# MECHANISMS OF SOIL CARBON STABILIZATION IN BLACK SPRUCE FORESTS OF INTERIOR ALASKA: SOIL TEMPERATURE, SOIL WATER, AND WILDFIRE

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MECHANISMS OF SOIL CARBON STABILIZATION IN BLACK SPRUCE FORESTS OF INTERIOR ALASKA: SOIL TEMPERATURE, SOIL WATER, AND WILDFIRE

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**Abstract and Overview:**

The likely direction of change in soil organic carbon (SOC) in the boreal forest biome, which harbors roughly 22% of the global soil carbon pool, is of marked concern because climate warming is projected to be greatest in high latitudes and temperature is the cardinal determinant of soil C mineralization. Moreover, the majority of boreal forest SOC is harbored in surficial organic horizons which are the most susceptible to consumption in wildfire. This research focuses on mechanisms of soil C accumulation in recently burned (2004) and unburned (~1850-1950) black spruce (*Picea mariana* [Mill.] BSP) forests along gradients in stand productivity and soil temperature. The primary research questions in these three chapters address: 1) how the interaction between stand production and temperature effect the stabilization of C throughout the soil profile, 2) the quantity and composition of water soluble organic carbon (WSOC) as it is leached from the soil across gradients in productivity and climate, and 3) physiographic controls on organic matter consumption in wildfire and the legacy of wildfire in stable C formation (pyrogenic C, or black carbon). Soil WSOC concentrations increased while SOC stocks decreased with increasing soil temperature and stand production along the gradients studied. Stocks of BC were miniscule in comparison to organic horizon SOC stocks, and therefore the C stabilizing effect of wildfire was small in comparison to SOC accumulation through arrested decomposition. We conclude that C stocks are likely to be more vulnerable to burning as soil C stocks decline relative to C sequestered in aboveground woody tissues in a warmer climate. These findings contribute to refining estimates of potential changes in boreal soil C stocks in the context of a changing climate.

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**Co-authorship by chapter:**

This dissertation is composed of three separate papers with different suites of co-authors. I would like to acknowledge the terrific efforts of these co-authors, by chapter, below.

*Chapter 1:*

Kane, E.S., D.W. Valentine, E.A.G. Schuur, K. Dutta (2005), Soil carbon stabilization along climate and stand productivity gradients in black spruce forests of interior Alaska, *Can. J. For. Res.*, 35, 2118-2129.

*Chapter 2*:

Kane, E.S., D.W. Valentine, G.J. Michaelson, J.D. Fox, C-L. Ping (2006), Controls over pathways of carbon efflux from soils along climate and black spruce productivity gradients in interior Alaska, *Soil Biol. Biochem.*, 38, 1438-1450.

*Chapter 3*:

Kane, E.S., E.S. Kasischke, D.W. Valentine, M.R. Turetsky, A.D. McGuire (submitted), Topographic influences on wildfire consumption of soil organic carbon in black spruce forests of interior Alaska: implications for black carbon accumulation, *Global Biogeochem. Cycles*,

# Chapter 1: Soil carbon stabilization along climate and stand productivity gradients in black spruce forests of interior Alaska

## 1.1: Abstract

The amount of soil organic carbon (SOC) in stable, slow turnover pools is likely to change in response to climate warming because processes mediating soil C balance (net primary production and decomposition) vary with environmental conditions. This is important to consider in boreal forests, which comprise one of the world's largest stocks of SOC. We investigated changes in soil C stabilization along four replicate gradients of black spruce productivity and soil temperature in interior Alaska to develop empirical relationships between SOC and stand and physiographic features. Total SOC harbored in mineral soil horizons decreased by 4.4 g C $m^{-2}$ for every degree-day increase in heat sum within the organic soil across all sites. Furthermore, the proportion of relatively labile light fraction (density <1.6 g $cm^{-3}$) soil organic matter decreased significantly with increased stand productivity and soil temperature. Mean residence times of SOC (as determined by $\Delta^{14}C$) in dense fraction (>1.6 g $cm^{-3}$) mineral soil ranged from 282-672 years. The oldest SOC occurred in the coolest sites, which also harbored the most C and had the lowest rates of stand production. These results suggest that temperature sensitivities of organic matter within discrete soil pools, and not just total soil C stocks, need to be examined in order to project the effects of changing climate and primary production on soil C balance.

**1.2: Introduction**

It is widely accepted that positive feedbacks exist between the increasing atmospheric concentration of $CO_2$ and global warming (e.g., Keeling et al. 1996, Chapin et al. 2000), and therefore the ability of soil to accumulate and preserve mineralizeable organic matter has received growing interest. The likely direction of change in Soil Organic Carbon (SOC) in the boreal forest biome, which harbors the world's second largest SOC stock, is of marked concern because climate warming is projected to be greatest in high latitudes (McGuire et al. 2000) and temperature is the cardinal determinant of soil C mineralization.

Black spruce (*Picea mariana* (Mill. B.S.P.)) forests are a major component of the boreal system (Gower et al. 2001) and are the dominant cover type of Alaska, where they cover $26 \times 10^6$ ha (Heath et al. 2003) and harbor ~250 Mg C $ha^{-1}$ (Johnson and Kern 2003). Our understanding of the soil C storage potential of black spruce forests is unclear because soil C balance in these forests, as in most forests, is determined by the balance between inputs from net primary productivity (NPP) and outputs from decomposition, both of which are likely to change under a warmer climate.

Productivity in black spruce forests is largely limited by the rate of nutrient release from decomposing organic matter (Van Cleve et al. 1981), which often increases with higher soil temperatures (Van Cleve et al. 1990) and varies with landscape position (Van Cleve and Yarie 1986, Van Cleve et al. 1991). Landscape position, in turn, directly influences soil temperature and insolation on a seasonal basis. Dramatic temperature inversions are common in the winter and cause cold air to settle in valley bottoms. On the other hand, uplands are often cooler in the summer months in interior Alaska (Slaughter and Viereck 1986). Soils on northern aspects receive less insolation than do those on southerly slopes and therefore are cooler (e.g., Slaughter and Long 1975) and more likely to contain permafrost (Rieger 1983). The interaction between reduced nutrient release from arrested decomposition and decreased stand production on cold, north-facing aspects provides a niche for moss expansion (Oechel and Van Cleve 1986). Moss development and organic matter accumulation, in turn, increase the thermal buffering capacity of the soil (Bonan 1991), which significantly reduces the soil heat sum (Van Cleve and Viereck 1981). In addition to affecting local soil temperatures, relatively small-scale changes in topography can impact the lateral and vertical movements of water, which affects soil nutrient transformations and uptake in black spruce forests (Yarie and Van Cleve 1983, Grant 2004). For example, forests near summits are more prone to nutrient loss whereas toe-slope forests are in regions of nutrient accumulation (Ping et al. in review). Therefore, differences in physiographic features can create natural gradients in temperature and nutrient mineralization, which in turn impact stand production and the soil decomposition environment (Viereck et al. 1983). However, it is not yet clear how variation across the landscape in the relative rates of productivity and decomposition impact soil C accumulation and its resistance to decay in black spruce forests.

There have been few investigations of how SOC changes along landscape-induced thermal gradients in black spruce forests of interior Alaska. Dyrness and Grigal (1979) investigated vegetation-soil relationships along a 3 km elevational/slope gradient.

Although the authors did not quantify SOC, they determined that the thickness, mass, and N concentration of the organic soil horizons all generally increased downslope. However, relating soil properties to production and climate in this study was hampered by the low number of black spruce stands (n = 6), the narrow range in stand productivities investigated, and the lack of temperature data. Two later investigations have shown a decrease in mineral soil C and an increase in nutrient mineralization in organic horizons with increasing soil temperature. However, these studies did not control for stand type (Viereck and Van Cleve 1984) or drainage class (Van Cleve et al. 1981), both of which affect SOC storage in boreal forests (Rapalee et al. 1998). Furthermore, the low number of black spruce stands compared in these two studies (n = 5 and 2, respectively) precluded isolation of any pattern between SOC and stand productivity and temperature in interior Alaska.

The bioavailability of surface organics decreases during humification, resulting in compounds that decay by orders of magnitude more slowly than fresh litter (Aber et al. 1990, Berg and Meentemeyer 2002). Soil organic C content and lability often decrease along a continuum with depth (Melillo et al. 1989, Winkler et al. 1996), however the amount of relatively labile organic matter associated with mineral soil varies depending on the degree of organic matter illuviation (Kaiser et al. 2002), the sorptive capacity of the mineral soil (Kaiser and Guggenberger 2003, Moore 2003), and the decomposition environment within the mineral soil (Rodionov et al. 2000, Elberling and Brandt 2003). Since organic matter in the mineral soil increases in density as it decomposes (Baisden et al. 2002b), fractionating mineral soils into different density separates can shed light on the degree of organic matter decomposition (Schulten and Leinweber 1999) and lability (McLauchlan and Hobbie 2004) in a soil horizon. Therefore, quantifying the relative amounts of light and dense mineral associated organic matter along gradients in stand production (and hence, nutrient mineralization) might show how soil C stabilization is affected by the cycling of surface organic matter, but this has received little study.

To determine how the complex interplay between stand production, nutrient mineralization, and soil temperature affects soil C stabilization, we investigated total SOC along four replicate gradients in black spruce productivity and climate in interior Alaska. Because turnover of soil pools is faster in warmer, more productive sites, we hypothesized: 1) total SOC decreases as stand productivity and soil temperature increase because productivity depends on the nutrients released with soil organic matter turnover, which also increases with temperature, the net result of which being that 2) SOC content in more stable soil pools is proportional to the amount of actively cycling SOC in surface soil pools.

**1.3: *Study Sites***

Four replicate gradients in black spruce productivity and soil temperature were established across interior Alaska, ranging in latitude from 63°-65°N (~365 km) and in longitude from 142°-148°W (~550 km). Study areas were named for the roads used to access them and were established (from west to east) off of the Parks Highway (P), Murphy Dome (M), the Elliott Highway (E), and the Taylor Highway (T) (Figure 1). Each study area consisted of three sites, which were differentiated by their relative level of stand production, or Site Index. Site Index (SI) is defined as the height of stand dominants attained after 50 years, and sites with low (L, 2.5-4.6 m), medium (M, 4.6-7.5 m), or high (H, 8.1-12.5 m) SI values occur within each study area. Site hierarchy, stand properties, and landscape characteristics are outlined in Table 1.

Vegetation at the low productivity stands was dominated by *Ledum groenlandicum* and *Vaccinium* spp. (*vitis-idaea and uliginosum*) shrubs. *Betula nana* was also common at the TL site. *Vaccinium* spp. were common in the forb layers at the medium and high productivity stands, with *Equisetum sylvaticum* also being common at the high productivity stands. Moss species were determined by harvesting all living moss in three 0.2 $m^2$ plots at each site. Dominant moss genera at all sites included *Hylocomium* and *Pleurozium*, as most sites were moderately to well drained. *Sphagnum* was also common at the PL site, however, as permafrost impeded drainage despite its relatively high slope and elevation.

Soils in interior Alaska, and the Yukon-Tanana uplands in particular, are dominated by a mantle of mica-rich loess that continues to blow in from the outwash plain of the Tanana Valley (Pewe and Reger 1983, White et al. 2002). Although a thorough soil survey has not been conducted across much of interior Alaska, incipient loess parent material was dominant at all sites. Intermittent permafrost denotes a Gelisol, and the PL and EM soils were keyed as Typic Historthels and the EL and TL soils were Aquic Haplorthels. Inceptisols occurred at all of the other sites. The ML and EH sites were keyed into the Cryaquept Great Group because the matrix showed redoximorphic features. The relatively low SOM content and a lack of long periods of inundation placed soils from the other sites into the Cryochrept Great Group. The Histic Subgroup applied to the ML, PM, and EH sites. Inceptisols were within the Cryic temperature regime.

**1.4: Methods**

Methods of site establishment generally followed the protocols of the Growth and Yield Program at the University of Alaska, Fairbanks (E.C. Packee), and this study incorporated four sites from that study. Briefly, each site consisted of three 20 $m^2$ plots, which were spaced 20 m apart (either in a triangle formation or perpendicular to slope). We sought backslopes in order to keep control for drainage, nutrient accumulation, and nutrient mobility across sites (Table 1). The primary landscape attributes selected were aspect and local relief because they determine insolation, which in turn drives changes in soil temperature and stand productivity. The colder, low-productivity stands occurred on north facing slopes and typically received less insolation annually. Control over temperature and production at the T sites also included adiabatic cooling, which was driven mainly by elevation. Furthermore, only mature stands were selected. Chen and others (2002) have shown that black spruce stand production dominated total aboveground productivity after approximately 30 years in Canadian uplands. Overstory production within a SI class declined gradually in later stages of stand development. Therefore, we avoided young (<60 years) and old (>150 years) stands so as to not confound changes in stand ontogeny with site/climatic limitations to stand production.

Stand density was determined by tallying every tree (>1.27 cm diameter) in each plot (0.04 ha) at each site. The height and diameter at breast height (1.37 m) of every tree in ten 0.001 ha randomly distributed plots at each site were measured. Age of dominant trees (at breast height) in each stand was determined by felling or coring 4-6 dominant trees at each site. The heights attained by dominant trees at 50 years of age (SI, m) from each stand in this study were reconstructed through use of the height-age relationship developed by Rosner (2004). Actual ages of dominant trees (age at soil surface) in each SI class were derived from ages at breast height using growth-age relationships at cross sections obtained at 1.2 m intervals along each tree harvested (following Carmean 1972). Site Index equations were developed from 292 trees harvested from 33 sites across interior Alaska (Rosner 2004), of which four sites have been incorporated in this study (Table 1). The utility of SI as a measure of site productivity is generally limited to mature, even-aged, undisturbed, mono-specific stands (Carmean and Lenthall 1989) and these criteria were found at all of the sites in this study.

Soil temperatures from within the middle of the fibric and hemic soil horizons, from a depth of 5 cm into the mineral soil, and from air temperatures at 100 cm were monitored continuously from June 2003 to June 2004 at each site. Soil temperatures were monitored by genetic horizon because temperatures obtained from a fixed depth could lead to comparisons of pedogenically dissimilar horizons (such as organic vs. mineral soils) across sites. Readings were recorded hourly using HOBO H8 data recorders (Onset Computer Corporation, Bourne MA, USA). Radiation shields were used to protect air temperature thermistors from direct solar radiation. Solar insolation (watt hours $m^{-2}$ $yr^{-1}$) was calculated for each site using Solar Analyst 1.0 (2000) software (Helios Environmental Modeling Institute, KS, USA). This software uses Digital Elevation Model (DEM) input in reconstructing the angular distribution of sky visibility versus obstruction (from horizon angle) in conjunction with a raster representation of the apparent position of the sun as it varies through time.

### 1.5: *Soil Analyses*

Soil descriptions were obtained from 4 randomly distributed pits dug at each plot, for a total of 12 pits at each site. Soil pits were approximately 60 $cm^2$, and dug into 5 cm of B1 horizon. Horizons were delineated based on the USDA NRCS classification scheme (Schoenberger et al. 2002). Organic soils were differentiated in the field by the degree of humification into fibric (Oi), hemic (Oe), and sapric (Oa, which rarely occurred) soil horizons. All living, green moss was discarded. Slightly or intermediately decomposed brown moss was designated as part of the Oi or Oe horizon, respectively. Mineral soils were separated into A and B1 (5 cm) horizons. Three cores (5.08 cm diameter) were obtained from the "best face" of each descriptive pit. Parsed soil cores were bulked by Oi, Oe and Oa, A, and B1 (depth of 5 cm) horizons at each descriptive pit.

Subsamples of homogenized organic soils were deprived of >2 mm roots and soil wood. Mineral soils were passed through a 2 mm sieve. All soils were dried at 105°C. Organic soils were ground in a ball mill, but grinding of the fine-textured loess mineral soil was not required. Initial tests revealed that the carbonate content of mineral soil was negligible and therefore soils did not undergo an acid treatment. Carbon content was determined through combustion in a LECO 2000 CNS analyzer. Samples were run at least in duplicate and the difference between replicates was <5% of the value.

The A horizons were homogenized and 1.8 g was subsampled from 6 pits at each site, chosen at random, to be separated by density into fractions <1.6 g $cm^{-3}$ (light) and >1.6 g $cm^{-3}$ (dense) using sodium polytungstate solution (Sometu Co., Sherman Oaks, CA, USA). The choice of density was based on earlier studies in which fresh plant materials were separated from mineral associated organic matter (Golchin et al. 1994). Although these are somewhat operational definitions, the light fraction is composed mainly of relatively undecomposed vegetation whereas the dense fraction represents a mixture of more refractory material associated with the mineral soil (Schulten and Leinweber 1999, Baisden et al. 2002b, McLauchlan and Hobbie 2004). We followed the density fractionation procedure described in detail by Baisden et al. (2002b), following Golchin et al. (1994). Generally, 1.8 g of A horizon soil was weighed into 50 mL centrifuge tubes filled with 35mL of sodium polytungstate solution (SPT). The tubes were gently shaken, allowed to sit for >1 hour, washed with an additional ~5 mL of SPT from a wash bottle, and centrifuged to aid in separation of particles <1.6 g $cm^{-3}$. The floating, >1.6 g $cm^{-3}$ fraction was aspirated onto a pre-baked quartz fiber filter (Pall, Tissuquartz), rinsed with de-ionized water (DI), dried, and weighed. Residual soil was re-suspended in 35 mL of SPT and then was subjected to 6 minutes of ultrasonic disruption using a Biosonik probe style sonifier (Bronwill Scientific, Rochester, NY, USA) with an energy output of approximately 40 J $mL^{-1}$ (following North 1976). This liberated a negligible amount of soil organic matter (SOM) across all sites (<0.03 g $g^{-1}$ initial mass), showing that there was little occluded organic matter due to the low clay content and weak structure of the loess soils. This small mineral associated <1.6 g $cm^{-3}$

fraction was aspirated onto a quartz filter as before, but no further analysis was conducted on this fraction. The dense (>1.6 g $cm^{-3}$) fraction was then rinsed with DI and collected as before. Recovery of fractionated soil across all samples deviated by 0.03 g $g^{-1}$ initial dry weight. The small amount of dissolved organic matter lost with the SPT was not recovered and could not be accurately calculated.

Stable isotope measurements were performed with a GEO 20-20 duel-inlet isotope ratio mass spectrometer coupled with a PDZ Europa solid preparation module. Results are expressed in standard $\delta^{13}C$ notation, as the deviation (‰) relative to the isotopic ratio of Pee Dee belemnite standard. The samples used for $\delta^{13}C$ analysis were the same as those used for total C analysis and in the density fractionations. Samples were analyzed at least in duplicate and the deviance was <0.2‰. For radiocarbon analysis, dense fraction (>1.6 g $cm^{-3}$) soils were converted to $CO_2$ by combustion at 900°C for 2 hours in quartz glass tubes. Sample gas was purged of all non-$CO_2$ gasses by condensing in liquid nitrogen at -70°C. Purified $CO_2$ was then graphitized with an iron powder catalyst (99.99%) in a method modified from Vogel et al. (1984). Graphite was sent to the Keck Carbon Cycle/Accelerator Mass Spectrometry facility at the University of California, Irvine (USA) for radiocarbon analysis. Isotopic values for $\Delta^{14}C$ are reported as the deviation (‰) from the $^{14}C/^{12}C$ ratio of oxalic acid, and the analytical precision was 3.8‰. The $\Delta^{14}C$ values were also corrected for mass-dependent fractionation through use of the $\delta^{13}C$ value (Stuiver and Polach 1977). We used the $\Delta^{14}C$ values to estimate the average age of SOM in dense fraction mineral soil from the A horizon following methods described in detail by Trumbore (2000).

### 1.6: *Data Analyses*

Site means in soil properties were composed of plot level means (n = 3). Two soil samples were chosen at random from each plot for $\delta^{13}C$ isotope analyses (n = 6). Density fractionations were bulked at the plot level, and one of the bulked dense fractions from each site was randomly chosen for radiocarbon analysis. However, graphite quality from the iron catalyzed reduction of $CO_2$ was unsuitable for radiocarbon analysis at 4 sites and therefore n = 8. All cross-site comparisons involving soil temperature have an n = 10 because the TH and TM sites were consumed in forest fires in the spring of 2004.

