alaskan Fires for 1999, 2004, and for all years 1988–2004. Linear fits shown at the top for linear segments of small fires 1–3 log (HA) and large fires 3.5–4.5 log(HA).

mafrost landscapes might be due fundamentally to differences in the deep organic layers that, under normal circumstances perpetuate thermal and biological properties through repeated fire cycles. Differences in permafrost and nonpermafrost sites were not evident from combustion chemistries nor chemistries of the shallow soils in this fire. The similar fuel and carbon stocks in shallow organic layers (LN/L/D, and F layers; Fig. 2) suggests that, under these conditions of burning, the regrowth in shallow layers that occurs after each fire may not be significantly different across permafrost and nonpermafrost sites. Thermal and biological differences, therefore, can be attributed to the deep organic mats on the permafrost sites and to historical effects that formed and sustained the deep organic layers. This theory is consistent with findings of Carrasco *et al.* (2006) in which thick humic layers persist over millennia due to the historic effects of permafrost, cold temperatures and organic matter recalcitrance in subsurface soil horizons. Moisture content and depth to water tables are particularly important to the preservation of these deep organic layers because moisture protects the substrate from losses to fire and because moisture conducts cold winter temperatures into the frozen peat. Based on these ideas we have developed a conceptual model (Turetsky *et al.*, 2005) in which the mineral substrate and its drainage capacity exerts a primary control over near-surface permafrost in black spruce boreal forests; drainage becomes increasingly important to vegetation and microbial processes when permafrost thaws, for example, through fire cycles (Harmon *et al.*, 1986; Yoshikawa *et al.*, 2003) because secondary succession and decomposition are highly sensitive to soil drainage. With the maintenance of poor drainage conditions, either through high water-holding capacity of organic soil layers or characteristics of the underlying sediment (Reiger *et al.*, 1979) deep organic layers in permafrost environments can be kept wet and cool during summer seasons even after burning (Fig. 3). Therefore, deep organic layers must first form from a variety of conditions (aggrading permafrost, moisture conditions that resist fire) and deep organic layers may be required for persistence of permafrost. Once permafrost has aggraded, a positive feedback between cold temperatures, poor drainage, and the chemical recalcitrance of moss layers enhances the preservation of thick, deep organic mats by resisting decomposition and combustion. Changing climatic regimes could interrupt these feedbacks by increasing fire severity or initiating permafrost degradation, both of which might lead to the degradation of deep organic layers. In turn, removal of these deep soils will change soil climatic conditions. Vegetation that colonizes drier boreal systems, such as aspen or white spruce, have higher decomposition rates (Turetsky *et al.*, 2005) and are more responsive to changes in temperatures (O'Neill *et al.*, 2002; Q^{10} lower in black spruce). Thus, changes in vegetation type would accelerate decomposition, further contributing to a loss of organic carbon and to a thinner organic mat.

The importance of the organic mats in the governance of heat is apparent from relationships of surface temperature to organic mat thickness, relationships that have been shown by other workers as well (Yoshikawa *et al*., 2003; Kasischke & Johnstone, 2005). Average annual temperatures and average summer temperatures are inversely proportional to thickness of the organic (O) layers (Table 3; Fig. 4), which suggests that while radiant heat, as affected by shading and albedo (Yoshikawa *et al*., 2003), may contribute to site variation, the organic mat below 5 cm where temperature is measured is governing to a large extent the thermal behavior at the soil surface. Lachenbruch (1994), who explored the relationship of heat penetration through a two-layer medium in similar soils, demonstrated that the range of diurnal temperature fluctuations is highly dependent on properties of both the upper (generally organic) and lower (generally mineral) soil layers and that the depth of temperature fluctuations varies greatly with soil properties such as density and moisture content. Of particular importance to our case might be moisture content of the organic matter, which we found difficult to calibrate for the variety of soil conditions (see Yoshikawa *et al*., 2004; Overduin *et al*., 2005). The influence of the deep humified layers, mostly found in the permafrost sites (Fig. 2), on heat properties is particularly important owing to its high density and moisture retention (Yoshikawa *et al*., 2003). Indeed, the reduction in organic mat thickness by burning was not statistically significant (Table 2) despite the significant differences in stock of organic matter (Table 1) emphasizing again that the presence of dense, deep organic horizons just above the mineral soil (Fig. 2) play an important role in site differences of organic matter stocks and surface temperatures alike.