Differences among sites in mean annual temperature and soil profile depths were tested with one-way analysis of variance pairwise contrasts at $\alpha = 0.05$ (Scheffe error protection). Pearson correlation coefficients demonstrated relationships between variables ($\alpha = 0.05$). Multiple regression analysis (forward selection procedure) developed empirical relationships between SOC and stand and physiographic properties (n = 12). Predictor variables were accepted at $\alpha = 0.05$ and if more than 5% of the variance could be explained. All descriptive statistics were performed with Analyze-it statistical module (Analyze-it Software, Ltd. Leeds, UK).

### 1.7: Results

### 1.8: *Stand Productivity and Soil Temperature*

Mean stand basal area and diameter were highly correlated with SI ($r = 0.76$ and 0.83, respectively; $p < 0.004$). Site Index, basal area, and stand diameter all increased with mean annual temperature ($r = 0.70, 0.55, 0.73$, respectively) and mean growing degree days (GDD) >0°C ($r = 0.69, 0.89, 0.83$, respectively; $p < 0.05$) of the organic soil horizons (see Table 2). Temperature measurements taken in the mineral soil (5 cm) did not significantly explain any of the variance in SI ($p = 0.14$), probably because the rooting depth of black spruce is largely limited to the organic horizons (e.g., Viereck and Johnston 1990).

Soil temperatures by genetic horizon varied among sites ($p < 0.001$). Mean annual temperatures (MAT) changed by ~2.5 °C in the organic horizons and by ~2.3°C at a depth of 5 cm in the mineral soil across all sites (Table 2). Total organic matter depths varied significantly between study sites (Table 2). Soil heat sums decreased significantly with organic matter depth across all sites, with low productivity sites exhibiting deeper organic soil horizons (Figure 2). Mineral soil temperatures decreased marginally on sites with deeper total organic matter accumulation ($r = -0.59$ for both summed GDD and MAT at a depth of 5 cm into the mineral soil; $p = 0.07$). Mean annual soil temperatures obtained at a fixed depth (5 cm into the mineral soil) correlated well with MAT in the Oe horizon across sites ($r = 0.88$, $p < 0.001$).

**1.9:** ***Soil Organic Carbon, Productivity, and Climate***

The SOC content of different genetic horizons on an area basis (kg C $m^{-2}$) varied substantially along the four black spruce productivity gradients (Figure 3). The proportion of total SOC in surface mineral soil horizons (A and 5cm of B1) shifted from 52% (±3.3) in low productivity stands to 41% (±1.1) in stands exhibiting higher growth rates (Figure 3). Low-productivity black spruce stands harbored significantly more C in the mineral soil horizons than did sites with higher stand productivity (Figure 4). Carbon stocks in the combined organic horizons varied significantly among sites, and SOC in the

Oi horizons decreased with increasing SI across all sites ($r = -0.72$, $p = 0.01$). Moreover, total SOC (mineral and organic horizons) was inversely related to SI ($r = -0.69$, $p = 0.01$; Figure 2) and stand basal area ($r = -0.72$, $p < 0.01$) across all sites. The observed decrease in total SOC with increased SI (Figure 2) was significant even when the "PL" site (SI = 2.49m) was removed from the analysis ($R^2 = 0.37$, $p < 0.05$).

Organic C harbored in mineral soil horizons decreased significantly with an increase in summed GDD within Oe soil horizons at a rate of 4.4g C $m^{-2}$ for every GDD increase in heat sum (Figure 5). This observed relationship between mineral SOC and summed GDD (Oe) across all sites (Figure 5) was robust and significant even when the "TL" site (158 GDD) was removed from the analysis ($R^2 = 0.57$, $p = 0.02$). Furthermore, the most productive sites (SI = 8.5 – 12.0 m) also exhibited the warmest soil temperatures and had the lowest mineral soil C content (Figure 5). Soil organic C in the mineral soil horizons decreased with increased summed GDD within Oi horizons ($r = -0.84$, $p < 0.01$), and also with increased MAT in the Oe horizons ($r = -0.74$, $p = 0.02$).

Depth and SOC content of Oi horizons decreased with an increase in the heat sum of the Oe horizons across all sites ($r = -0.67$ and $-0.60$, $p = 0.03$ and $0.07$, respectively). However, the SOC content of the Oe and Oa horizons increased with heat sum ($r = 0.67$, $p = 0.03$) across all sites. Due to this opposing trend, summed SOC in all genetic horizons did not vary with temperature ($p = 0.13$), despite a marked decrease in mineral soil C with an increase in soil temperature and stand productivity (Figure 5).

Total SOC accumulation was inversely related to insolation ($r = -0.73$, $p = 0.01$; Table 1). Mean annual air temperatures (1 m) increased with higher insolation ($r = 0.77$,

$p < 0.001$) and organic soil horizon depths decreased with increasing insolation across all sites ($r = -0.85$, $p < 0.001$). However, the negative relationship between insolation and total SOC accumulation was confounded somewhat by the occurrence of cooler, low productivity sites at relatively higher elevations where there was little obstruction of sunlight (i.e., the "TL" and "ML" sites). Elevation accounted for part of this variability and explained an additional 35% of the variance in total SOC accumulation when combined with insolation ($R^2 = 0.88$, $p < 0.001$).

Across all sites, the light fraction (<1.6 g $cm^{-3}$) SOM in the A horizon (g $g^{-1}$ dry soil) decreased with increasing soil MAT ($r = -0.67$ and $-0.71$ for Oe and mineral soil horizons, respectively; $p < 0.05$) and increasing GDD ($r = -0.75$ and $-0.72$ for Oe and mineral soil horizons, respectively; $p < 0.05$). Furthermore, the most productive sites (SI = 8.5 – 12.0 m) also exhibited the warmest soil temperatures and had the lowest proportion of light fraction SOM in the A horizon (Figure 6). This observed relationship between the proportion of light fraction mineral soil and summed GDD (Oe) across all sites (Figure 6) was significant even when the "TL" site (158 GDD) was removed from the analysis ($R^2 = 0.49$, $p = 0.04$). Light fraction SOC on an area basis (kg C $m^{-2}$) also decreased with an increase in summed GDD (Oe) across all sites ($r = -0.62$, $p = 0.06$). Site Index was unable to explain any additional variance in light fraction C content (kg C $m^{-2}$, $p = 0.16$), but SI explained 35% of the variance in decreasing light fraction A horizon mass (g $g^{-1}$ dry soil, $p = 0.04$).

**1.10: *Soil Carbon Cycling and Turnover***

The degree of $\delta^{13}C$ enrichment in dense fraction A horizon mineral soil relative to organic horizons was negatively related to C storage in underlying B1 mineral soil horizons (Figure 7A). Furthermore, as $\Delta^{14}C$ enrichment in dense fraction A horizon mineral soil increased, C storage in the underlying B1 horizon decreased across all sites (Figure 7B). Total C accumulation in the mineral soil horizons (kg C $m^{-2}$, A + B1) showed a marginal inverse relationship with $\Delta^{14}C$ enrichment ($r = -0.64$, $p = 0.09$). Mean Residence Times (MRT) of C in dense fraction A horizon SOM ranged from 282-672 years (Figure 7B). Soil C turnover rate ($MRT^{-1}$) and soil temperature were both negatively related to SOC storage in the mineral soil horizons (Figures 7 and 5, respectively).

As an independent test of how a change in fractionation density might affect the pool isolated, the A horizon from an additional black spruce site near Delta Junction, AK (King et al. 2002) was fractionated at 2.0 g $cm^{-3}$, and analyzed for total C and $\Delta^{14}C$ in a similar manner. The $\Delta^{14}C$ value of this fraction was –18.2‰ (Mean residence time of 415 years), and the SOC content of the underlying B1 horizon was 2.00 kg C $m^{-2}$. These values fall within 95% confidence limits of the relationship between $\Delta^{14}C$ in the >1.6 g $cm^{-3}$ fractions and B1 horizon SOC across 8 sites in this study (Figure 7B). This suggests that loess-based mineral soil fractions >1.6 and >2.0 g $cm^{-3}$ are composed of pedogenically-similar SOM.

## 1.11: Discussion

### 1.12: *Soil Carbon Cycling and Turnover*

The proportional amount of labile SOM in the mineral soil decreased with increasing temperature and stand productivity (Figure 6), but the extent of these changes in SOM lability and the implications for increased C mineralization in a warmer climate are uncertain. Carbon mineralization in mineral soils at the cooler sites, with a higher proportion of light-fraction SOM, could exhibit a higher sensitivity to soil warming (Raich and Schlesinger 1992, Kirschbaum 1995, Bekku et al. 2003). On the other hand, C mineralization from soils obtained from the warmer sites, with a lower proportion of light-fraction SOM, could be relatively insensitive to warming (Liski et al. 1999, Giardina and Ryan 2000, Dalias et al. 2001). Previous research suggests that SOM diminishes asymptotically at higher temperatures because there is likely a limit to the amount of decomposition that is sensitive to temperature (e.g., Melillo et al. 1989, Aber et al. 1990), but our empirical data show linear decreases in SOC and light fraction SOM in the mineral soil with increasing heat sum. Feedbacks between temperature, SOM depth, and SOC turnover may explain part of the linear decrease in SOC observed at warmer temperatures.

The stable isotopic composition ($\delta^{13}C$) of the light and dense soil fractions can be used to corroborate information gained from their relative quantities (Baisden et al. 2002a,b). Although several factors can affect the fractionation of $^{13}C$ in soils (Ehleringer et al. 2000, Fernandez et al. 2003), SOM generally becomes more enriched in $^{13}C$ as turnover processes preferentially mineralize $^{12}C$ (Nadelhoffer and Fry 1988, Boutton 1996). Soil organic matter density and $\delta^{13}C$ enrichment both increase with decomposition (Amundson and Baisden 2000, Baisden et al. 2002b) and follow a continuum from litter to mineral soil (Melillo et al. 1989, Balesdent et al. 1993, Ehleringer et al. 2000). Therefore, the greater degree of $\delta^{13}C$ enrichment in dense fractions relative to the lighter Oe horizons observed in the warmer, more productive stands (Figure 7A) suggests C mineralization between the organic and mineral soil horizons increased with temperature and production.

The amount of $^{14}C$ incorporated into different density fractions is also capable of distinguishing turnover rates of SOM (Trumbore 2000, Baisden et al. 2002a,b). The $^{14}C$ signature of SOM reflects that of the atmospheric $CO_2$ concentration at the time of fixation into the original plant material. The low $\Delta^{14}C$ values (<0‰) observed in dense fraction soils from the cooler, low productivity sites indicates that C exchange with modern (<40 years) atmospheric $^{14}C$ concentrations, which were enriched via thermonuclear weapons testing in the 1960's, is very slow (i.e., MRT>300 years). The higher $\Delta^{14}C$ values observed in the warmer, more productive sites reflect higher rates of exchange (i.e., higher C turnover) with the atmosphere (Figure 7B).

**1.13:** ***Stand-soil interactions***

The lower mineral soil C storage on more productive (Figure 4), warmer (Figures 5 and 6) sites with faster soil C turnover rates (Figure 7), suggests that decomposition in the mineral soil may be more temperature sensitive than net primary productivity. However, total SOC in combined mineral and organic soil horizons did not significantly decrease with increased soil heat sum (Figure 5), most likely because of increased C inputs to the soil in warmer, more productive forests. Mack and others (2004) have similarly shown that while total SOC decreased with increased plant production (with nutrient availability increased via fertilization) in the Alaskan tundra, SOC in upper organic soil horizons (>5 cm) increased with nutrient fertilization. Chen et al. (2002) demonstrated that although proportionally more of the total C fixed in low productivity black spruce stands of Canada (SI=3) was allocated belowground, total C inputs to the soil, above and belowground, were higher in more productive stands (SI=11). Ruess et al. (2003) have similarly shown that belowground production increased in step with aboveground production in floodplain black spruce stands in interior Alaska. If these trends in stand production apply to this study, and one considers that SOC in organic soil horizons did not significantly change with productivity or climate, it follows that turnover of forest-derived C inputs to the organic soil must have increased with stand production and temperature. However, the products of SOM turnover did not accumulate in the mineral soil (Figure 7), which could be due in part to the low sorptive capacity for the products of SOM turnover within incipient, loess-dominated mineral soil (Ugolini et al. 1987, Dahlgren and Marrett 1991, Moore 2003).

**1.14: *Biophysical Controls over Temperature and the "Thermal Advantage"***

Moss significantly increases surface insulation and therefore exerts major control over soil temperatures (Rieger 1983, Bonan 1991, Beringer et al. 2001). It is no surprise, then, that soil temperatures in all horizons decreased with increasing organic matter thickness (particularly fibric material) across the productivity/temperature gradients in this study (Figure 2). Air temperatures (1 m) did not significantly vary across all sites

(Table 2), but soil temperatures in the lower productivity sites (with deeper Oi) remained cooler and had a lower heat sum (Figure 2). These data suggest that the moss layer increased the thermal inertia of soils relative to air temperature by insulating the soil.

Swanson and others (2000) hypothesized that peat accumulation in cold climates increases with the divergence between warmer air temperatures and cooler soil temperatures, which gives a "thermal advantage" to biological processes occurring in the air (primary productivity) relative to biological processes occurring in the soil (microbial decomposition and mineralization). We tested this hypothesis across our four productivity/temperature gradients by subtracting the heat sum (GDD >0 °C) of the Oe soil horizon from the air (1 m) GDD. We found that the thermal environment of primary production relative to soil decomposition ($GDD_{(air\text{-}Oe)}$) explained 57% of the variance in total SOC storage across all sites (Figure 8). Furthermore, $GDD_{(air\text{-}Oe)}$ explained 8% more of the variance in mineral soil C storage than did soil heat sum alone (Figures 5 and 8).

Another measure of the advantage of primary productivity over soil decomposition environment was the annual potential amount of incoming insolation at each site (Table 1). The thermal advantage of productivity relative to decomposition was lower on sites with greater amounts of insolation because organic matter depths, and hence soil insulating properties, decreased. This expectation was consistent with the decrease in total SOC harbored in all horizons with increasing insolation across all sites.

### 1.15: *SOC in Alaska compared to other boreal forests*

Upland spruce forests of interior Alaska harbor considerably more SOC than previous reports for other boreal forests. We sampled to 5 cm B horizon in this study, and Ping at al. (1997) identified an additional 5.9-7.6 kg C $m^{-2}$ in underlying B and C horizons to 1m depth in upland forests of interior Alaska that exhibited similar bulk densities and C concentrations to the sites in this study (total profile SOC ranging from 16.9-78.7 kg C $m^{-2}$). These SOC values for interior Alaska are 1 to 19 x greater than previous reports for upland boreal forests spanning Sweden, Finland, Norway, and Denmark (Callesen et al. 2003). Moreover, the mean amount of SOC reported for organic horizons in these Fennoscandian forests was 29%-66% less than values observed in this study, and was 38% and 64% less than organic horizon SOC reported for upland white spruce (Steltzer, 2004) and upland black spruce (Ping et al. 1997) in northwest and interior Alaska, respectively.

The observed decline in mineral soil C with increasing heat sum and stand productivity in interior Alaska differs from previous reports along thermal/stand productivity gradients in Fennoscandian boreal forests (Liski and Westman, 1995, 1997). Callesen et al. (2003) demonstrated that total SOC (across 3 texture classes to a depth of 1 m) in mixed boreal forests spanning four Nordic countries increased from 4.1 to 12.0 kg $m^{-2}$ as mean annual temperature increased from 0 to 7.5 °C. Warmer temperatures and increased parent material development (Liski et al. 1998) across the Nordic study region are in sharp contrast with the cooler temperatures (Table 2; Ping et al. 1997) and incipient Gelisols and Inceptisols of interior Alaska (Allan 1969, Ping et al. *in review*). These data suggest that SOC accumulation in upland boreal forests of interior Alaska is largely controlled by the degree to which decomposition is arrested at cooler temperatures, whereas soil textural properties exert more control over SOC in other boreal forests.

**1.16: Conclusions**

Near surface SOC decreased with increased stand productivity (and hence, probable increases in nutrient availability) across four replicated gradients in black spruce stand production and climate in interior Alaska. In addition, the distributions of pools within the mineral soil (a decrease in total SOC and proportion of light fraction SOM) and within the organic soil (a decrease in Oi horizon depth and SOC) were apparently driven by interactions between increased soil temperature and stand productivity across all sites. Variation in total SOC across all sites was best explained by the difference in above and belowground heat sums, which we suggest is largely mediated by the insulative properties of surface organic horizons. The interactions between increased stand productivity, organic matter accumulation, and soil heat sum have altered the quality and quantity of different SOM pools through differences in decomposition environment and plant C inputs, even though total SOC appeared insensitive to temperature in some cases. These data suggest that temperature responsiveness of organic matter decomposition within discrete soil pools, and not just total soil C stocks, needs to be examined in order to project the effects of changing climate and primary production on soil C balance.

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TABLE 1.1. Biophysical properties of the twelve black spruce sites in this study.

| | Study Sites[†] | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Stand Characteristics | PL | EL | TL | ML | EM | PM | TM | MM | PH | MH | EH | TH |
| Site Index[‡] (m) | 2.49 | 4.32 | 4.53 | 4.61 | 4.62 | 4.77 | 7.40 | 7.54 | 8.12 | 8.78 | 11.64 | 12.47 |
| Dominant Age | 150 | 130 | 60 | 60 | 140 | 120 | 60 | 60 | 120 | 120 | 120 | 80 |
| Dominant Height (m) | 5.41 | 8.66 | 5.18 | 5.28 | 9.62 | 8.95 | 8.38 | 8.53 | 13.34 | 14.02 | 16.61 | 15.32 |
| Trees $ha^{-1}$ | 3085 | 4800 | 1111 | 1700 | 2525 | 6975 | 2667 | 6500 | 2475 | 6941 | 2683 | 3679 |
| Basal Area ($m^2$ $ha^{-1}$) | 4.60 | 15.06 | 1.84 | 3.54 | 17.48 | 21.22 | 13.51 | 22.72 | 21.32 | 26.80 | 21.47 | 31.38 |
| Landscape Properties | | | | | | | | | | | | |
| Elevation (m) | 427 | 335 | 993 | 549 | 366 | 470 | 877 | 549 | 455 | 412 | 427 | 730 |
| Aspect (°) | 360 | 330 | 283 | 330 | 330 | 315 | 328 | 180 | 165 | 190 | 360 | 147 |
| Slope (%) | 30 | 5 | 13 | 8 | 13 | 15 | 3 | 8 | 8 | 5 | 13 | 16 |
| Hillslope Position[§] | BS | FS | H BS | H BS | BS | H BS | RI | BS | BS | BS | H BS | BS |
| Insolation (watt hrs. $m^{-2}$ $yr^{-1}$) x$10^5$ | 6.4 | 8.3 | 10.3 | 8.8 | 7.5 | 8.7 | 10.8 | 10 | 9.6 | 9.7 | 8.1 | 10.3 |
| Latitude (°N) | 64.770 | 65.108 | 63.660 | 64.958 | 65.106 | 64.767 | 63.460 | 64.955 | 64.765 | 64.977 | 65.102 | 63.397 |
| Longitude (°W) | 148.280 | 147.882 | 142.290 | 148.241 | 147.882 | 148.298 | 142.468 | 148.241 | 148.300 | 148.014 | 147.882 | 142.489 |

[†]Study site codes are denoted by their location near the Parks Highway (P), Elliott Highway (E), Taylor Highway (T), and Murphy Dome (M). Stands are ranked by productivity, with Low (L), Medium (M), or High (H) Site Index values occurring within each study area. Previous research incorporating or proximate to a given site includes: Barney et al. 1978 (PM), Rosner 2004 (TL, TM, TH, EH), Van Cleve et al. 1981 (PM), Viereck et al. 1983 (PL, PM), Vogel et al. 2005 (PM), and the Growth and Yield Program, University of Alaska, Fairbanks: E.C. Packee (TL, TM, TH, EH).

[‡]Site Index is the height (m) obtained by dominant trees in the stand after 50 years.

[§]Geomorphic descriptions as described by the USDA NRCS (2002) are: BS=Backslope, FS=Footslope, RI=Rise in locally flat area. H (High) and L (Low) describe microrelief.

TABLE 1.2. Mean soil profile depths (by genetic horizon) and temperature data for each horizon across the four black spruce gradients in productivity and temperature. Numbers followed by different letters are significantly different (one-way analysis of variance pairwise comparisons; $\alpha = 0.05$).

| Temperatures and Depths by Horizon | | Study Sites[†] PL | EL | TL | ML | EM | PM | TM | MM | PH | MH | EH | TH |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Depth (cm) | Oi | 16.17b | 11.42ab | 11.83ab | 6.25a | 10.00ab | 5.58a | 4.92a | 4.25a | 5.08a | 4.00a | 5.83a | 5.08a |
| (SE)[‡] | | (1.48) | (1.45) | (1.90) | (0.69) | (1.09) | (0.31) | (0.50) | (0.30) | (0.47) | (0.30) | (0.41) | (0.48) |
| | Oe, Oa | 25.58b | 18.67ab | 19.17ab | 15.33ab | 24.42ab | 16.75ab | 12.67a | 12.58a | 14.67ab | 12.75a | 21.08ab | 13.33ab |
| | | (1.03) | (1.72) | (1.64) | (1.31) | (1.11) | (0.90) | (1.37) | (0.95) | (0.99) | (0.84) | (1.22) | (0.96) |
| | A | 27.91b | 19.91ab | 21.83ab | 17.67ab | 25.75ab | 18.75ab | 16.33ab | 14.25a | 17.00ab | 15.50ab | 22.42ab | 16.42ab |
| | | (0.99) | (1.77) | (1.85) | (1.41) | (1.05) | (0.97) | (1.42) | (0.99) | (0.93) | (0.82) | (1.28) | (0.92) |
| MAT[§] (°C) | air (1 m) | -0.69a | -1.81a | -2.32a | -1.86a | 0.30a | -1.68a | | -1.95a | -1.40a | -1.84a | -0.26a | |
| (SE) | | (0.77) | (0.70) | (0.74) | (0.74) | (0.63) | (0.74) | | (0.72) | (0.73) | (0.74) | (0.64) | |
| | Oi | -0.06abc | -0.06abc | -0.72ab | 0.60bc | 1.53c | -1.93a | | 1.66c | 1.28bc | 1.00bc | 1.64c | |
| | | (0.33) | (0.33) | (0.25) | (0.29) | (0.35) | (0.45) | | (0.31) | (0.32) | (0.40) | (0.37) | |
| | Oe, Oa | -0.09abc | -0.87ab | -1.14a | 0.53cd | 0.63cd | 0.30bcd | | 1.35d | 1.30d | 1.02cd | 1.28d | |
| | | (0.20) | (0.19) | (0.12) | (0.11) | (0.23) | (0.26) | | (0.21) | (0.24) | (0.20) | (0.21) | |
| | mineral | -0.36bc | -1.38a | -1.37a | 0.26de | -0.61b | 0.49ef | | 0.89f | 0.68ef | 0.48def | 0.10cd | |
| | (5 cm) | (0.15) | (0.14) | (0.08) | (0.06) | (0.07) | (0.10) | | (0.11) | (0.10) | (0.06) | (0.05) | |

[†]Study site codes are the same as in Table 1 and are denoted by their location off of the Parks Highway (P), Elliott Highway (E), Taylor Highway (T), and Murphy Dome (M).

[‡]SE, Standard Error of the measurement.

[§]MAT is the Mean Annual Temperature (°C) from June 2003 - June 2004. Forest fires in 2004 prevented the collection of complete data sets at the TM and TH sites.

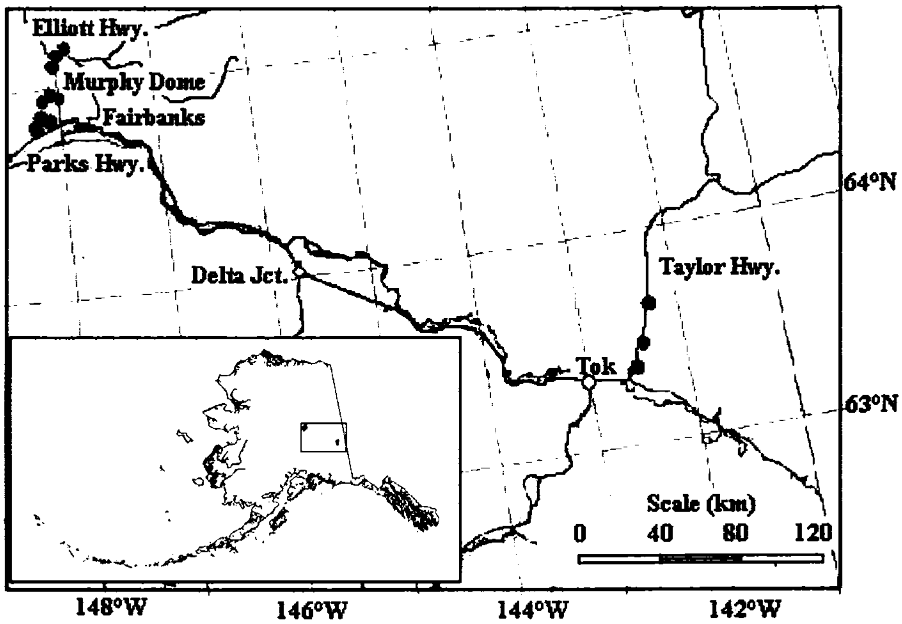


FIGURE 1.1. Distribution of the twelve study sites (solid circles) in the Tanana Valley uplands of interior Alaska. Inset figure shows the relative position of the study region in the state of Alaska, USA. Low (L; SI = 2.5-4.6 m), Medium (M; SI = 4.6-7.5 m), and High (H; SI = 8.1-12.5 m) productivity black spruce stands occur in the four study areas located off the Parks Highway (P), the Elliott Highway (E), Murphy Dome (M), and the Taylor Highway (T).

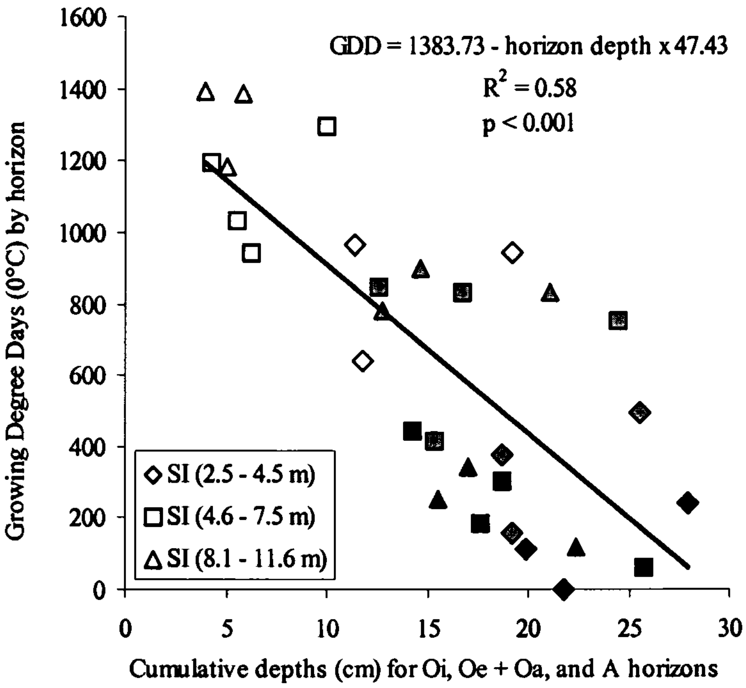


FIGURE 1.2. Growing degree-day heat sums by horizon. Open symbols represent Oi horizons, shaded symbols represent Oe and Oa horizons, and solid symbols represent the A horizons. Sites are ranked by productivity (Site Index; stand height (m) attained at 50 years).

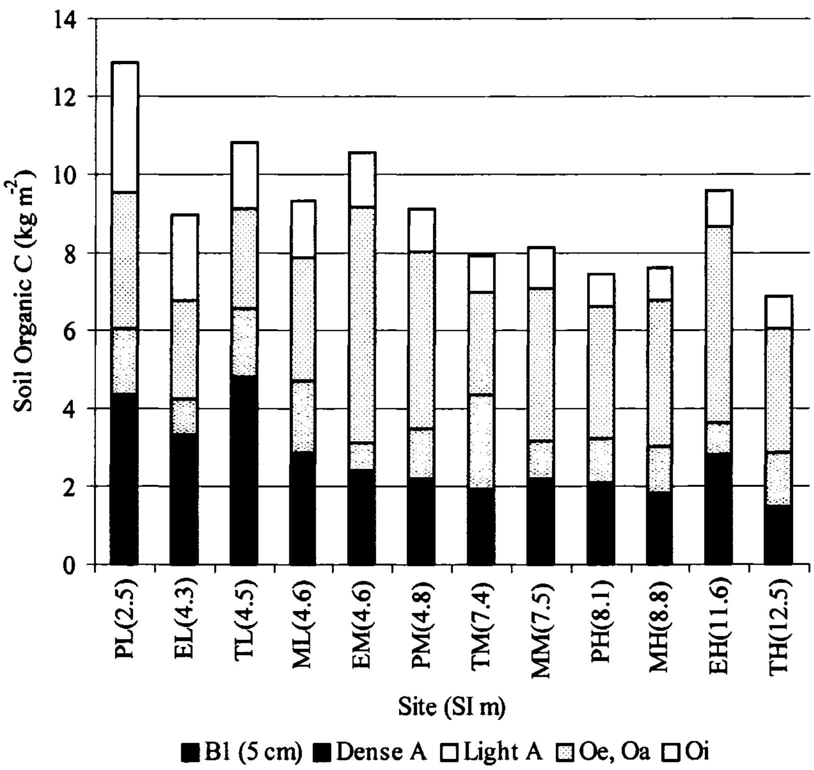


FIGURE 1.3. Soil organic carbon content (kg C $m^{-2}$) by horizon. Soil organic carbon content (kg C $m^{-2}$) for mineral (A and 5 cm B1) and organic (Oi, Oe, Oa) soil horizons across four black spruce productivity and temperature gradients. Mineral soil was fractionated into "dense" (>1.6 g $cm^{-3}$) and "light" (<1.6 g $cm^{-3}$) soil organic matter. Sites are ranked by productivity (Site Index; stand height (m) attained at 50 years).

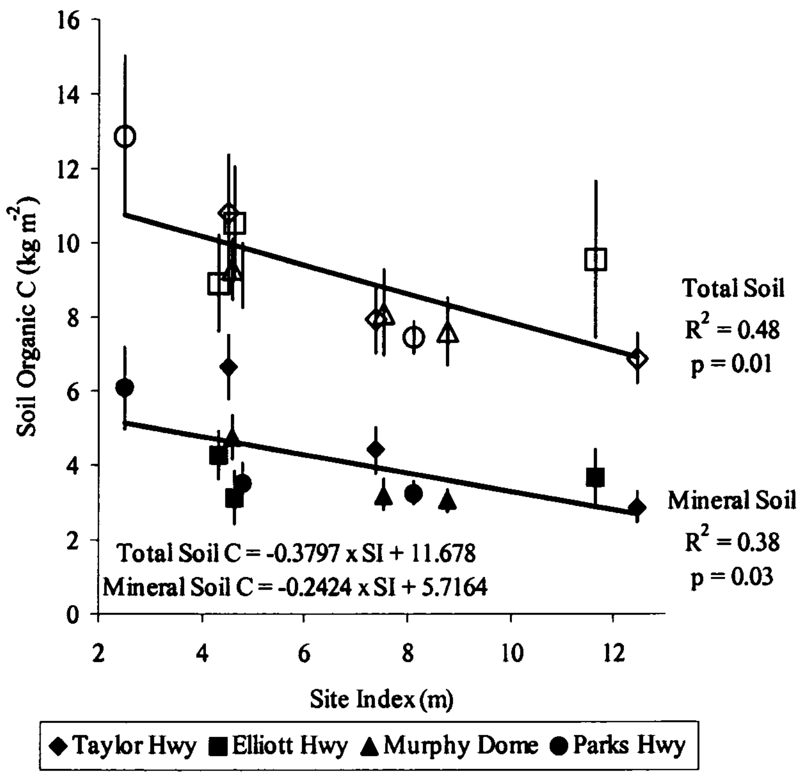


FIGURE 1.4. Total soil organic carbon content vs. stand production. Total soil organic carbon content in all horizons (open shapes), and in the mineral soil horizons (solid shapes) decreases with stand productivity (Site Index; stand height (m) attained at 50 years) across four study areas in interior Alaska. Error bars represent Standard Errors of the mean.

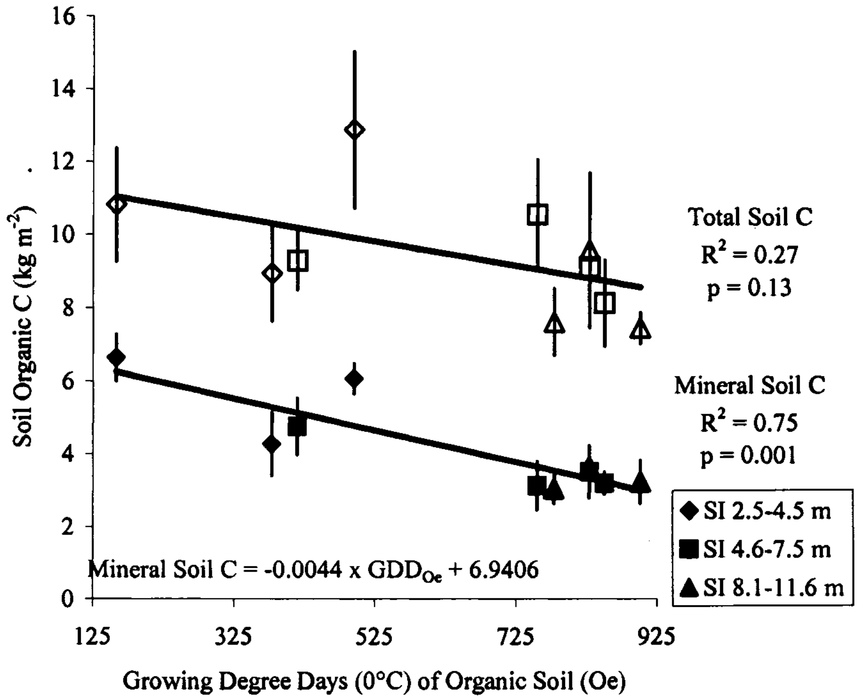


FIGURE 1.5. Total soil organic carbon content vs. soil heat sums. Soil organic carbon content in the mineral soil (solid shapes) decreases with increasing heat sum (Growing Degree Days >0 °C) in the Oe horizon across 4 black spruce productivity gradients. Total soil organic carbon content (open shapes) showed no significant change with heat sum ($\alpha = 0.05$). Stand productivity was grouped into three Site Index (SI, m) classes. Error bars represent Standard Errors of the mean.

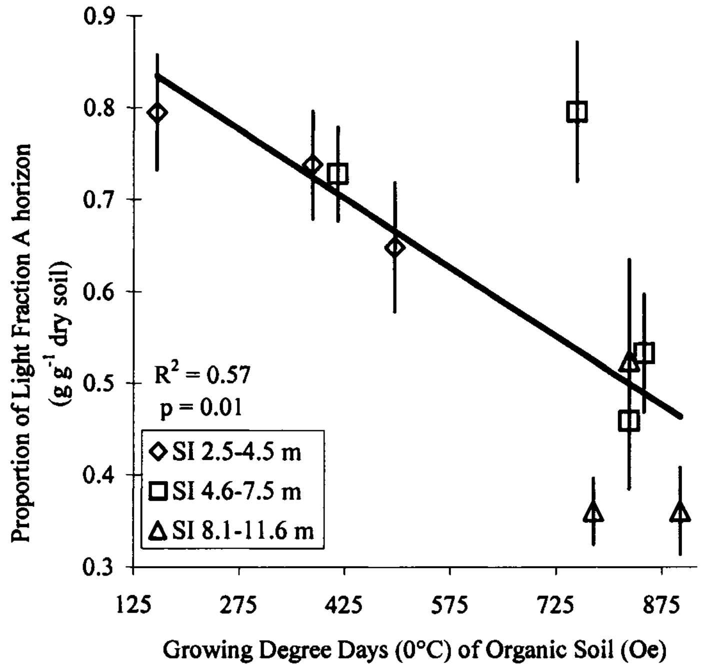


FIGURE 1.6. The proportion of light fraction A horizon (< 1.6g $cm^{-3}$) decreases with heat sum (Growing Degree Days >0 °C) in the Oe soil horizon across 4 black spruce productivity gradients. Stand productivity was grouped into three Site Index (SI, m) classes. Error bars represent Standard Errors of the mean.

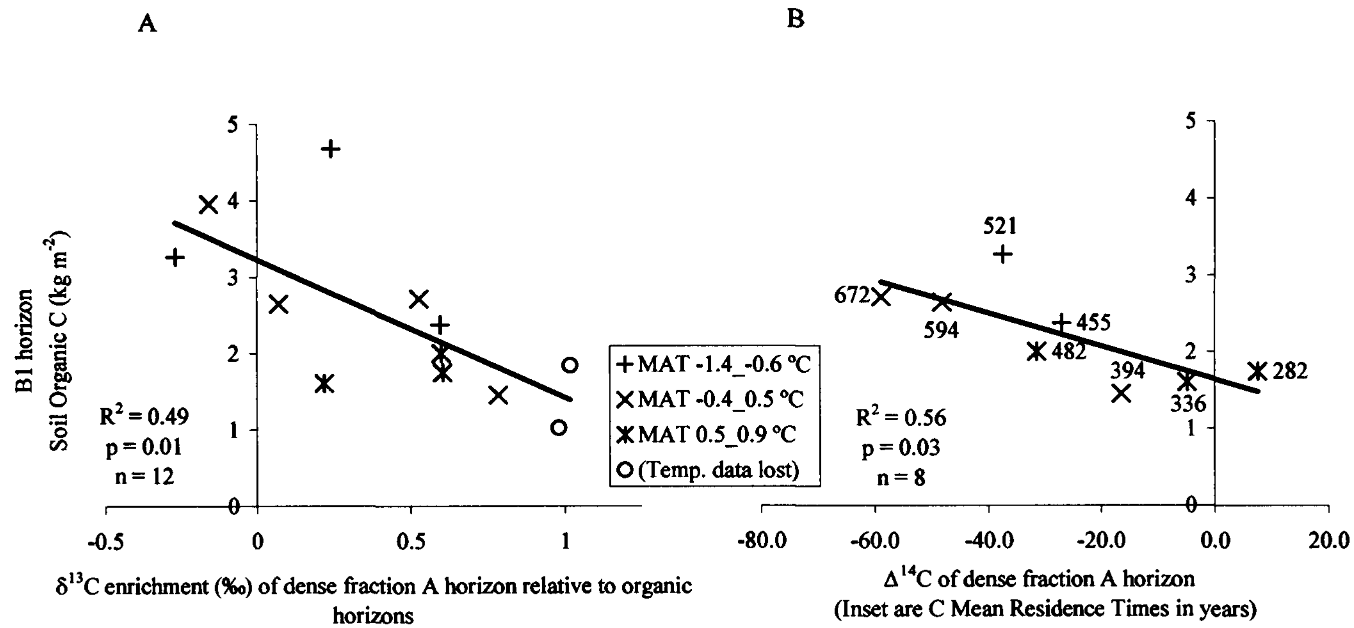


FIGURE 1.7. Carbon isotope enrichment in the dense fraction A horizon mineral soil increases as the soil organic C content of the underlying B1 mineral soil horizon decreases. A) The $\delta^{13}C$ isotope enrichment in the dense fraction A horizon mineral soil (> 1.6g $cm^{-3}$) relative to organic soil horizons vs. B1 horizon soil organic C. B) The $\Delta^{14}C$ content of dense fraction A horizon mineral soil vs. B1 horizon soil organic C. Inset numbers are Mean Residence Times (years) of C in the dense fraction, as calculated from the degree of $\Delta^{14}C$ incorporation.