A spatially explicit understanding of combustion requires that both ground fuels and fractional combustion losses (e.g. burn severity) be defined and monitored separately in order to detect changes in the coupling of climate-fire-carbon cycling in boreal systems. For example, Kasischke & Bruhwiler (2003) demonstrated different fractional combustion losses for various vegetation/thermal regimes, as supported by this study. It is important to note, however, that the 1999 Donnelly Flats fire is not representative of all fires for a number of reasons related to both climatic (e.g. fire weather) and landscape (e.g. fire recurrence, fuel availability) controls that vary throughout the boreal region and from year to year. Other wildfires, for example, may force more variation in emissions across the landscape, especially if differences in surface moisture conditions are great at the time of the fire. As has been shown in numerous studies (Kasischke *et al*., 2000; Yoshikawa *et al*., 2003; Zhuang *et al*., 2003), thermal (e.g. moisture and temperature) conditions that control plant production also have controlling effects on decomposition and ground-fuel storage. Further analyses should be conducted on landscape patterns for a variety of fires under a variety of fire-weather conditions in order to elucidate 'legacy' effects of climate-fire interactions reflected in old and deep organic layers. Further studies should also be conducted on the potential to detect organic mat thickness from surface temperature or soil moisture content from surface temperature. For example, the predictive statistics for organic mat thickness (Equations 1, 2) could be used to map organic mat thickness if factors such as shading were accounted for. While studies have begun to address these complex temporal and spatial questions (e.g. Rupp *et al*., 2001; Turetsky *et al*., 2004), the models that estimate and forecast fire emissions require a clear and spatially explicit link to past climate-fire-soil interactions.

A number of investigators (e.g. Malamud *et al*., 1998; Diaz-Delgado *et al*., 2004) have noted that landscape patterns of wildfire and regrowth occur in a way that is 'self organized' because the landscape has a 'memory' for the distribution of fuels on the landscape. Based on our data herein and on a previous modeling study (Carrasco *et al*., 2006), we hypothesize that the underlying mechanisms responsible for the large regional patterns of burning and regrowth in the discontinuous permafrost zone are due largely to the long-lasting effects of deep, cold environments that persist through fire cycles. This is exemplified by similarities in combustion (Table 1) and differences in pre and postburn stocks of organic matter at depth (Fig. 2) that exert controls over near-surface temperatures (Fig. 4). Moreover, postburn thermal properties maintain the differences between permafrost and nonpermafrost landscapes, particularly in summer months when soil temperature exerts control over plant production and decomposition; with different soil temperatures on permafrost and nonpermafrost sites, biotic responses are likely to maintain those differences through secondary succession (Johnstone & Kasischke, 2005) and plant-soil responses to temperature. It would follow, therefore, that as long as permafrost distributions and wildfire behavior are subjected to 'moderate' fires like the Donnelly Flats wildfire, these landscape patterns are likely to be perpetuated (Fig. 5; separation along Vector 2). Threshold or irreversible change, however, might occur if permafrost degrades or if combustion is severe enough to penetrate to the mineral soil and ameliorate the differences in organic mat thickness. For example, dramatic decreases in organic matter thickness at all sites would presumably manifest convergence of near-surface temperatures of permafrost and nonpermafrost sites (Fig. 4, toward thin organic mats). This is consistent with the work of Johnstone & Kasischke (2005), who found that while long-lasting, legacy effects may persist through multiple fire cycles influence patterns of soil/temperatures and postfire succession, the occurrence of deep burning fires can radically change the successional trajectory.

There is some indication that fractal dimension (slope, Fig. 5) of burn frequency and area may indicate significant system differences between years of high area burn vs. years of low area burn. The Alaska 2004 fires, for example, were much more pervasive than 1999 (6.5 million vs. 1 million acres, respectively; Alaska Fire History database, available at http://agdc.usgs.gov/data/blm/fire/). In addition, 2004 fractals were more similar for small vs. large fires than in 1999, when fractals were markedly different for small vs. large fires (Fig. 5). From the fractal analysis, it appears that severe fire weather ameliorates the differences between small and large fires because the fractals are more similar for the year with severe fire weather. It is possible that burn severity (in our case depth of burn) might be correlated with frequency/area relationships. For example, as the fractal in 2004 is less steep than other years, it indicates a disproportionate increase in large over small fires during that high-burn year, and large fires are thought to burn more severely than small fires (Conard *et al*., 2002). The potential link between burn area and burn depth presents an underlying mechanism (organic mat thickness) that might account for the system-level shifts discussed by Hollings (1986) as indicative of change in ecosystem state.

The possibility of remotely sensing burn severity through the correlation of organic mat thickness to surface temperature (e.g. Fig. 4) may allow both fractal analysis of burn area and frequency in two dimensions plus fire severity in the third dimension. Analyses must consider both fractal analyses of frequency–area relationships, as well as spatial and temporal patterns of organic matter combustion or other measures of burn severity. In particular, seasonal climatic and short-term fire weather conditions should be considered, as they can contribute to patterns of burn severity through relationships to thaw depth. Indeed, climate change and its consequences for permafrost distributions and fuel moisture may decouple aspects of fire behavior (burn area and severity). Further studies should include a spatially explicit analysis of soil, vegetation, burn severity, burn perimeter, and fire weather in a number of stands for a number of years in order to test and monitor these ecosystem properties and their associations.