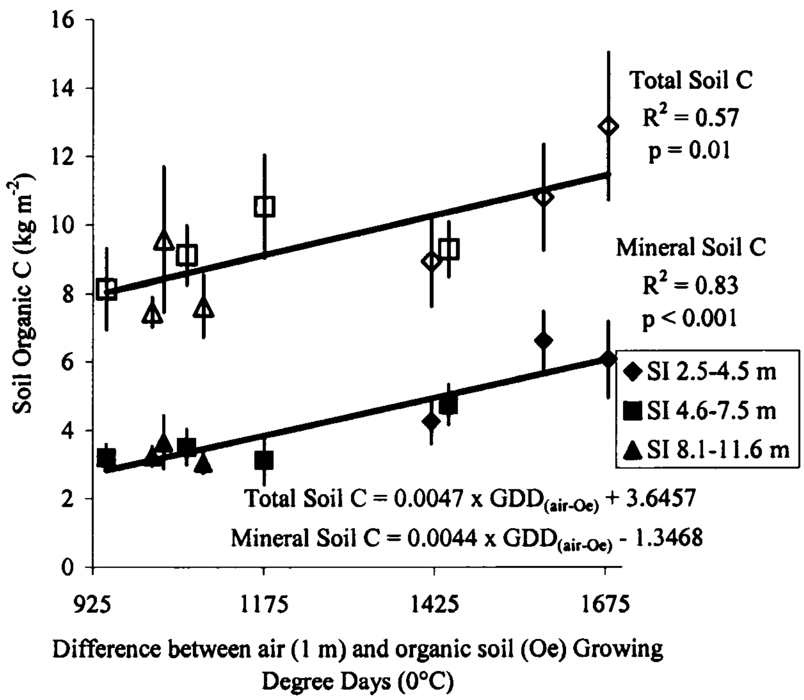


FIGURE 1.8. The difference between air (1m) and soil (Oe horizon) heat sums (Growing Degree Days >0 °C) compared with total (open shapes) and mineral (solid shapes) soil organic C storage. Stand productivity was grouped into three Site Index (SI, m) classes. Error bars represent Standard Errors of the mean.

# Chapter 2: Controls over pathways of carbon efflux from soils along climate and black spruce productivity gradients in interior Alaska

## 2.1: Abstract

Small changes in C cycling in boreal forests can change the sign of their C balance, so it is important to gain an understanding of the factors controlling small exports like water-soluble organic carbon (WSOC) fluxes from the soils in these systems. To examine this, we estimated WSOC fluxes based on measured concentrations along 4 replicate gradients in upland black spruce (*Picea mariana* [Mill.] B.S.P.) productivity and soil temperature in interior Alaska and compared them to concurrent rates of soil $CO_2$ efflux. Concentrations of WSOC in organic and mineral horizons ranged from 4.9–22.7 g C $m^{-2}$ and from 1.4–8.4 g C $m^{-2}$, respectively. Annual WSOC fluxes (4.5–12.0 g C $m^{-2}$ $y^{-1}$) increased with annual soil $CO_2$ effluxes (365–739 g C $m^{-2}$ $y^{-1}$) across all sites ($R^2 = 0.55$, $p = 0.02$), with higher fluxes occurring in warmer, more productive stands. Although annual WSOC flux was relatively small compared to total soil $CO_2$ efflux across all sites (<3%), its relative contribution was highest in warmer, more productive stands which harbored less soil organic carbon. The proportions of relatively bioavailable organic fractions (hydrophilic organic matter and low molecular weight acids) were highest in WSOC in colder, low-productivity stands whereas the more degraded products of microbial activity (fulvic acids) were highest in warmer, more productive stands. These data suggest that WSOC may be a mechanism for increased soil C loss if the climate warms and therefore should be accounted for in order to accurately determine the sensitivity of boreal soil organic C balance to climate change.

**2.2: Introduction**

It is widely accepted that positive feedbacks exist between the increasing atmospheric concentration of $CO_2$ and global warming (e.g., Chapin et al. 2000), and therefore the ability of soils to accumulate and preserve mineralizeable organic matter has received growing interest. The likely direction of change in soil organic carbon (SOC) in the boreal forest biome, which harbors the world's second largest SOC stock, is of marked concern because climate warming is projected to be greatest in high latitudes (e.g., Arctic Climate Impact Assessment 2004) and temperature is the cardinal determinant of soil C mineralization. In boreal forests, SOC balance is often the small residual of two much larger C fluxes (inputs from net primary production and losses via soil heterotrophic respiration), and therefore small changes in these two fluxes can change the sign of C balance in these systems (Goulden et al. 1998), and thereby alter the sensitivity of boreal forest C balance to climate change.

Perhaps the least studied component of ecosystem C cycling in boreal forests is water-soluble organic C (WSOC) movement with the mass flow of water from upland soils. Across a boreal landscape in northern Canada (BOREAS study), Moore (2003) showed that WSOC measured *in situ* from upland organic soil horizons ranged from 2.0-6.3 g WSOC $m^{-2}$ throughout the growing season. Although these fluxes were small relative to SOC stocks measured to 5 cm in the mineral soil in upland black spruce forests from the same study (0.01-0.10%; Trumbore and Harden 1997), accounting for this flux could offset 4.3%-45.3% of mean area-weighted SOC accumulation per year across the region (13.9-46.4 g C $m^{-2}$ $y^{-1}$;

Rapalee et al. 1998). Moreover, black spruce stands in the same study region have been shown to vary widely in their net C uptake (10 ± 50 g C $m^{-2}$ $yr^{-1}$; Goulden et al. 1998) and therefore even small soil WSOC fluxes could negate the C sink status of these systems. A better understanding of the mechanisms behind soil WSOC flux is necessary in ascertaining the sensitivity of boreal forest C balance to changing regimes of climate and primary production.

Water throughput in forest soils is largely controlled by the sorptive capacity of soil organic matter (SOM), topography, and evapotranspiration (e.g., Fisher and Binkley 2000). In black spruce forests of interior Alaska, aspect and landscape position play pivotal roles in mediating insolation, soil temperature, and stand productivity (e.g., Viereck et al. 1983). In turn, organic horizon depths diminish with increased nutrient mineralization occurring in warmer, more productive stands (Van Cleve and Barney 1981, Van Cleve et al. 1983a). This is important to consider because partially decomposed organic soils can contain >80% water by volume and therefore the depths of organic horizons exert considerable control over surface soil hydrology (Boelter and Verry 1977, Van Cleve et al. 1983b, Van Cleve et al. 1993, D'Amore and Lynn 2002). Moreover, peak soil water flows are often restricted to organic horizons because mineral soils are usually still frozen during spring snowmelt, especially when permafrost is present (Kane et al. 1992, MacLean et al. 1999). However, biophysical controls over the interaction between soil water flux and WSOC concentration in boreal forests have received little study.

Decomposition environment and leachates from plant-derived inputs may exert considerable control over the concentration of WSOC in boreal forests. In temperate systems, WSOC concentrations increase with litter fall inputs to and turnover within surface organic soil horizons (Huang and Schoenau 1996, Park et al. 2002) because of subsequent leaching of the products of microbial decay (McDowell and Likens 1988). Soil WSOC concentrations increase with microbial activity (Williams and Edwards 1993), fungal abundance (Guggenberger et al. 1994, Högberg and Högberg 2002), temperature (Christ and David 1996), and with almost any conditions that enhance SOC mineralization (Kalbitz et al. 2000, Michalzik et al. 2001). Neff and Hooper (2002) have also shown a positive correlation between $CO_2$ and WSOC fluxes in laboratory incubations of Alaskan soils, which suggests that WSOC concentrations may increase as microbial processing within soils increases. Field measurements of these fluxes and their environmental controls in boreal forests are rare and poorly understood, however.

Soil respiration increases with temperature and turnover of plant-derived inputs in boreal forests (Schlentner and Van Cleve 1985, Rustad et al. 2001, Vogel et al. 2005). Mineralization of soil WSOC also contributes greatly to total soil respiration (Qualls and Haines 1992, Neff and Asner 2001, Kalbitz et al. 2005), but the degree of WSOC mineralization in high latitudes is highly dependent on its composition (Michaelson et al. 1998). Ping and others (2001) found that microbial $CO_2$ production (and hence, activity) increased with the proportion of relatively labile soil WSOC fractions in Alaskan tundra soils. Hydrophilic neutral (HiN) fractions of WSOC generated the most $CO_2$ in incubation experiments, and were therefore the most labile (Michaelson et al. 1998). The bioactivity of other fractions decreased in the order: Hydrophobic neutrals (HoN), low molecular weight fulvic acids (LMWFA), humic acids (HA), and fulvic acids (FA). This continuum of WSOC bioactivity may be used to gain insight as to the environmental controls over WSOC fractions in the field. For example, Michaelson and Ping (2003) suggested that the ratios of hydrophilic to hydrophobic WSOC (Hi:Ho; including all neutral and acid fractions) were higher in colder, permafrost dominated sites due to temperature preservation of more labile organic matter. In addition, saturated soils in arctic Alaska, where $O_2$ diffusivity may limit microbial activity, exhibited higher HA:FA ratios than did better drained soils of the region (Ping et al. 1988, Ping et al. 2001). Quantifying relative changes in soil WSOC forms and concentrations, in addition to soil respiration, along gradients in temperature and productivity should therefore provide insight as to the mechanisms behind WSOC generation and its influence on soil $CO_2$ efflux (e.g., Weller et al. 1995, Stottlemyer 2001).

To determine how the complex interplay among stand production, nutrient mineralization, and soil temperature affects soil WSOC dynamics, we measured WSOC concentrations, composition, and soil $CO_2$ efflux along four replicate gradients in black spruce stand production and soil temperature in interior Alaska. Because turnover of SOM is faster in warmer, more productive sites, we hypothesized: 1) the products of SOM turnover, and therefore soil WSOC concentrations, increase with soil respiration in warmer, more productive stands, which suggests that 2) the lability of WSOC fractions decreases as soil respiration, temperature, and stand production increase, and that 3) fluxes of WSOC increase with soil respiration, stand production, and temperature because of diminished soil water holding capacity and increased WSOC concentrations.

## 2.3: Materials and Methods

### 2.4: *Sites and Soil Profile Descriptions*

Site establishment, temperature records, and soil profile and stand descriptions and analyses in this study were previously described in detail by Kane et al. (2005). Briefly, four replicate gradients in black spruce (*Picea mariana* (Mill.) B.S.P.) productivity and soil temperature were established across interior Alaska, ranging in latitude from 63°-65°N (~365 km) and in longitude from 142°-148°W (~550 km) (Figure 1). Study areas were named for the roads used to access them and were established (from west to east) off of the Parks Highway (P), Murphy Dome Rd. (M), the Elliott Highway (E), and the Taylor Highway (T). Each study area consisted of three sites differentiated by their relative level of stand production, or Site Index. Site Index (SI) is defined as the height of stand dominants attained after 50 years, and sites with low (L, 2.5-4.6 m), medium (M, 4.6-7.5 m), or high (H, 8.1-12.5 m) SI values occur within each study area. Each site in this study is denoted by its location and productivity level (e.g., Parks Highway, low-productivity (PL)). Site Index equations had been developed from 292 trees harvested from 33 sites across interior Alaska (Rosner 2004), of which four sites have been incorporated into this study. The utility of SI as a measure of site productivity is generally limited to mature, even-aged, undisturbed, mono-specific stands (Carmean and Lenthall 1989) and these criteria were met at all of the sites in this study. Moreover, SI has been directly linked to above and below ground biomass production in black spruce stands (Chen et al. 2002).

Each site consisted of three 20 $m^2$ plots, which were spaced 20 m apart (either in a triangle formation or perpendicular to slope, within a uniform stand type). We sought backslopes in order to control for drainage, nutrient accumulation, and nutrient mobility across sites. The primary landscape attributes selected were aspect and local relief because they determine insolation, which in turn drives changes in soil temperature and stand productivity. The colder, low-productivity stands occurred on north facing slopes and received less insolation annually. Warmer, high-productivity stands occupying south facing aspects had shallower organic horizons and harbored significantly less SOC than colder, low productivity forests (Kane et al. 2005). Intermittent permafrost occurred at the PL, EM, EL, and TL sites (Typic Historthels and Aquic Haplorthels) and Inceptisols occurred at the other sites (Cryaquepts and Cryochrepts). General soil physical and climate data are presented for each productivity level in Table 1.

Twelve soil pits (approximately 0.6 $m^2$, to a depth of 5 cm into the B horizon) were dug at each site. Genetic horizons were delineated in the field into the Oi, Oe+Oa, A and B horizons (following Schoenberger et al. 2002). Three soil cores (5.08 cm diameter) were obtained from the best face of each descriptive pit and were parsed and bulked by horizon in the field. Roots (> 2 mm diameter) were removed from organic soils by hand, and mineral soils were passed through a 2 mm sieve. Soils were dried, ground in a ball mill, and analyzed for SOC content on a LECO 2000 CNS analyzer (LECO Co., St. Joseph, Michigan) ($n = 12$). Water holding capacity was determined for organic and mineral soil horizons at each site by homogenizing field-moist soil cores on a tray in the laboratory, placing a subsample in a pre-weighed and saturated Whatman #5 (Whatman Inc., Florham Park, New Jersey) qualitative filter in a funnel, and then saturating the sample with deionized water. The sample was left until it ceased to drip water, weighed wet, dried at 105 °C, and then reweighed. Mineral soils from the A horizon were subsampled from 6 pits at each site, chosen at random, to be separated by density into fractions <1.6 g $cm^{-3}$ (light) and >1.6 g $cm^{-3}$ (dense) using sodium polytungstate solution (Sometu Co., Sherman Oaks, California), as described in detail by Baisden et al. (2002), following Golchin et al. (1994). The choice of density was based on earlier studies in which the light fraction was composed mainly of relatively undecomposed, labile organic matter whereas the dense fraction represents a mixture of more refractory material associated with the mineral soil (Golchin et al. 1994, Schulten and Leinweber 1999, Baisden et al. 2002, McLauchlan and Hobbie 2004). Soil temperatures were continuously recorded in the middle of the Oi, Oe, and 5 cm of mineral soil horizons, as well as at a depth of 10 cm (n = 4), from June 2003 through September 2004 using HOBO H8 portable data recorders at each site (Onset Co., Bourne, MA).

### 2.5: *Water-Soluble Organic Carbon*

Nine soil cores were obtained randomly on a 20 x 20 m sampling grid at each site in May, June-July, and again in September 2004. Cores were parsed in the field into organic (Oi+Oe+Oa) and mineral soil (5 cm, A+B). All cores were immediately sealed in polyethylene bags and were kept on ice in an insulated cooler while being transported to laboratory refrigerators kept at approximately 4 °C. All organic soil samples were extracted for WSOC content within 24 hours, and mineral soils were extracted within 48 hours. Water-soluble organic C concentrations from subsets of mineral soils extracted 24 and 48 hours after collection did not significantly differ ($p = 0.30$, ± 5.3 mg C $l^{-1}$).

The method for extracting WSOC from the soil followed Huang and Schoenau (1996), which was modified from McGill et al. (1986). Briefly, field moist soil samples were homogenized on a tray and roots >2 mm were removed. Then 20 g of organic soil or 40 g of mineral soil were gently shaken on a rotary table with 100 ml ($n = 9$ per horizon per date) of deionized water for 1 hour, filtered through a Whatman GF/A filter, and then passed through a Whatman 0.45 μm membrane filter. Soil extracts were preserved at pH 2 using $H_3PO_4$ and refrigerated at 4 °C prior to analysis. Each field moist soil sample was subsampled to determine moisture content (gravimetrically) and WSOC was adjusted to an oven-dry basis. Bulk density and depth measurements for each soil horizon were used to relate mg WSOC kg oven dry soil$^{-1}$ on an area basis at each site (g WSOC m$^{-2}$).

Three zero-tension lysimeters (85 x 19 cm) were installed perpendicular to slope at the organic-mineral soil interface at the end of the growing season in 2003 at the PL, PM, and PH sites. Water was harvested from these, and standing soil water was obtained where available, throughout May 2004. Three first order streams adjacent to upland sampling points were also sampled. These samples were filtered (0.45μ) and WSOC concentrations were compared to DI $H_2O$ extracted WSOC from complementary soil cores obtained nearby (< 10m, $n = 9$).

Water-soluble organic C fractions were determined in soil extracts obtained from the PL, PM, and PH sites. Three replicate samples (bulk product of 9-12 cores each) of organic and mineral soil were obtained at each site in September 2004. These soils were collected and prepared as previously described. We equilibrated 200 g of organic soil and 400 g of mineral soil (field moist) with 1 l of deionized water for ~12 hours. Samples were filtered through a 0.45 μm polysulfanone filter and concentrations of each WSOC fraction were determined on tandem XAD8-XAD4 resin columns described by Ping et al. (2001), following Malcolm (1992). In this study WSOC was separated based on sorption to and desorption from the exchange resins and sequential acid and base treatments, into four fractions (HA, soluble C at pH = 1; FA, eluted from XAD8 resin; HiN, soluble C passing XAD4; and LMWFA, eluted from XAD4 resin). Total Ho and Hi organic matter were determined by retention on XAD8 and XAD4 resins, respectively. Total WSOC and concentrations in different fractions were determined on an OI model 700 total organic carbon analyzer (OI Analytical, College Station, Texas).

**2.6: *Soil $CO_2$ Efflux***

Soil respiration measurements were made using an infrared gas analyzer (IRGA; EGM-4 gas analyzer, PP Systems, Haverhill, MA) with a dynamic soil respiration chamber (SRC-2, PP Systems). Respiration collars (18) were randomly located on three 20 x 20 m sampling grids (separate from WSOC collection) at each site. The respiration collars (10.2 cm diameter schedule 40 polyvinyl chloride) were inserted to a depth of 2.5 cm at least 1 week prior to initial measurements and were left in place for the duration of the experiment. Care was taken to ensure that the collars remained at the same depth throughout the measurement period. All vascular plants were removed from the collars prior to measuring. Measurements were made approximately bi-weekly at each site, from May through September in 2004 (n = 5 or 6). A portable thermometer was used to measure soil temperature at 10 cm below the surface concurrently with soil respiration measured at each collar. Changes in collar-specific air volumes (caused by changes in microtopography and the presence of moss) were measured directly through use of the ideal gas law by injecting 25 $cm^{-3}$ of 3500 mg $m^{-3}$ $CO_2$ into the soil respiration chamber and measuring the subsequent dilution of $CO_2$ concentration with the IRGA. All flux values have been adjusted for these collar specific volumes.

**2.7:** ***Soil Water Balance***

We used a simple water balance method, in which inflow is partitioned into export plus a change in storage (e.g., Ward 1975), to calculate the potential for soil water flux across the climate and temperature gradients. For the purposes of this model, inflow consisted solely of precipitation (P), export consisted of estimated actual evapotranspiration (AET), and the storage term (S) was based on the water holding capacity and depths of soil horizons measured at each site. The potential outflow (O) of water from the soil in each month was determined by difference, such that: $P-O-AET-\Delta S=0$.

Monthly mean precipitation values were obtained from the ten-year record (1990-2000) collected in upland black spruce stands at the Bonanza Creek Long Term Ecological Research site (http://www.lter.uaf.edu). Winter precipitation was measured for the same site over the same period by the National Water and Climate Center, SNOTEL data network (http://www.wcc.nrcs.usda.gov/snow). Precipitation was also measured throughout the growing season at each study area using standard 0.18 $m^2$ rain gauges (U.S.F.S. No. 5100-451A). Monthly means for the maximum, minimum, and mean water years from 1990-2000 were used as inflow values in calculating soil water balance on monthly timesteps. The storage term (S) was solved iteratively for each site by assuming that storage at the end of the mean water year scenario was equal to storage at the beginning of the water year (i.e., annual $\Delta S$=0 mm). Potential evapotranspiration (PET) was estimated for each site by the Thornthwaite (1948) method (Ward 1975), which used monthly mean values for temperature, precipitation, and day length (determined by latitude). Potential evapotranspiration was related to AET following Brooks et al. (1991). It was assumed here that if PET=0 mm and mean monthly temperature was below freezing, then all precipitation for a given month was snow. Monthly snow accumulated until mean monthly PET was positive. When mean monthly temperatures were not <0 °C (PET>0 mm), water equivalents of accumulated snow were treated as inflow.

The potential flux of WSOC out of the soil was estimated by multiplying the measured concentration of WSOC by outflow for a given month ($O_{mo}$), and was then summed for a water year as described in equation 1.

$$\sum_{mo=1}^{12}\left[\frac{g\,WSOC\,\mathrm{m}^{-2}{}_{mo}}{(S_{mo}+O_{mo})}\right]\times O_{mo}=g\ \mathrm{WSOC\,m^{-2}\,y^{-1}} \qquad [1]$$

Only spring and fall (April-May and September-October) WSOC concentration and O values contributed to annual flux determination because these were the only two times throughout the year that the water balance showed a surplus of soil water across all sites (soil water storage was well below capacity in the summer and there was no free water when temperatures <0 °C). We accounted for potential reductions in WSOC flux with continued flushing (e.g., Boyer et al. 1997) by applying the relationship between water and WSOC fluxes at 0-10 cm presented by Neff and Asner (2001).

**2.8:** ***Data Analysis***

Multiple regression analysis was used to develop empirical relationships for predicting soil $CO_2$ efflux at each site. Forward selection procedures were used in accepting predictor variables at $\alpha = 0.05$ and if an additional 5% of the variance could be explained by adding another factor. Soil respiration was the dependent variable, and soil temperature, moisture, elevation, and SI were predictor variables. Soil respiration was normally distributed across all sites (Shapiro-Wilk test, $W = 0.96$, $p = 0.21$) and soil respiration was natural log transformed because empirical relationships were logarithmic. Site means for each measurement date were used to develop the models. Growing season soil $CO_2$ effluxes were estimated for each site using an empirical relationship developed across all sites and measurement days. Mean daily soil temperatures and soil moisture from actual measurements on sample dates and interpolated values between sampling dates were used in the multiple regression. Winter fluxes were estimated using an equation developed for similar stands (which included the PM site) by Vogel et al. (2005). This equation related the length of the snow-free season to cumulative flux ($R^2 = 0.69$, $p < 0.05$). Soil respiration measurements at the mid-level productivity sites (EM and MM) were withheld from the model development so that they could be used in validation. Differences among sites in WSOC composition were tested using 1-way ANOVA pairwise comparisons, LSD. Pearson correlation coefficients demonstrated relationships between variables ($\alpha = 0.05$). Descriptive statistics were performed with Analyze-it statistical module (Analyze-it Software, Ltd. Leeds, UK) and regressions were developed with PC SAS (version 8.2, SAS institute, Cary, North Carolina).

**2.9: Results**

**2.10: *WSOC Concentrations***

Mean [WSOC] in organic soil horizons were an order of magnitude higher than those measured in surface mineral soils (Figure 2). There were no trends between organic and mineral soil horizon [WSOC] throughout the growing season of 2004. Soil [WSOC] in organic horizons increased by factors of two and five at two stands (TM and TH, respectively) after forest fires burned them in June (Figure 2).

Mean total [WSOC] in organic and mineral soils increased with soil heat sums (growing degree days >0 °C, GDD) across all sites ($R^2 = 0.54$, $p = 0.02$). Stand production (SI, m) also increased with GDD in organic horizons across all sites ($r = 0.69$, $p < 0.05$), and [WSOC] in organic and mineral soils was higher in the warmer, more productive stands. Interaction between SI and GDD did not explain any additional variance in [WSOC]. Correlations between [WSOC] and SI in the organic ($r = 0.65$, $p = 0.04$) and mineral ($r = 0.93$, $p < 0.001$) soil horizons were most significant in the fall and summer, respectively. The amount of WSOC per unit SOC (g WSOC kg $SOC^{-1}$) also increased with GDD and stand production across all sites (Figure 3).

**2.11: *WSOC Composition and Soil Decomposition***

The most bioavailable fraction, HiN, dominated WSOC extracted from mineral and organic soil horizons in the Parks Highway sites (PL, PM, and PH; Figure 4). The proportion of Hi to Ho in organic and mineral horizons was highest in the cold, low productivity stand (PL). The most degraded fraction isolated, FA, comprised the lowest percentage of total WSOC in the PL site. The proportion of HA to FA was greater at the PL site than at the warmer, more productive PH site in both the organic (4x) and mineral (2x) soil horizons (Figure 4).

Soil WSOC expressed on an area basis (g WSOC $m^{-2}$) was highest in the most productive stands exhibiting the most decomposed SOM (Figure 5). Organic horizon WSOC measured in the fall decreased as the proportion of relatively undecomposed fibric SOC increased across all sites ($r = -0.76$, $p = 0.01$). Similarly, mineral horizon WSOC measured in the summer decreased as the proportion of relatively undecomposed light fraction A horizon (<1.6 g $cm^{-3}$) increased across all sites ($r = -0.75$, $p = 0.01$).

**2.12: *Soil $CO_2$ Efflux***

Site means in soil respiration measured from May-September, 2004 increased with SI ($r = 0.67$, $p = 0.05$; Table 2) across all sites. Soil temperature (°C), SI (m), elevation (m), and soil moisture (g $g^{-1}$) were all significant predictor variables for natural log transformed soil respiration, explaining an additional 39%, 21%, 8%, and 6% of the respective variance upon their inclusion (Table 2; $R^2 = 0.59$, $p < 0.001$). There was an interaction between elevation and soil moisture variation ($F = 5.58$, $r = 0.36$, $p = 0.02$) and between SI and soil moisture ($F = 7.51$, $r = -0.36$, $p = 0.008$). A weak interaction between soil moisture and soil temperature occurred ($F = 3.61$, $p = 0.06$), with warmer temperatures occurring in drier sites. There was no interaction between soil moisture and elevation, elevation and SI, or elevation and soil temperature measured throughout the growing season.

Measurements withheld from the multiple regression determination (EM and MM sites) were well within the prediction interval ($\alpha = 0.05$) of modeled vs. observed values ($R^2 = 0.70$, $p < 0.001$; Figure 6). Mean modeled soil $CO_2$ efflux throughout the growing season (Table 2) and for days on which measurements were taken (Figure 6) closely tracked observed values across all sites ($R^2 = 0.71$, $p = 0.004$ and $R^2 = 0.59$, $p < 0.001$, respectively).

**2.13: *Soil Water Flux***

Although total snowfall throughout the winter of 2003-2004 was average (water equivalent of 123 mm fell from November 2003 through April 2004 at Bonanza Creek LTER; Brian Charlton, personal communication), the following summer in which WSOC samples were collected was one of the driest on record, with an average of 12.0 ± 1.4 mm of precipitation measured throughout June and July across the study region. For this reason, mean monthly precipitation values over the period 1990-2000 (347 mm $y^{-1}$) were used in contrasting seasonal water balances across all sites (Table 3). Monthly mean precipitation only exceeded the storage capacity of the organic soil and losses through estimated actual evapotranspiration in the spring and in the fall months. Most of the estimated annual outflow (93%) occurred in April at most of the sites. At the Elliott Highway study area, however, cooler spring temperatures resulted in most of the annual outflow (95%) occurring in May. Since most potential soil water flux occurred early in the season, the effects of summer drought in 2004 on total WSOC fluxes are likely to be minimal. The surface mineral soil was frozen throughout the spring (Table 1), and therefore water flux was only calculated for organic horizons. At all of the sites, soil water storage was below capacity in June, July, and August. Organic matter depth (and hence, soil water storage) drove estimated mean annual soil water outflow ($r = -0.70$, $p = 0.03$). The storage term decreased as SI increased across all sites (Table 3; $r = -0.74$, $p = 0.01$), but no trends were observed between outflow and SI.

Estimated annual outflow for the maximum precipitation water year observed over the same period (470 mm $y^{-1}$) increased significantly, as inputs exceeded the soil water storage capacity more often (Table 3). In the dry water year scenario (215 mm $y^{-1}$) soil water storage decreased across all sites and outflow was significantly reduced.

**2.14: *WSOC concentrations, fluxes, and $CO_2$ efflux***

Cumulative soil $CO_2$ efflux estimates for the snow free part of the year ranged from means (± standard error) of 379 ± 39 g C $m^{-2}$ in the cold, low productivity stands to 563 ± 69 g C $m^{-2}$ in the warmer, high productivity stands. Cumulative winter soil respiration estimated from the duration of snow cover at each site (following Vogel et al. 2005) ranged from 37.0-42.5 g C $m^{-2}$, and annual soil $CO_2$ efflux estimates ranged from 360-739 g C $m^{-2}$ $y^{-1}$. Mean total soil WSOC stocks increased with cumulative soil $CO_2$ evolution throughout the growing season, with higher soil respiration and WSOC occurring in warmer, more productive stands (Figure 7). This relationship was significant even when the EH site (growing season efflux of 700 g C $m^{-2}$) was removed from the analysis ($R^2 = 0.51$, $p < 0.05$).

Soil [WSOC] measured in standing surface water pools and tensionless lysimeters in May were about 64% lower than [WSOC] in complementary soils extracted with DI $H_2O$

(Figure 8). Therefore, the relationship between observed and extracted soil [WSOC] (Figure 8) was used to scale down potential soil WSOC fluxes estimated by soil water extraction (Equation 1). Potential WSOC fluxes decreased seasonally from a mean (± standard error) of 5.7 ± 0.6 g C $m^{-2}$ through May and June to 2.0 ± 0.5 g C $m^{-2}$ through August and September across all sites. Annual WSOC flux estimates ranged from 4.5 to 12.0 g C $m^{-2}$ $y^{-1}$ and increased with annual soil $CO_2$ effluxes, with larger and more varied fluxes occurring in warmer, more productive stands (Figure 9). Estimated decreases in WSOC fluxes with cumulative outflow were minor (~8%). Annual WSOC fluxes were not correlated to soil temperatures, SI, slope, elevation, or SOC across all sites.

**2.15: Discussion**

**2.16: *WSOC Concentration and Soil Respiration***

Growing season concentrations of WSOC in organic and mineral soils in this study (Figure 2) were comparable to those reported by Huang and Schoenau (1996), who used the same extraction procedure in aspen forest soils of north-central Canada. In their study, WSOC increased in the end of the growing season, which they attributed to the leaching of soluble organic matter from increased plant litter inputs occurring in the fall. We observed the most significant increase in organic horizon WSOC with stand productivity in September (Figures 3 and 5), when inputs from understory leaf litter and root turnover to the soil would be the highest. The most significant increase in mineral soil WSOC with stand productivity was in the middle of summer, when root production and activity in black spruce stands is the highest (Ruess et al. 2003, Vogel et al. 2005). Moreover, WSOC stocks in organic and mineral soil horizons were highest in warmer, more productive stands where SOM was most decomposed (Figure 5). These data support the expectation that soil WSOC concentrations increase as plant derived inputs to and decomposition within the soil increases.

Empirical estimates of soil respiration along climate and stand productivity gradients in this study were consistent with other values reported for black spruce forests of interior Alaska (Schlentner and Van Cleve 1985, Ruess et al. 2003), and nearly encompassed the range of fluxes reported in the literature for this forest type (e.g., Vogel et al. 2005). Since stand productivity depends on nutrient release from decomposing organic matter, which increases with soil temperature (e.g., Van Cleve et al. 1983a), it is not surprising that soil respiration increased with stand production and temperature ($p < 0.001$) across all sites. The interaction between elevation and moisture was also a predictor of soil respiration across all sites ($p = 0.01$), mostly because elevation explained fluctuations in soil moisture ($r = 0.36$, $p = 0.02$), which have been shown to limit soil respiration (Fierer and Schimel 2002, Kane et al. 2003). Other studies have shown that the influence of soil moisture on soil respiration within black spruce stands depends mostly on its interaction with soil temperature (Schlentner and Van Cleve 1985, Ruess et al. 2003), and we observed a weak interaction between these factors.

It is important to note that $CO_2$ and WSOC fluxes primarily occur at different times of the year; the former occurs mainly in the summer and fall and the latter mainly in the spring. Summer WSOC pools probably play a greater role in $CO_2$ generation, whereas the water-soluble products of microbial and root activity may exert more control over WSOC concentrations and fluxes in the spring and fall. Notwithstanding, increased mean

WSOC pools occurring with higher soil respiration throughout the growing season (Figure 7) suggests that microbial and root activity plays a role in WSOC generation. Therefore it is valuable to compare annual fluxes of $CO_2$ and WSOC (Figure 9) because cumulative soil respiration integrates total microbial and root activity, which play different roles in WSOC generation throughout the year.

The largest increases in soil [WSOC] were observed in organic horizons after forest fires at two sites on the Taylor Highway (TM and TH; Figure 2). This is consistent with Shibata et al. (2003), who reported a 50% increase in [WSOC] in gravitational water collected from surface organic soil, with no change in mineral soil [WSOC], following wildfire in an upland black spruce stand in interior Alaska. The lack of a change in mineral soil [WSOC] is probably due to a combination of rapid mineralization and increased export, since water storage in organic horizons is diminished following fire. Owing to the marked increase in organic horizon [WSOC] following fire, elevated soil WSOC export, though short lived (Petrone 2005), likely comprises a significant amount of total C losses in the post burn environment in boreal forests. However, the bioavailability of fire affected [WSOC] and its contribution to total soil C evasion is not known.

**2.17:** ***Soil Water Balance and WSOC flux***

Since organic horizon depth and water holding capacity both determined soil water storage in this study, and both of these factors decreased with increasing stand production and soil temperature, it is not surprising that storage decreased with increasing SI (Table 3). The co-occurrence of lower soil water storage and higher WSOC stocks in warmer, more productive stands (Table 3, Figure 5) apparently contributed to higher WSOC fluxes (Figure 9).

Snowmelt infiltration into mineral soil is controlled in part by the previous fall's precipitation and therefore soil moisture content upon freezing. If the soil freezes at a high moisture content, then pores are blocked by ice and spring infiltration is inhibited (Kane 1980). However, if fall soil moisture near the surface is low then snowmelt may readily infiltrate mineral soils that are essentially freeze-dried and exert considerable attraction for water. It is unlikely that significant amounts of WSOC moved into mineral soils in the spring of 2004 because mineral soils were frozen at the time of estimated peak outflow across all sites (Table 1 and Table 3) and precipitation from the previous fall (2003) had been typical. These findings are consistent with the work of Rovansek et al. (1996), who used a similar simple water balance approach to demonstrate that water movement in surface horizons only occurred immediately following snowmelt in the Alaskan arctic. Moreover, they showed that water flow through mineral soil was a minor part of the water balance, totaling less than 1% of annual ET. Similarly, Carey (2003) showed that most (69%) of the annual WSOC exported from a subarctic catchment in Canada was controlled by the snowmelt flush through surface organic horizons.

Since most of the estimated outflow across all sites occurred during snowmelt, changes in winter precipitation would likely have the biggest impact on annual soil water movement. For example, outflows during the high water year (470 mm) scenario were higher than the mean water year scenario (Table 3), but mostly occurred in the spring. During the dry water year (215 mm) scenario, outflows declined from the 10-year mean as expected, and only occurred in the spring. However, combined changes in temperature and precipitation could have neutral effects on outflows, depending on the directions of change. For example, increasing mean monthly temperatures by 5 °C resulted in roughly a 36% decrease in soil water outflows, mainly because of increased estimated AET, and therefore increased drought earlier in the spring. On the other hand, the co-occurrence of warming and increased precipitation (such as in the high water year scenario) resulted in small increases in estimated outflows (7-10%), which occurred in the spring. These analyses suggest that changes in the proportion of annual precipitation occurring as snowfall and the length of the snow-free season exert major control over annual soil water balance.

Since the concentration of WSOC in streams typically increases with water throughput in boreal watershed studies (e.g., Stottlemyer 2001, Schindler et al. 1997, Carey 2003), it follows that WSOC fluxes would increase in high water years, barring any major decline in WSOC flux with continued flushing (Boyer et al. 1997). Across all sites, a 37% change in annual precipitation evenly distributed across all months resulted in a change in estimated annual outflow by approximately 54% (Table 3), which could cause changes in mean (± standard error) annual WSOC fluxes across all sites from 9.7 ± 0.9 to 5.9 ± 0.8 g C $m^{-2}$ $y^{-1}$ in high and low water years, respectively (Figure 9). Although implications of changing precipitation regime and temperature on WSOC are hard to ascertain because of seasonal changes in soil water balance, Alaska's climate is projected to become considerably warmer and wetter in the next century (Arctic Climate Impact Assessment 2004). These data suggest that warmer, wetter conditions (especially increased snowfall) could increase soil WSOC fluxes in boreal forests.

**2.18: *WSOC Composition and Soil Organic C***

Changes in WSOC composition along the Parks Highway stand productivity and climate gradient reflect soil decomposition environment (Figure 4). The release of Hi organic matter has been closely related to warmer periods when microbial activity is high (Bourbonniere 1989) and Hi organic matter has more fractions directly produced by microbes than does Ho organic matter (Guggenberger et al. 1994). However, the relationship between Hi WSOC and microbial activity can be confusing because Hi WSOC is both a product of (McKnight et al. 1985) and a highly labile substrate for microbial activity (Qualls and Haines 1992, Michaelson et al. 1998, Dai et al. 2000). We suggest that the decreasing Hi:Ho from the colder, low productivity site (PL) to the warmer, more productive site (PH) reflects temperature protection of more labile substrates (i.e., arrested decomposition at cooler temperatures; Michaelson and Ping 2003). The PH site had higher soil respiration and lower proportions of the more labile LMWFA compounds than the cooler, less productive PL site. Furthermore, the HA:FA ratio, which reflects the potential for soil microbial activity (Ping et al. 2001, following Orlov 1996), was highest in the PL site. The proportion of FA increased significantly with stand productivity, which indicates a higher degree of WSOC degradation (i.e., humification; Dai et al. 2000) at the warmer, more productive PH site (Figure 4).

Since WSOC concentrations increased with SI and temperature, the absolute amounts of more labile fractions were roughly constant across the Parks Highway gradient even though the proportion of labile fractions decreased with increasing SI and temperature. Most of the increase in soil WSOC concentration measured at the PH site was attributed to an increase in FA (7.2 vs. 1.6 g C $m^{-2}$ (FA) for the PH and PL sites, respectively). These data suggest that internal cycling of more labile fractions was higher in the warmer, more productive stand whereas degradation of WSOC was retarded in the cooler, low-productivity site (e.g., Kaiser et al. 2002).

The lack of a relationship between mineral SOC and WSOC flux from the organic horizons suggests that eluviation of surface WSOC has little effect on soil C accumulation in underlying horizons. Likely factors contributing to this include WSOC composition and mineralization (Michaelson et al. 1998, Moore and Matos 1999, Cleveland et al. 2004), restricted thaw depth during peak outflow (MacLean et al. 1999, Carey 2003), and the low sorptive capacity for WSOC within incipient, loess-dominated mineral soil (Ugolini et al. 1987, Dahlgren and Marrett 1991, Neff and Asner 2001). Additionally, there is some evidence that the sorption of different WSOC forms in Inceptisols decreases with SOC content, and that the sorption of Ho, though preferential, competes with Hi (Kaiser et al. 1996). Higher Hi:Ho ratios in soils with more SOC (e.g., the cold, low productivity site) could partially explain the lack of a relationship between WSOC flux and SOC. These data suggest that most WSOC was not stabilized in mineral soil, and therefore probably mineralizes rapidly (Kalbitz et al. 2005).

**2.19:** ***Soil Respiration and WSOC flux***

The interaction between soil temperatures, stand productivity (Figure 3), and soil decomposition environment (Figure 5) apparently controlled the stocks (Figure 7), lability (Figure 4), and potential flux of WSOC within soil horizons (Figure 9).

However, WSOC export from upland spruce forests in interior Alaska is less clear because soil decomposition environment likely influences WSOC mineralization. Increased soil $CO_2$ efflux in stands with higher WSOC flux estimates (Figure 9) and a marked decline in WSOC concentrations from uplands to adjacent streams (Figure 8) suggest that a significant amount of WSOC mineralizes as it moves down slope with the mass flow of water. For example, in the Caribou Poker Creek Research Watershed in interior Alaska, soil WSOC concentration decreased markedly from upland to valley bottom dominated by black spruce (12~62 vs. ~7 mg C $l^{-1}$, respectively; MacLean et al. 1999, Shibata et al. 2003). Additionally, Schindler et al. (1997) showed that WSOC outflow from Canadian mixed boreal forests (29.0-41.9 g C $m^{-2}$ $y^{-1}$ in 1972-1974 and 19.7-19.9 g C $m^{-2}$ $y^{-1}$ in 1988-1990) mineralized considerably in adjacent lakes, with *in situ* losses accounting for 21-82% of WSOC inflows. Kling (1991) also suggested that much of the $CO_2$ released from Alaskan arctic water bodies was from the mineralization of terrestrial WSOC, based on evidence from stream and soil lysimeter $pCO_2$ data. However, exactly how much WSOC is mineralized along a continuum from uplands to stream catchments is unclear because WSOC is both the product of and a substrate for microbial activity. Notwithstanding, these data suggest that WSOC contributes more to total soil C evasion in warmer, more productive upland stands.