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References

Canadian Agricultural Services Coordinating Committee (1998) *The Canadian System of Soil Classification*. NRC Canada Research Press, Ontario.

Carrasco J, Neff JC, Harden JW (2006) Modeling physical and biogeochemical controls over carbon accumulation in a boreal forest soil. *Journal of Geophysical Research – Biogeosciences*, **111**, G02004.

Clow GD, Urban FE (2002) Large permafrost warming in northern Alaska during the 1990's determine from GTN-P borehole temperature measurements. *EOS Transactions AGU*, **83**, abstract B11E-04.

Conard SG, Sukhinin AI, Stocks BJ *et al*. (2002) Determining effects of area burned and fire severity on carbon cycling and emissions in Siberia. *Climatic Change*, **55**, 197–211.

Diaz-Delgado R, Lloret F, Pons X (2004) Spatial patterns of fire occurrence in Catalonia, NE, Spain. *Landscape Ecology*, **19**, 731–745.

Dyrness CT, Norum RA (1983) The effects of experimental fires on black spruce forest floors in interior Alaska. *Canadian Journal of Forest Research*, **13**, 879–893.

FAO (1988) *FAO- UNESCO soil map of the world. Revised legend*. FAO/UNESCO/ISRIC World Soil Resources reports, No. 60 (reprinted 1990).

French HF, Goovaerts P, Kasischke ES (2004) Uncertainty in estimating carbon emissions from boreal forests. *Journal of Geophysical Research*, **109**, doi: 10.1029/2003JD003635.

Harden JW, Meier R, Silapaswan C *et al*. (2003) Soil drainage and its potential for influencing wildfires in Alaska. In: *Studies by the US Geological Survey in Alaska, 2001, US Geological Survey Professional Paper 1678* (ed. Galloway J), pp. 139–144. US Geological Survey, Menlo Park, CA.

Harden JW, Neff JC, Sandberg DV *et al*. (2004) Chemistry of burning the forest floor during the FROSTFIRE experimental burn, interior Alaska, 1999. *Global Biogeochemical Cycles*, **18**, doi: 10.1029/2003GB002194.

Harmon ME, Franklin JF, Swanson FJ *et al*. (1986) Ecology of coarse woody debris in temperate ecosystems. *Advances in Ecological Research*, **15**, 133–307.

Hollings CS (1986) The resilience of terrestrial ecosystems: local surprise and global change. In: *Sustainable Development of the Biosphere* (eds Clark WC, Munn RE), pp. 292–317. Cambridge University Press, Cambridge, MA.

Johnstone JF, Kasischke ES (2005) Stand-level effects of soil burn severity on postfire regeneration in a recently burned black spruce forest. *Canadian Journal of Forest Research*, **35**, 2151–2163.

Kasischke ES, Bruhwiler L (2003) Emissions of carbon dioxide, carbon monoxide, and methane from boreal forest fires in 1998. *Journal of Geophysical Research*, **108**, doi: 10.1029/2001JD000461.

Kasischke ES, Christensen NL, Stocks BJ (1995) Fire, global warming, and the carbon balance of boreal forests. *Ecological Applications*, **5**, 437–451.

Kasischke ES, Johnstone JF (2005) Variation in postfire organic layer thickness in a black spruce forest complex in interior Alaska and its effects on soil temperature and moisture. *Canadian Journal of Forest Research*, **35**, 2164–2177.

Kasischke ES, O'Neill KP, French HF *et al*. (2000) Controls on patterns of biomass burning in Alaskan boreal forests. In: *Fire, Climate Change, and Carbon Cycling in the Boreal Forest. Ecological Studies*, Vol. 138 (eds Kasischke ES, Stocks BJ), pp. 173–196. Springer-Verlag, New York.

Krupnik I, Jolly D (2002) *The earth is faster now. Indigenous observations of arctic environmental change*. Arctic Research Consorium of the United States, 376 pp.

Lachenbruch AH (1994) *Permafrost, the active layer, and changing climate*. Open-file Report 94-694, US Geological Survey, 43 pp.

Liu H, Randerson JT, Linfours J *et al*. (2005) Changes in the surface energy budget after fire in the boreal ecosystems of interior Alaska: an annual perspective. *Journal of Geophysical Research – Atmospheres*, **110**, D13101.

Malamud BD, Morein G, Turcott DL (1998) Forest fires: an example of self-organized critical behavior. *Science*, **218**, 1840–1841.