**2.20: Conclusions**

We suggest that interactions among soil temperature, stand productivity, and soil decomposition extent largely control WSOC concentration, its lability, and the potential flux of WSOC within soil horizons in upland black spruce forests of interior Alaska.

Annual fluxes of WSOC and $CO_2$ and their driving factors were highly correlated, and the contribution of WSOC flux to total soil C evasion was highest in warmer, more productive stands, which contained the least amount of soil organic carbon. These data suggest that WSOC may be a mechanism for increased soil C loss if the climate warms. Further research tracking the fate of WSOC, as related to the biophysical drivers of its lability and mobility, is necessary in order to accurately determine the sensitivity of boreal SOC balance to climate change.

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Table 2.1. General soil physical and chemical properties. Values are means by low, medium, and high stand productivity (Site Index, m) level. Values in parentheses are standard errors of the mean.

| Site (SI, m)[d] | Horizon[a] | Depth (cm) | Bulk density ($g\ cm^{-3}$) | WHC[b] (mm) | Growing Degree Days[c] Spring | Fall | Annual | Total C ($kg\ C\ m^{-2}$) |
|---|---|---|---|---|---|---|---|---|
| Low | Organic | 19.7 (1.4) | 0.07 (0.03) | 114.0 (10.0) | 8 | 23 | 344 | 5.04 (0.64) |
| (SI 2.5-4.5 m) | A | 21.8 (1.5) | 0.42 (0.04) | | | | | 1.78 (0.35) |
| | B1 | 26.8 | 0.99 (0.13) | 27.0 (4.1) | 0 | 8 | 118 | 3.66 (0.45) |
| Medium | Organic | 16.6 (1.1) | 0.08 (0.02) | 76.1 (10.8) | 41 | 105 | 712 | 5.37 (0.54) |
| (SI 4.6-7.5 m) | A | 18.8 (1.1) | 0.37 (0.04) | | | | | 1.58 (0.27) |
| | B1 | 23.8 | 1.02 (0.12) | 29.7 (1.1) | 0 | 102 | 248 | 1.99 (0.30) |
| High | Organic | 15.5 (1.0) | 0.07 (0.01) | 60.1 (4.0) | 87 | 100 | 836 | 4.68 (0.58) |
| (SI 8.1-11.6 m) | A | 17.8 (1.0) | 0.42 (0.02) | | | | | 1.53 (0.20) |
| | B1 | 22.8 | 1.11 (0.14) | 50.4 (4.6) | 0 | 84 | 239 | 1.68 (0.25) |

[a] Horizons include all of the organic layers (Oi, and Oe+Oa), the A horizons, and 5 cm of B horizon (n = 48).

[b] Water Holding Capacity for combined mineral and organic horizons (n = 8).

[c] Growing Degree Day heat sum (0°C) measured in the middle of the Oe and in 5 cm of mineral soil horizons.

[d] Site Index (SI) is the height (m) attained by dominant trees in a stand at 50 years of age.

Table 2.2. Soil $CO_2$ efflux (μmol $m^{-2}$ $s^{-1}$). Measurements occurred from May through September, 2004 and are compared to values empirically modeled for the same days on which measurements were taken ($R^2 = 0.71$, $p = 0.004$). Growing season means are presented (n = 5 or 6).

| | Mean Respiration (μmol $CO_2$ $m^{-2}$ $s^{-1}$) | | |
|---|---|---|---|
| Site (SI, m)[a] | y (SE)[b] | $\hat{y}$[c] | SSR |
| PL (2.5) | 2.79 (0.50) | 2.75 | 1.48 |
| EL (4.3) | 2.14 (0.31) | 2.35 | 0.54 |
| TL (4.5) | 3.64 (0.44) | 3.43 | 0.37 |
| ML (4.6) | 1.68 (0.31) | 1.90 | 0.60 |
| EM (4.6)[d] | 3.01 (0.53) | 3.00 | 2.13 |
| MM (7.5)[d] | 3.18 (0.51) | 3.17 | 1.69 |
| PH (8.1) | 3.04 (0.34) | 3.36 | 0.31 |
| MH (8.8) | 3.97 (0.60) | 2.85 | 0.60 |
| EH (11.6) | 4.19 (0.75) | 4.38 | 3.63 |

[a] Site Index (SI) is the height (m) attained by dominant trees in a stand at 50 years of age.

[b] Mean soil $CO_2$ efflux (μmol $m^{-2}$ $s^{-1}$) measured from May -September, 2004. Values in parentheses are standard errors of the mean.

[c] Mean modeled soil $CO_2$ efflux (μmol $m^{-2}$ $s^{-1}$) estimated for days when measurements were taken using percent moisture (MP, g $g^{-1}$), SI (m), elevation (elev, m), and mean daily temperature (T, 10 cm, °C) at each site: $\ln \hat{y} = T \times 0.126 + SI \times 0.089 + MP \times 0.009 + elev \times 0.001 - 1.298$ ($R^2 = 0.59$, $p < 0.001$, SE = 0.35, n = 41). SSR = the sum of squares of the residual for y vs. $\hat{y}$.

[d] The mid-productivity sites were not included in the creation of the multiple regression so that observed values from these sites could be used in validation.

Table 2.3. Water output (mm) from the organic soil horizons. Water output (mm) from the organic soil horizons uses a simple water balance model operating on monthly timesteps. Values are reported for a mean, maximum, and minimum water year observed in interior Alaska between 1990-2000.

| Site | Outflow[a] | | | | | Evapotranspiration | | | | Storage | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| (SI, m)[e] | Mean | | | | | | AET[b] | | | | Δ storage[d] | |
| | Spring | Fall | Total | Max. | Min. | PET[c] | Mean | Max. | Min. | Initial | Max. | Min. |
| PL (2.5) | 130 | 0 | 130 | 193 | 69 | 591 | 216 | 265 | 164 | 21 | 12 | -18 |
| EL (4.3) | 75 | 2 | 77 | 154 | 10 | 538 | 270 | 316 | 221 | 25 | 0 | -16 |
| TL (4.5) | 132 | 22 | 154 | 228 | 80 | 511 | 193 | 242 | 141 | 25 | 0 | -6 |
| ML (4.6) | 152 | 5 | 157 | 231 | 91 | 542 | 189 | 238 | 137 | 21 | 0 | -13 |
| EM (4.6) | 73 | 1 | 74 | 151 | 8 | 550 | 272 | 318 | 223 | 21 | 0 | -16 |
| PM (4.8) | 148 | 9 | 157 | 231 | 87 | 547 | 190 | 238 | 137 | 15 | 0 | -9 |
| MM (7.5) | 158 | 13 | 171 | 245 | 97 | 530 | 175 | 224 | 123 | 12 | 0 | -5 |
| PH (8.1) | 146 | 10 | 156 | 230 | 84 | 552 | 190 | 239 | 138 | 11 | 0 | -8 |
| MH (8.8) | 150 | 14 | 164 | 238 | 89 | 546 | 182 | 231 | 130 | 9 | 0 | -4 |
| EH (11.6) | 75 | 8 | 83 | 160 | 11 | 539 | 264 | 309 | 214 | 14 | 0 | -10 |

[a] Water balance is solved for mean (347 mm), maximum (470 mm), and minimum (215 mm) observed water years (1990-2000) such that precipitation = outflow + AET + Δ storage.

[b] Actual evapotranspiration

[c] Potential evapotranspiration calculated by the Thornthwaite method.

[d] Initial storage term was determined by horizon depth and its water holding capacity. Change in storage in the mean water year scenario was assumed to be zero. Change in storage in maximum and minimum water years is deviation from storage in the mean water year scenario.

[e] Site Index (SI) is the height (m) attained by dominant trees in a stand at 50 years of age.

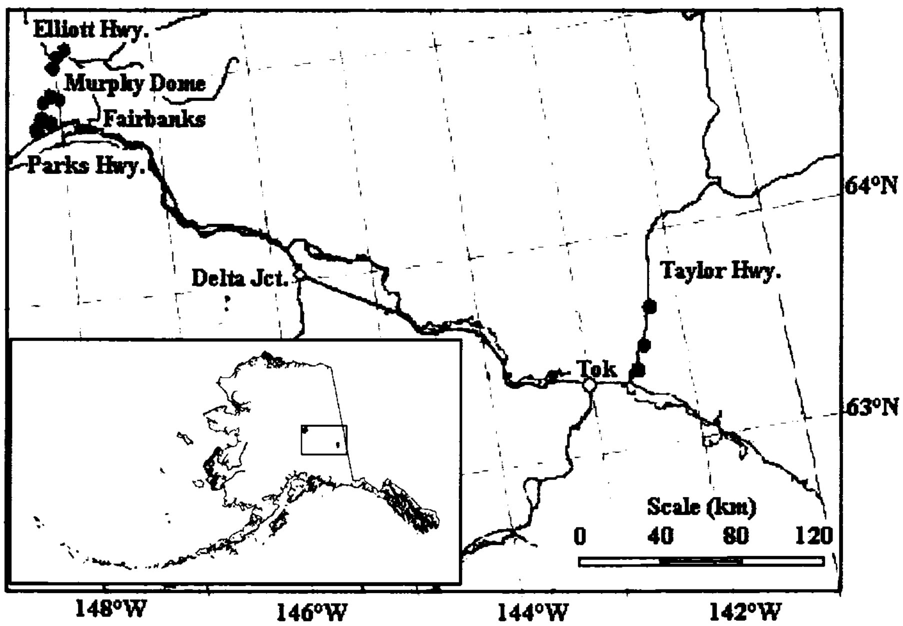


FIGURE 2.1. Distribution of the twelve study sites (solid circles) in the Tanana Valley uplands of interior Alaska. Inset figure shows the relative position of the study region in the state of Alaska, USA. Low (L; SI = 2.5-4.6 m), Medium (M; SI = 4.6-7.5 m), and High (H; SI = 8.1-12.5 m) productivity black spruce stands occur in the four study areas located off the Parks Highway (P), the Elliott Highway (E), Murphy Dome (M), and the Taylor Highway (T).

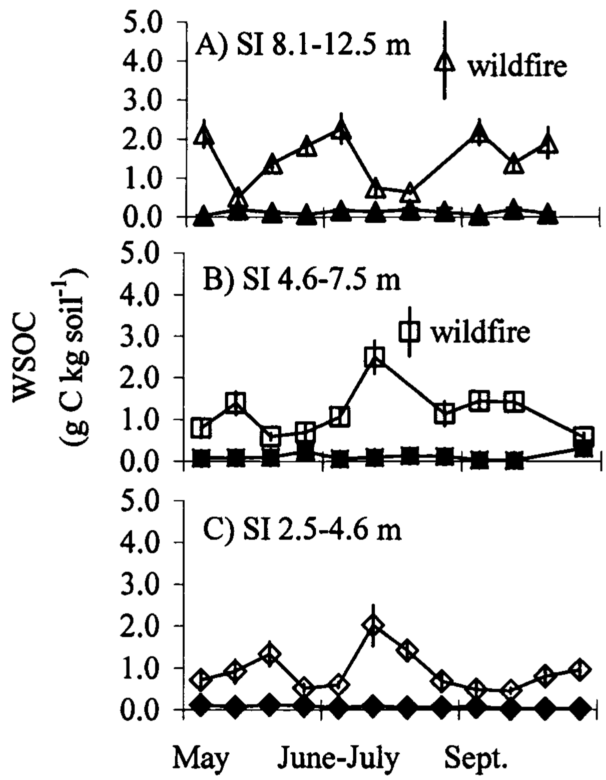


FIGURE 2.2. Water soluble organic carbon concentrations (g WSOC kg soil$^{-1}$). Measurements were for combined organic horizons (Oi, Oe, Oa; open shapes) and for 5 cm of mineral soil (A + B; opaque shapes) throughout the growing season of 2004. Wildfires occurred at the TM and TH sites in June and samples were not obtained from these sites in September. Error bars are standard errors of the mean.

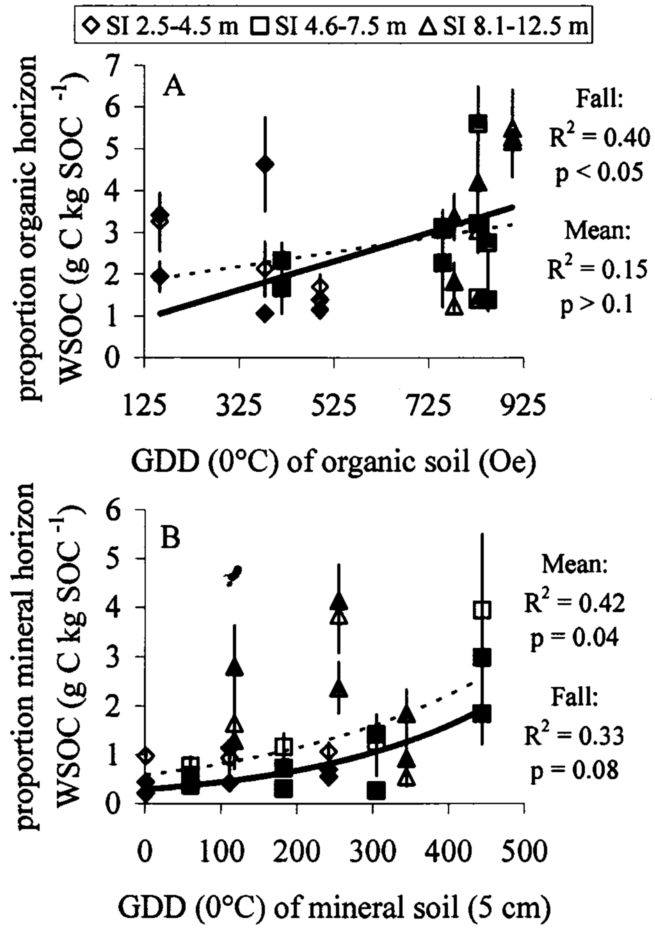


FIGURE 2.3. Water-soluble organic C per unit SOC (g WSOC kg SOC$^{-1}$). Concentrations increased with soil heat sums (growing degree days; GDD) in organic (A) and mineral (B) soil horizons. Sites are grouped by their relative level of stand productivity, with high (triangles), medium (squares), and low (diamonds) SI (m) values occurring in each study area. Samples were collected in the spring (open), summer (shaded), and fall (opaque). Error bars are standard errors of the mean.

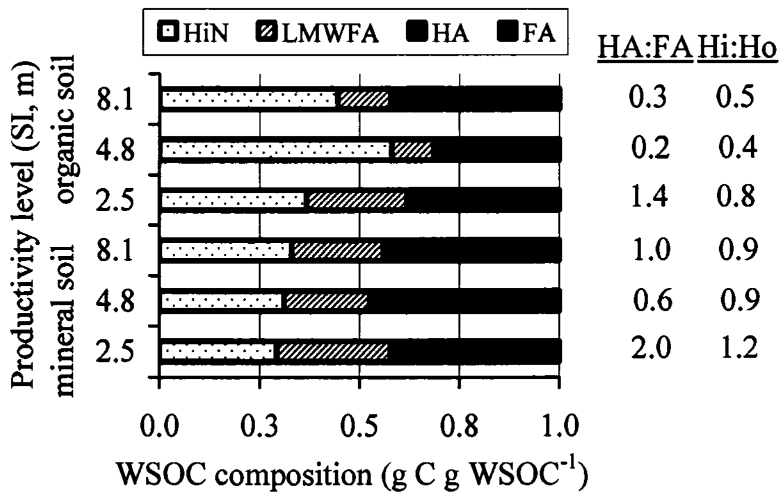


FIGURE 2.4. Percentage of HA, FA, LMWFA, and HiN. Percentage of humic acids (HA), fulvic acids (FA), low molecular weight fulvic acids (LMWFA), hydrophilic neutrals (HiN), and the ratio of total hydrophilic (Hi) to hydrophobic (Ho) acids in soil water extracts from organic and mineral soils along the Parks Highway stand productivity gradient.

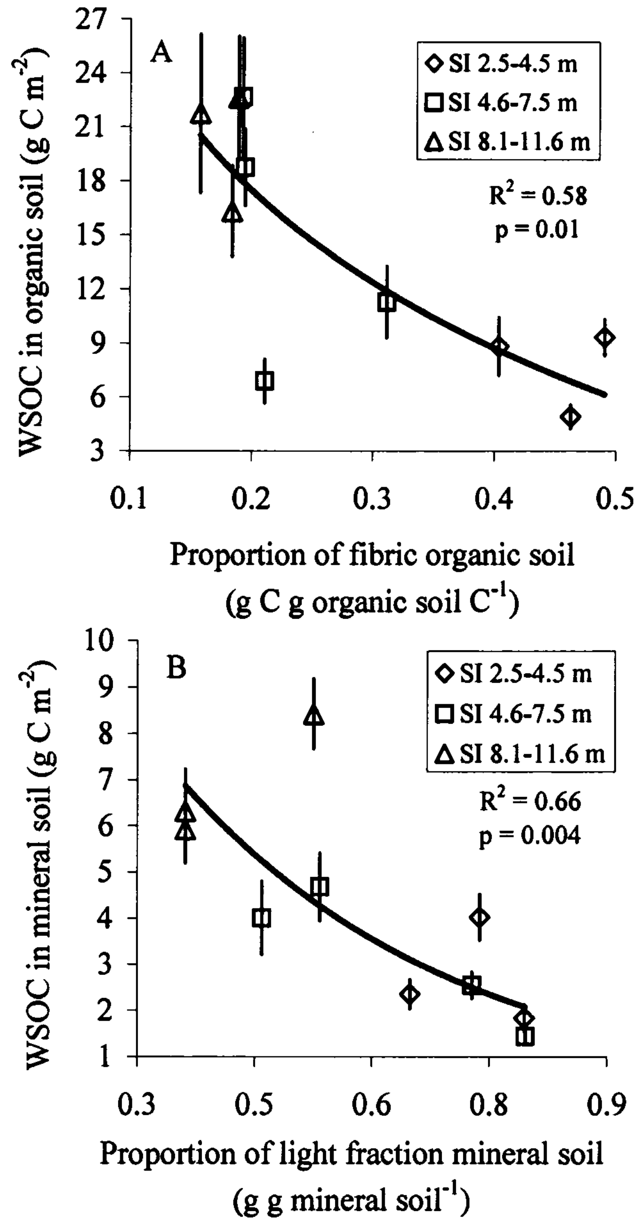


FIGURE 2.5. WSOC stocks (g C $m^{-2}$) vs. degree of SOM decomposition. (A) Fall concentrations of WSOC were higher at sites with a lower proportion of fibric SOC in the organic horizons. (B) Summer concentrations of WSOC were higher at sites with a lower proportion of light fraction (< 1.6 g $cm^{-3}$) SOM in the mineral (5 cm) horizons. Error bars are standard errors of the mean.

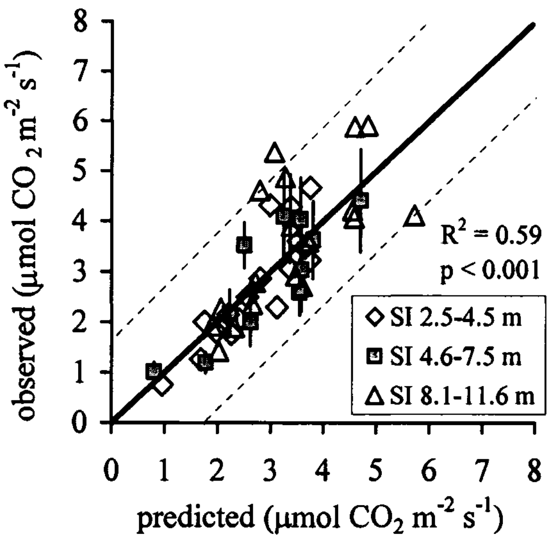


FIGURE 2.6. Predicted vs. observed soil respiration (µmol $CO_2$ $m^{-2}$ $s^{-1}$). Predicted soil respiration (µmol $CO_2$ $m^{-2}$ $s^{-1}$) vs. flux values observed throughout the growing season (May – September) in 2004. Observed values from the mid productivity sites (MM and HM) were plotted (squares) to validate predicted values, as these observations were not included in the empirical relationship depicted in Table 2. Dashed lines represent a 95% prediction interval about the values. Error bars are standard errors of the mean.

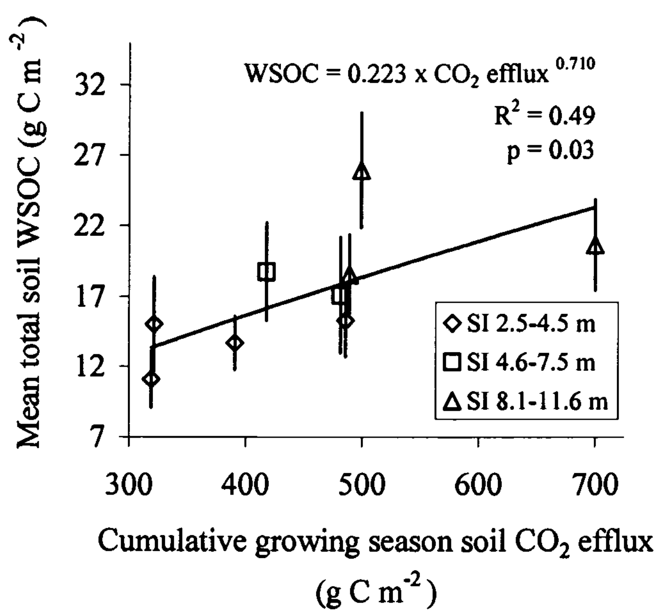


FIGURE 2.7. Mean total WSOC stocks increased with total soil $CO_2$ evolution. Sites are grouped by their relative level of stand productivity, with high (triangles), medium (squares), and low (diamonds) SI (m) values occurring in each study area. Error bars are standard errors of the mean.

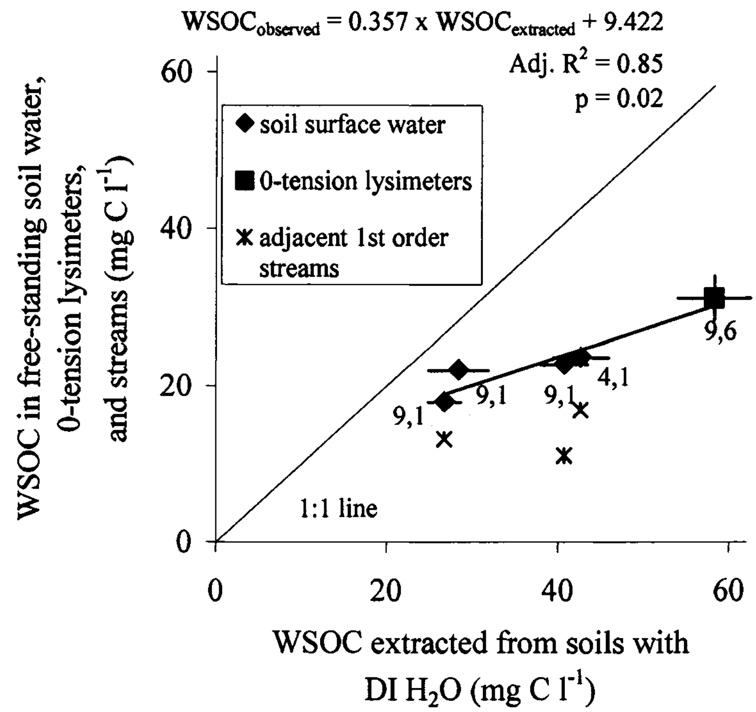


FIGURE 2.8. Observed WSOC concentrations in free-standing soil water and tensionless lysimeters are used to scale WSOC concentrations from soil core extracts. Inset values are sample n for X and Y, respectively. Error bars are standard errors of the mean.

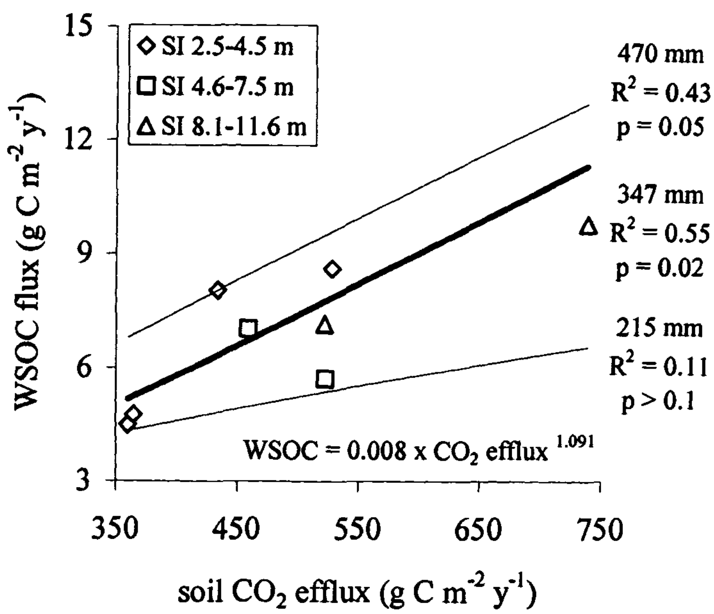


FIGURE 2.9. Annual flux estimates of WSOC vs. soil $CO_2$ efflux (g C $m^{-2}$ $y^{-1}$). Dashed lines are trends through annual WSOC fluxes estimated for a high (470 mm) and a low (215 mm) water year vs. soil respiration (points not shown). Sites are grouped by their relative level of stand productivity, with high (triangles), medium (squares), and low (diamonds) SI (m) values occurring in each study area.

# Chapter 3: Topographic influences on wildfire consumption of soil organic carbon in black spruce forests of interior Alaska: implications for black carbon accumulation

## 3.1: Abstract

Little is known of what primarily controls landscape patterns of soil organic matter consumption in wildfire, or what the legacy of fire is in stable soil carbon formation in boreal forests. We measured characteristics of soil organic carbon (SOC), residual organic matter, and black carbon (BC) in surface mineral soils along opposed north- and south-facing toposequences in recent (2004) and old (~1860-1950) burn sites throughout interior Alaska. North-facing and flat areas harbored more than twice the SOC in surface organic horizons (2.4 ± 0.5 and 3.0 ± 0.9 kg C $m^2$, respectively) than did south-facing forests (1.4 ± 0.4 kg C $m^2$) after wildfires. Mineral soil black carbon stocks ranged from 112 ± 24 g C $m^{-2}$ to 173 ± 43 g C $m^{-2}$ on north- and south-facing aspects, respectively. Black carbon comprised 6 ± 0.7% and 11 ± 2% of mineral horizon SOC stocks on north- and south-facing aspects, respectively. Across all sites, BC stocks and concentrations increased with decreasing soil moisture content, and drier south-facing forests harbored the most BC in surface mineral soils. These data supported the expectation that deeper, wet ground fuels occurring in north-facing and toe-slope forests burned less completely and produced less BC than on drier south-facing aspects. While warmer and drier forests harbored less total SOC, wildfire was a mechanism by which more stable C pools (BC) could aggrade even though the turnover rates of C in other soil pools also increased (as determined by $\Delta^{14}C$). These data provide a basis for

understanding how changing patterns of organic layer depths and soil moisture mediate the consumption of SOC in wildfire, as well as the long-term accumulation of BC, in the context of changing fire patterns likely to occur in a warmer climate.

**3.2: Introduction**

Over the past four decades, the arctic and boreal regions of the Northern Hemisphere have undergone a higher degree of warming than any other region on earth [Serreze et al., 2000; Chapin et al., 2005]. As a result, much research is focused on understanding how climate change is affecting the structure and functions of the region's ecosystems [Oechel et al., 1997; Kasischke and Stocks, 2000; Chapin et al., 2006]. Two climate-related phenomena are particularly important in boreal forests: fire and permafrost.

As a result of climate warming, the frequency of fire in the North American boreal forest region has more than doubled over the past four decades [Gillett et al., 2004; Kasischke and Turetsky, 2006], and the increasing temperatures forecasted by general circulation models are likely to result in further increases [Flannigan et al., 2005]. Fire is the most frequent disturbance in boreal forests, and serves as an important control over ecosystem processes through its effects on surface characteristics, which in turn, mediate solar radiation reflectance at the soil surface [Dyrness et al., 1986]. Variation in depth of burning of surface organic layers in boreal forests regulates soil temperature and moisture and therefore has consequences for post-fire soil C cycling [O'Neill et al., 2002; Bergner et al., 2004], tree recruitment [Landhaeusser and Wein, 1993; Johnstone and Kasischke, 2005; Johnstone and Chapin, 2006], permafrost maintenance [Yoshikawa et al., 2003;

Harden et al., 2006], and long-term soil carbon storage [Kasischke et al., 1995; Harden et al., 2000].

The formation of permanently frozen ground (or permafrost) is a unique characteristic of high latitude ecosystems which incorporates important feedback linkages with surface organic material remaining after wildfire. The presence of permafrost in boreal forests is strongly regulated by physiography (slope and aspect), soil drainage, and the presence of deep layers of bryophytes and organic soils that form over cold, wet soils because of low rates of decomposition [Viereck et al., 1983; Oechel and Van Cleve, 1986]. North-facing and toe-slope forests receive less insolation than do those on southerly slopes and therefore are cooler, wetter, and more likely to contain permafrost [Rieger, 1983; Slaugher and Viereck, 1986]. In addition, soil drainage in many sites occupied by black spruce depends on the seasonal thawing of permafrost, with drier soil conditions occurring later in the growing season when the depth to frozen soil is greatest. Since the degree of burning of surface organic layers in boreal forests regulates ground-temperature and moisture [Kasischke and Johnstone, 2005], understanding patterns of wildfire consumption of surficial organic layers is necessary in investigating the complex interplay between surficial organic matter accumulation and permafrost maintenance.

The amount of surface fuel consumption (SFC) within a particular burn event is determined largely by organic layer moisture and fuel type [Dyrness and Norum, 1983; Alexander et al., 1991; Miyanishi and Johnson, 2002; Kasischke and Johnstone, 2005; Harden et al., 2006]. Organic soils can contain more than 80% water by volume [e.g., Boelter and Verry, 1977] and therefore organic horizon depth influences the total moisture of organic matter that is exposed to combustion [Kasischke and Johnstone, 2005]. Wildfire extent and SFC in a given area likely vary with antecedent moisture conditions coincident with organic horizon thickness, but exactly how patterns of burning are affected by landscape biophysical properties are not well documented for boreal forests.

Physiography plays an important role in determining forest structure, which determines the fuel types available for combustion [Johnson, 1992; Ryan, 2002]. The distribution of fuels available for combustion in black spruce forests changes from predominantly bryophytes, surface duff, and shrubs on colder, north-facing sites where tree density is relatively low to crown-derived fuels on warmer sites where stand growth is not as limited by nutrient mineralization. The amount of organic matter burned is likely to be higher in warmer sites with a greater abundance of crown derived fuel types [Van Wagner, 1973; Alexander et al., 1984] and drier (shallower) organic layer fuels [Kasischke et al., 2000]. However, low-lying branches and abundant lichens on lower branches (fire ladders) are common traits of black spruce that make wildfire nearly inevitable wherever it dominates the stand, regardless of landscape position [Dyrness et al., 1986; Harden et al., 2001; Amiro et al., 2001]. Therefore, measuring changes in SFC, as well as residues from burning, in black spruce forests varying in slope position and aspect should constrain estimates of net SOC loss in wildfires within this boreal forest type.

Although the fraction of surficial soil organic carbon (SOC) released to the atmosphere in boreal forest fires can be considerable (between 16%-100% of SOC stocks; Kasischke et al. [1995], Harden et al. [2000]), little is known of the fate of fire-transformed C in soils. Pyrogenic C in soils exists along a continuum from black carbon (BC) in partly charred plant material to more graphitic BC particles, owing to microclimatic variation in burn temperature, burn duration, fuel type, and oxygen availability during a fire [Sekiguchi et al., 1983; Shindo, 1991; Kuhlbusch, 1995; Gleixner et al., 2001]. Although these factors make BC in soils hard to consistently identify [e.g., Goldberg, 1985; Schmidt and Noack, 2000], its composition of highly aromatized components affords operational definitions based on its resistance to thermally and chemically oxidizing conditions [Kuhlbusch, 1995; Schmidt et al., 2001; Preston and Schmidt, 2006]. Moreover, the highly-condensed molecular structure of BC suggests a high resistance to microbial degradation. Decay resistant BC therefore represents a sink for rapidly cycling C between the atmosphere and biosphere [Kuhlbusch and Crutzen, 1995] and is thought to comprise a significant amount of net biome production in fire-prone ecosystems [Schulze et al., 1999]. It follows that the contribution of BC to stable SOC pools should increase with successive fire cycles [Harden et al., 2000; Schulze et al., 2000], but since BC stocks accumulated over time are subject to combustion in a severe fire it is hard to ascertain the C legacy of past wildfires [Schmidt, 2004].

Here, we report on processes that regulate the depth of burning, as well as SOC and BC accumulation, in the soils of black spruce (*Picea mariana* Mill.) BSP) forests in interior Alaska. In this study we capitalize on the severe wildfires that occurred in the summer of 2004, which burned a record 2.7 x $10^6$ ha. Sites included opposed north- and south-facing toposequences in forests affected by recent (2004) and old (~1860-1950) fire events. Since duff moisture exerts major control over the severity and extent of burning within a given wildfire, and north-facing aspects are often colder and wetter than south-facing aspects in high-latitude systems, we hypothesized that SFC would be higher on southerly slopes compared to north-facing and toe-slope forests. We focused on BC accumulation at the organic/mineral soil interface (A horizons) to investigate long-term pyrogenic C accumulation across sites because organic horizon C stocks are vulnerable to burning. We expected to find greater BC accumulation in surface mineral soils from south facing aspects, as a byproduct of more biomass burned over time and shallower organic soils, than in soils from north-facing aspects.

## 3.3: Methods

### *3.4: Study sites*

Site establishment, temperature records, and soil profile and stand descriptions and analyses for 12 unburned stands (~60-160 y old) in this study were previously described in detail by Kane et al. [2005]. Briefly, four replicate gradients in black spruce productivity and soil temperature were established across interior Alaska, ranging in latitude from 63°-65°N (~365 km) and in longitude from 142°-148°W (~550 km) (Figure 1). Study areas were named for the roads used to access them and were established (from west to east) along the Parks Highway (P), Murphy Dome Rd. (M), the Elliott Highway (E), and the Taylor Highway (T), with low (L), mid (M) and high (H) levels of stand production occurring in each study region. The primary landscape attributes selected were aspect and local relief because they determine insolation, which in turn drives changes in soil temperature and moisture. Intermittent permafrost occurred at the PL, EM, EL, and TL sites (Typic Historthels and Aquic Haplorthels) and Inceptisols occurred at the other sites (Cryaquepts and Cryochrepts).

In addition to these 12 sites, we sampled 9 unburned stands located within or near the burn units studied on the Steese and Taylor Highways, and 4 unburned stands located in the Fairbanks region (Figure 1). We sampled 9 sites on south slopes, 11 on north slopes, and 5 on toe slopes (< 4%). General soil physical and climate data are presented for each landscape position in Table 1.

In June and July of 2005, black spruce forest stands that burned during the summer of 2004 were sampled. Our sites were located within two different fire events – the 217,000 ha Boundary fire (BY) that started on 13 June 2004 and the 115,000 ha Porcupine fire (PE) that started on 21 June 2004 (Figure 1). Test sites were located along opposed north- and south-facing toposequences, from uplands to lower toe-slopes and flat areas. Mean stand age (as determined in the laboratory by counting incremental growth rings on basal sections obtained from dominant trees in a given stand) of the BY fire sites was approximately 135 y (north = 134 y, south = 137 y, $n$=25), and of the PE fire sites was 90 y (north = 88 y, south = 95 y, $n$=34), prior to burning in 2004.

**3.5: Soil characteristics in unburned stands**

**3.6:** ***Soil profile descriptions and analyses***

Twelve soil pits (approximately 0.6 $m^2$, to a depth of 5 cm into the B horizon) were dug at each unburned site. Genetic horizons were delineated in the field into the fibric (Oi), mesic and humic (Oe+Oa), and the mineral soil A and B horizons (following

Schoenberger et al. [2002]; see Harden et al. [2004] for comparison between U.S. and Canadian classification). Macroscopic soil charcoal pieces were also harvested from A horizons in the field; all soil sampling methods and SOC determination were as previously reported by Kane et al. [2005].

Mineral soils from the A horizon were subsampled from 6 pits at each unburned site, chosen at random, to be separated by density into fractions < 1.6 g $cm^{-3}$ (light) and > 1.6 g $cm^{-3}$ (dense) using sodium polytungstate solution (Sometu Co., Sherman Oaks, California) as previously described by Baisden et al. [2002], following Golchin et al. [1994]. Elemental analysis (C, H, N) and loss on ignition (400 °C for 7 h) were determined on light fractions; percent oxygen was calculated by difference, following Haumaier and Zech [1995].

Stable isotope measurements were performed with a GEO 20-20 duel-inlet isotope ratio mass spectrometer coupled with a PDZ Europa solid preparation module. Results are expressed in standard $\delta^{13}C$ notation, as the deviation (‰) relative to the isotopic ratio of Pee Dee belemnite standard. The samples used for $\delta^{13}C$ analysis were the same as those used for total C analysis and in the density fractionations. Charcoal samples were washed in HCl for several hours and rinsed with DI prior to analysis [Bennett et al., 1990]. Samples were analyzed at least in duplicate such that replicates differed by <0.2‰ of the mean.

We analyzed the $\Delta^{14}C$ (‰) content of dense fraction (> 1.6 g $cm^{-3}$) soils to determine C mean residence times (turnover $rate^{-1}$) within this specific SOC pool [see Baisden et al., 2002]. Mean residence times were estimated through use of a homogenous one-pool model of $\Delta^{14}C$ incorporation, as previously reported by Kane et al. [2005], following the methods described in detail by Trumbore [2000].

**3.7: *Black carbon determination***

We measured the BC content of surface mineral soils (A horizons) obtained from 6 soil pits at each unburned site, as described in detail by Kuhlbusch [1995]. Briefly, this method entailed elemental analysis of soil C after two pretreatments: a solvent extraction followed by a thermal treatment. The solvent extraction entailed placing ~1.3 g of pulverized soil into a 50 mL centrifuge tube and treating it with different solvents (10-20 mL) in the order: 2x NaOH (1M), 1x 70 mass percent $HNO_3$, 5x 1M NaOH, 1x 1 mass percent HCl, and 2x de-ionized water. The sample was placed in an ultrasonic bath (10-20 minutes), centrifuged (3,000 RPM), and the supernatant was decanted, in between solvent treatments (11 solvent treatments per sample). The sample was then dried, pulverized again, and placed in a chamber (in a pure oxygen flow at 1,000 mL $min^{-1}$) inside an extra resistance wire tube furnace maintained at 340 ± 4 °C for 2 h. Darco 4x12 granular charcoal (Norit Americas Inc.; Atlanta, Georgia) was pulverized and used as a blank sample in the thermal treatments (mean loss of 0.03 g $g^{-1}$ by mass). The volume of the thermal treatment chamber was 2.6 L, enabling 12 samples in ceramic boats to be heated at a time. The C that was resistant to the solvent extractions and the thermal treatment, defined by Kuhlbusch [1995] as BC, was determined by the difference in C contents in each fraction using a LECO CNS analyzer (LECO Co., St. Joseph, Michigan).

In a comparative analysis of BC determination in soils, Schmidt et al. [2001] found that BC concentrations could vary by over two orders of magnitude for an individual sample, depending on the method-dependent operational definition of BC. The method used in this study [Kuhlbusch, 1995] yielded the third lowest amount of BC in the comparison of 6 methods investigated by Schmidt et al. [2001], and therefore should be considered a somewhat conservative measure of BC in soils. Australia's Commonwealth Scientific and Industrial Research Organization (CSIRO) estimated the BC content of reference materials, a German Mollisol and an Australian Vertisol, to be 4.7 and 10.4 g C $kg^{-1}$ respectively, using photo-oxidation and nuclear magnetic resonance (Skjemstad et al. [1999]; www.geo.unizh.ch/phys/bc/). Consistent with Schmidt et al. [2001], our analysis of these BC reference soils provided ~60% lower BC yields than the CSIRO's determinations ($1.63 \pm 0.15$ and $4.63 \pm 0.12$ g C $kg^{-1}$ for the Mollisol and the Vertisol, respectively). Therefore, estimates of soil BC yields reported herein are conservative in comparison with determinations by other methods.