Manies KL, Harden JW, Silva SR *et al*. (2004) *Soil data from Picea mariana stands near Delta Junction, Alaska of different ages and soil drainage types*. Open File Report 2004-1271, US Geological Survey, 19 pp.

Neff JC, Harden JW, Gleixner G (2005) Fire effects on soil organic matter content and composition in boreal Interior Alaska. *Canadian Journal of Forest Research*, **35**, 2178–2187.

Odum EP (1971) *Fundamentals of Ecology*. W.B. Saunders, Philadelphia, PA.

Oechel WC, Vourlitis GL, Hastings SJ *et al*. (2000) Acclimation of ecosystem CO_2 exchange in the Alaskan artic in response to decadal climate warming. *Nature*, **406**, 978–981.

O'Neill KP, Kasischke ES, Richter DD (2003) Seasonal and decadal patterns of soil carbon uptake and emission along an age-sequence of burned black spruce stands in interior Alaska. *Journal of Geophysical Research*, **108**, doi: 1029/2001JD000443.

O'Neill KP, Richter DD, Kasischke ES (2002) Environmental controls on soil CO_2 flux following fire in black spruce, white spruce, and aspen stands of interior Alaska. *Canadian Journal of Forest Research*, **32**, 1525–1541.

Osterkamp TE, Romanovsky VE (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes*, **10**, 17–37.

Overduin PP, Yoshikawa K, Kane DL, Harden JW (2005) Comparing electronic probes for volumetric water content of low-density feathermoss. *Sensor Review*, **5**, 215–221.

Pewe TL, Holmes GW (1964) *Geology of the Mt. Hayes D-4 Quadrangle, Alaska*. US Geological Survey, Washington, DC.

Reiger S, Schoephorster DB, Furbush CE (1979) *Exploratory Soil Survey of Alaska*. Soil Conservation Service, Washington, DC.

Rupp TS, Keane RE, Lavorel S *et al*. (2001) *Towards a classification of landscape-fire succession models* Newsletter of the Global Change and Terrestrial Ecosystems Core Project of the International Geosphere–Biosphere Program, 17, pp. 1–4.

Soil Survey Staff (1951) *Soil Survey Manual*. Agricultural Research Administration, United States Department of Agriculture, Washington, DC.

Turetsky MR, Amiro B, Bosch E *et al*. (2004) Historical burn area in western Canadian peatlands and its relationship to fire weather indices. *Global Biogeochemical Cycles*, **18**, GB4014.

Turetsky MR, Mack MC, Harden JW *et al*. (2005) Spatial patterning of soil carbon storage across boreal landscapes. In: *Ecosystem Function in Heterogeneous Landscapes* (eds Lovett GM, Jones C, Turner MG, Weathers KC), pp. 229–255. Springer, New York.

Turner MG, Collins SL, Lugo AL *et al*. (2003) Disturbance dynamics and ecological response: the contribution of long-term ecological research. *BioScience*, **53**, 46–56.

Turner MG, Gardner RH, O'Neill RV (2001) *Landscape Ecology in Theory and Practice: Pattern and Process*. Spinger-Verlag, New York.

Van Cleve K, Yarie J (1986) Interaction of temperature, moisture,and soil chemistry in controlling nutrient cycling and ecosystem development in the taiga of Alaska. In: *Forest Ecosystems in the Alaskan Taiga* (eds Van Cleve K, Chapin FS III, Flanagan PW, Viereck LA, Dyrness CT), pp. 160–189. Springer-Verlag, New York.

Viereck LA, Van Cleve K, Dyrness CT (1986) Forest ecosystems distribution in the taiga environment. In: *Forest Ecosystems in the Alaskan Taiga: A Synthesis of Structure and Function. Vol. Ecological Studies*, Vol. 57 (eds Van Cleve K, Chapin FS, Flannagan LA, Viereck LA, Dyrness CT), pp. 22–43. Springer-Verlag, New York.

Yoshikawa K, Bolton WR, Romanovsky VE *et al*. (2003) Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. *Journal of Geophysical Research*, **108**, doi: 10.1029/2001JD000438.

Yoshikawa K, Overduin PP, Harden JW (2004) Moisture content measurements of moss (Sphagnum spp.) layers using commercial sensors. *Permafrost and Periglacial Processes*, **15**, 309–318.

Zhuang Q, McGuire AD, O'Neill KP *et al*. (2003) Modeling soil thermal and carbon dynamics of a fire chronosequence in interior Alaska. *Journal of Geophysical Research – Atmospheres*, **108**, doi: 10.1029/2001JD001244.

Zoltai SC (1995) Permafrost distribution in peatlands of west-central Canada during the Holocene warm period 6000 years ago. *Geographie physique et Quaternaire*, **49**, 45–54.