**3.8: *Soil temperature and moisture***

We used end-of-growing-season mineral soil temperature and moisture along with measures of the depths of the organic layer profiles to confirm that flat, toe-slope and north-facing forests had cooler, wetter soils than south-facing stands, following Kasischke and Johnstone [2005]. In late September 2005, we dug three soil pits to permafrost or a depth of 30 cm at the 25 unburned sites. Soil temperature and moisture data were integrated over the top 30 cm (at 5 cm intervals) using a handheld digital thermometer and a Hydrosense volumetric water content meter (Campbell Scientific Inc., USA) (Table 1). Continuous soil temperature measurements within each genetic horizon at 12 stands were as previously reported by Kane et al. [2005]. Organic and mineral soil

(5 cm) gravimetric moisture contents were also determined (bi weekly) with soil cores obtained through the growing seasons of 2004-2005 at these 12 sites.

**3.9: Organic layer characteristics in burned stands**

At each site, we located a 40 m baseline transect in a homogeneous patch of burned forest. The baseline was oriented in a random direction, and bisected three 30 m transects, one located at the center, and two located at random distances in each direction from the center. Sample points were located every 5 m along each 30 m transect (7 per transect) and an additional 4 points were located along the baseline (at 5, 15, 25, and 35 m), for a total of 25 points per site. The depth of different organic layers (moss, lichen, residual char, fibric soil, mesic soil, and humic soil) was measured and samples from the different organic horizons at 10 to 21 points per plot were extracted to estimate bulk density and percent carbon [following Kasischke and Johnstone, 2005].

Kasischke and Johstone [2005] showed that adventitious roots on black spruce trees (roots formed above the base of the original root collar; LeBarron [1945]) in interior Alaska were located approximately 4 cm below the surface of the organic layer and therefore provide a baseline for pre-fire organic layer depth measurements. At the closest tree to each sample point, we measured the distance of the adventitious root above the burned organic layer (two measures were obtained, the minimum and maximum distance) and the distance from the top of the burned organic layer to the mineral soil. We used the average distance of the adventitious root above the mineral soil, plus 4 cm, as the total organic layer depth prior to burning (n = 25). Appendices containing residual soil depth data and SFC for each site in the BY and PE fire units can be found at: www.lter.uaf.edu/research_vignettes.cfm.

Soil organic C in burned stands was estimated using measured horizon depths of residual soil, and bulk density and C concentration data values measured for similar landscape positions as previously described. Measured depths from adventitious roots to residual soil were used to estimate SOC consumed in wildfire through use of an empirical relationship between organic horizon depth and SOC developed from published soil profile data in Alaskan black spruce forests (O'Neill [2000], King et al. [2002], Harden et al. [2004], Kane et al. [2005]; $r^2 = 0.71, p < 0.001, n = 18$).

**3.10: Data analysis**

Site means in soil properties were calculated from plot level means. Two soil samples were chosen at random from each plot for $\delta^{13}C$ isotope analyses ($n = 6$). Unburned cross-site comparisons involving annual soil temperature, gravimetric soil moisture, or BC have an $n = 10$ because two monitoring sites were burned in the PE unit fire of 2004.

Differences among sites in mean annual temperature and soil profile depths were tested with one-way analysis of variance pairwise contrasts at $\alpha = 0.05$ (Tukey error protection), or with analysis of covariance with landscape position as a covariate [Cody and Smith, 1997]. Pearson correlation coefficients demonstrated relationships between variables ($\alpha = 0.05$). Multiple regression analysis (forward selection procedure) developed empirical relationships between SOC and physiographic properties. Descriptive statistics were performed with Analyze-it statistical module (Analyze-it

Software, Ltd. Leeds, UK) and regressions and ANCOVA were developed through PC SAS (version 8.2, SAS institute, Cary, North Carolina).

### 3.11: Results

### 3.12: *Landscape controls on temperature and moisture*

North-facing aspects and toe-slopes (< 4% slope) exhibited cooler temperatures than southerly slopes both annually and seasonally (Table 1). Soil temperature measurements taken at the end of the growing season were highly correlated with mean annual soil temperatures ($r = 0.95$, $p = 0.004$, $n = 7$), supporting our assumption that end-of-growing-season soil temperature and moisture data can be used to analyze factors influencing these characteristics. Soil temperature decreased while soil moisture increased with increasing organic matter depth across all unburned sites (60-160 y old) in this study (Figure 2), which is consistent with previous observations from black spruce stands located in flat topography [Kasischke and Johnstone, 2005].

### 3.13: *Patterns of organic matter consumption in wildfire*

Total pre-fire organic horizon depths estimated from residual tree root collars (± standard error) were significantly higher on flat and north facing aspects (27.5 ± 1.5 and 24.5 ± 1.2 cm, respectively) than on southerly slopes (18.5 ± 1.1 cm; 1-way ANOVA pairwise comparisons with Tukey error protection, $F = 14.06$, $p < 0.001$; Figure 3). Mean post-fire organic horizon depths were significantly lower in south-facing forests (6.1 ± 1.3 cm) than in north-facing (11.5 ± 1.6 cm) and flat sites (16.6 ± 1.1 cm; $F = 13.14$, $p < 0.001$). The amount of residual deep duff (humic soil) varied with topographic position as well, with shallower depths occurring on southerly slopes (3.6 ± 0.6 cm) than on north-facing or flat areas (6.4 ± 0.7 and 8.7 ± 1.3 cm, respectively; $F = 10.07$, $p < 0.001$). Across all burn sites investigated, organic layer depths prior to burning explained 77% of the variance in post-fire organic layer depths ($p < 0.001$), with north facing and flat areas exhibiting the deepest pre- and post-fire organic layer depths (Figure 4). The relationship between pre- and post-fire organic layer depths did not change with landscape position (ANCOVA with three class variables; $F = 1.50$, $p = 0.24$), as there were no differences in depths of organic layer consumption among the different landscape positions investigated in these two early season fires (Figure 4).

Consistent with residual soil depth measurements, residual SOC was mostly contained within the moderately decomposed mesic horizon, and north-facing and flat areas harbored approximately 2.2 times the SOC in this horizon after wildfires than did south-facing forests (Figure 5). On average, an estimated 4.9 ± 0.3 kg SOC $m^{-2}$ was combusted on south-facing aspects and 4.0 ± 0.3 and 3.2 ± 0.3 kg SOC $m^{-2}$ were consumed on northern slopes and in flat areas, respectively. A higher fraction of total SOC was consumed on south-facing aspects than on both north-facing and toe-slope (< 4%) landscape positions in the Boundary and Porcupine unit fires of 2004 (Figure 5).

### 3.14: *Patterns of BC accumulation; $\delta^{13}C$, $\Delta^{14}C$, O:C*

Charcoal pieces found at the organic-mineral soil interface from mature forests in this study were more enriched in $\delta^{13}C$ (mean ± SE: -25.4‰ ± 0.2; $n$=10) than both light (<1.6 g $cm^{-3}$; -25.9‰ ± 0.1; $n$=12) and dense fraction (>1.6 g $cm^{-3}$; -26.4‰ ± 0.1; $n$=12) A horizon organic matter. Light fraction organic matter was more depleted than charcoal by 0.5‰, whereas the dense fraction was more depleted than charcoal by 1.0‰. Dense fraction C concentration in the A horizon increased with increasing $\Delta^{14}C$ enrichment across all sites ($r^2 = 0.55$, $p = 0.03$), with warmer and drier south-facing sites exhibiting the most $\Delta^{14}C$ (‰) enrichment. Moreover, the light fraction atomic O:C ratio was lower in soils from south-facing aspects (0.48 ± 0.01) than from north-facing aspects (0.50 ± 0.01; ANOVA pairwise, $p < 0.05$), and therefore more closely reflected the O:C ratio of soil charcoal (0.30 ± 0.02; Figure 6). The H:C ratio of soil organic matter did not vary significantly between aspects (Figure 6).

The acid insoluble fraction of carbon in surface mineral soils, which constituted the first pre-treatment in BC determination [Kuhlbusch, 1995], was significantly higher in the colder, north-facing sites than in south-facing forests, but differences in mean BC concentration were not significant between aspects (Table 2). Carbon concentrations in A horizon soils were higher in north facing sites. The proportion of BC to SOC changed significantly between aspects, with south facing forest soils having a BC:OC ratio approximately 1.7 times higher than was measured in north-facing forests (Table 2). Black carbon comprised roughly 11% and 6% of total A horizon SOC on south- and north-facing aspects, respectively. Surface mineral soil BC stocks and BC:OC both decreased with increasing soil moisture content measured at the end of the growing season across all sites (with one poorly-drained permafrost site (PL) removed as an outlier; Figure 7). Additionally, A horizon mean C residence times were higher in the cooler, wetter forests, and BC stocks and concentrations decreased as the mean residence time of dense fraction C increased across sites (Figure 8).

## 3.15: Discussion

### 3.16: *Patterns of organic matter consumption*

In the two burn units surveyed in this study, north-facing and toe-slope forests retained significantly more SOC following wildfire than was found on southerly slopes, owing to higher antecedent organic matter depths occurring in this landscape position (Figures 2 and 3). These results illustrate the feedback linkages between depth of the soil organic layer at the time of the burn and soil moisture in permafrost sites (Figure 2), as well as the feedback between soil moisture and the resultant impact on depth of burning during fires.

The organic layer depth/organic layer moisture feedback was important in mitigating the proportion of total organic layer depth consumed at the time when extreme fire weather and behavior occurred throughout east-central Alaska, which produced extended periods of low drought codes for each fire unit studied. The Fire Weather Index (FWI), which was developed by the Canadian Forest Service to integrate the effects of fuel moisture, drought, and wind on fire intensity and rate of spread, increased from averages of 7.6-7.9 throughout May to averages of 23.1-22.3 throughout June in the areas affected by the BY and PE fires, respectively (according to the Alaska Fire Service, Fire Weather Index Database). Additionally, the number of days with a FWI value >10 increased steadily throughout the fire season across both regions studied (10 d in May, 22 d in June, and 28 d in August). It is therefore likely that fires occurring later in the season, with increased drought and recession of permafrost, would burn deeper into the organic layers in north-facing and flat areas than in south facing areas with shallower depths to mineral soil, because mineral soil would limit the depth of consumption. Even in the two early season wildfires studied, consumption of nearly all soil organic horizons was common on south-facing slopes, which had the shallowest and driest organic layers, leaving only char on top of the mineral soil. These results suggest that while the feedback between increased organic layer depth and organic layer moisture is an important control over residual organic layer depths following wildfire in black spruce forests, increased drought or the recession of permafrost could allow for deeper SFC.

**3.17: Patterns of char deposition**

While it is possible that colder, wetter boreal forests harbor more char in mesic horizons owing to decreased depth of burning, over time charcoal can be transported from organic to surface mineral soil horizons as organic matter and vegetation aggrade [Carcaillet and Talon, 2001; Gavin, 2003; Ping et al., 2005]. Therefore, char is often concentrated at the organic/mineral soil interface in boreal forests (Alaska: Ping et al. [2005]; Canada: Harden et al. [2000], Preston et al. [2002]; Siberia: Schulze et al. [1999]; and across the loess-belt of Germany: Schmidt et al. [1999]).

Increasing decomposition extent of soil organic matter (SOM) has been linked to $\delta^{13}C$ enrichment [Nadelhoffer and Fry, 1988; Boutton, 1996; Amundson and Baisden, 2000] and increasing density of SOM fractions (e.g., > 1.6 g $m^{-2}$; Baisden et al. [2002], McLauchlan and Hobbie [2004]), and deviation from these trends may be due to the incorporation of charred remains [Baisden et al., 2002]. Char material is believed to have a density < 1.6 g $m^{-2}$ [Roscoe et al., 2001; Baisden et al., 2002], but soil charcoal is enriched in $^{13}C$ relative to matricial litter in field studies (King et al. [2002], Preston et al.

[2006]; ~2.0‰). Soil charcoal enrichment, as observed in this study, is probably due to changes in the isotopic composition of the fuel source because fractionation due to burning is minor [Leavitt et al., 1982; Turney et al., 2006]. Bole wood, for example, is isotopically enriched relative to fibric and mesic soil and constitutes the majority of aboveground fuel types (~43% of total biomass) available for combustion in black spruce forests [Schuur et al., 2003]. The $\delta^{13}C$ of light fraction (< 1.6 g $cm^{-3}$) SOM was likely influenced by soil charcoal incorporation because it was isotopically enriched relative to dense fractions, despite being composed of OM that has undergone less decomposition than dense fraction SOM. These data support the expectation that patterns of accumulated char materials in soil are reflected in light fraction SOM.

Analysis of the atomic H:C and O:C ratios of soil fractions can provide insight into the genesis of pyrogenic materials [Kuhlbusch and Crutzen, 1995; Preston and Schmidt, 2006] because heated SOM, as in a wildfire, has a higher degree of aromaticity and undergoes decarboxylation [Almendros et al., 1992]. The O:C ratio of char material declines with an increase in its temperature of formation [Sekiguchi et al., 1983] and also with an increase in the duration of heating [Shindo, 1991]. While it is true that soil humic acids also have low O:C ratios and could therefore confuse an interpretation of pyrogenic material in soil [Haumaier and Zech, 1995], we have previously observed lower humic acid concentrations in a subset of the warmer, south-facing stands than in cooler forests with deeper organic matter [Kane et al., 2006]. Therefore, the lower O:C in light-fraction mineral soils from drier, south-facing forests (Figure 6) probably reflects burn residues from hotter and/or longer duration wildfires, or a higher concentration of pyrogenic C, than occurred in the cooler and wetter stands.

### 3.18: ***Patterns of Black carbon accumulation***

In a recent synthesis of pyrogenic C in boreal forest soils, Preston and Schmidt [2006] ceded that BC estimates are very limited and difficult to compare due to variations in sampling and analysis, but offered that BC could be on the order of 100-200 g $m^{-2}$ in the forest floor. Moreover, reports of BC concentrations within the soil profile vary widely, even within a stand type. Concentrations of BC in a Siberian pine stand were reported to be very low in the litter and surface organic horizons, and were highest in the surface mineral soil (EB horizon; Schulze et al. [1999]). In contrast, Czimczik et al. [2005] found that most (up to 99%) of the BC was within organic horizons, and yields were an order of magnitude lower than those reported for other Siberian pine forests. While the range of BC stocks reported herein for surface mineral soils in interior Alaska (Table 2) mirrors the range offered by Preston and Schmidt [2006], our values underestimate total site BC by the potential amount harbored in organic horizons.

Given the high frequency of stand-replacing fires in spruce forests of interior Alaska (~100 years; Yarie [1981]), the high potential for charcoal production from biomass burning (72.9-93.2 g $m^{-2}$; Clark et al. [1998], Lynch et al. [2004]), and the persistence of char in soil (~200-1000 y; Harden et al. [2000]), one would expect far more char at the mineral soil surface than we report. Similarly, Czimczik et al. [2003] concluded that some form of *in situ* char degradation must be occurring in fire-prone Siberian forests, with char consumption in subsequent fires offering the most parsimonious explanation for relatively small BC stocks [Ohlson and Tryterud, 2000; Czimczik et al., 2005]. One would expect lower proportions of BC consumption in subsequent fires to occur in wetter forests with deeper OM depths, but we observed a negative trend between BC stocks and increasing soil moisture in this study. Black carbon incorporated with surface mineral soil, which is more likely to occur in drier south-facing forests with shallower organic horizons [Ping et al., 2005], is more protected from burning in subsequent wildfires.

Organic matter depths and soil moisture, which constrain the combustion process itself by influencing the temperature of combustion, $O_2$ concentration, and mode of combustion (e.g., flaming or smoldering), changed with landscape position in this study. Black carbon production occurs in smoldering fires in the absence of $O_2$, after fuels have undergone sufficient drying via distillation [Gleixner et al., 2001]. Declining BC stocks and concentrations with increasing soil moisture observed in this study (Figure 7) supports the expectation that deep, wet ground fuels occurring in north-facing and toe-slope forests burn shallower and produce less BC than do drier south-facing aspects. However, it is also possible that BC declines with increasing moisture to a point, after which increasing soil moisture protects BC stocks at the organic/mineral soil interface from burning in subsequent fires. This could explain why the wettest site in this study (PL) had the second largest BC stock in the surface mineral soil (Figure 7). Further research tracking the fate of BC in wetter endpoints along moisture gradients (i.e., treed bogs and fens) would elucidate the role soil moisture plays in organic matter consumption, BC production, and the protection of BC from oxidation in subsequent fires.

The physical characteristics constraining soil BC formation and accumulation are different than the driving factors of non-pyrogenic soil organic matter accumulation. While warmer, drier conditions are more amenable to combustion and BC production, soil organic matter turnover processes also increased in the surface soils of warmer, south-facing stands [Kane et al., 2005]. BC stocks and concentrations were higher in surface mineral soils of warmer, drier forests in spite of increased C turnover rates there (Figure 8). This hints at the recalcitrant nature of BC and suggests that there is preferential decomposition of other organic matter fractions in warmer, drier forests. These data suggest that while warmer, drier forests harbored less total SOC, wildfire was a mechanism by which more stable C pools (BC) could aggrade even though the turnover rates of C in other soil pools also increased. Notwithstanding, BC stocks were miniscule in comparison to SOC harbored in organic horizons (4% and 2% of organic horizon SOC on south- and north-facing aspects, respectively), and therefore the C stabilizing effect of wildfire is small in comparison to SOC accumulation through arrested decomposition, which predominates in cooler and wetter forests.

**3.19: Conclusions**

Patterns of organic matter depth, soil temperature, and soil moisture, all of which changed with physiography, likely determined the proportion of organic matter consumed from different landscape positions within two unique wildfires in Alaska. Warmer and drier south-facing forests lost a greater proportion of their total SOC stocks than did north-facing and toe-slope forests in wild fires occurring in 2004, and black carbon stocks measured in surface mineral soils from unburned forests (60 – 160 y old) suggested that this trend applied to past fire years as well. Black carbon content in surface mineral soils in unburned stands decreased with increasing soil moisture, supporting the expectation that less of the deeper, wetter surface soil fuels burned or were converted to BC. However, complete saturation of surface soils (as in permafrost dominated sites with perched water tables) could preserve BC stocks from oxidation in subsequent fires, meaning an increase in BC accumulation with increased moisture. While warmer, drier forests harbored less total SOC, wildfire was a mechanism by which more stable C pools (BC) could aggrade, even though the turnover rates of C in other soil pools also increased. Future work should track the fate of burn residues in boreal wetland soils, and also BC deeper in the soil profile and in organic horizons. These data are necessary in order to better understand the interaction between organic matter depth, soil moisture, and SFC in mediating the lasting effects of fire on SOC stocks in the context of changing fire patterns likely to occur in a warmer climate.

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Table 3.1. General soil physical and chemical properties of unburned stands. Means are presented by landscape position.

| Aspect | Mean Stand Age (years) | Organic matter (SE)[a] Depth (cm) | Bulk Density ($g\ cm^{-3}$) | Total C ($kg\ C\ m^{-2}$) | GDD[c] (>0 °C) | September means (SE)[b] $n$ | OM depth (cm) | Temperature (30 cm; °C) | Moisture (30 cm; $g\ g^{-1}$) |
|---|---|---|---|---|---|---|---|---|---|
| South | 95 | 13.3 (0.9) | 0.09 (0.01) | 4.43 (0.44) | 839 | 9 | 15 (1.2) | 3.6 (0.7) | 0.32 (0.05) |
| North | 108 | 20.1 (1.3) | 0.07 (0.01) | 5.74 (0.73) | 530 | 11 | 22.2 (1.4) | 1.8 (0.7) | 0.48 (0.06) |
| Flat/Toe-slope | 95 | 15.7 (1.5) | 0.07 (0.01) | 4.1 (0.46) | 379 | 5 | 20.5 (1.3) | 1.1 (0.2) | 0.73 (0.04) |

[a] Mean organic matter characteristics for all organic horizons (not including green moss); data from Kane et al. (2005).

[b] Mean values integrated from the top 30 cm of mineral soil from pits dug in September, 2005.

[c] Growing Degree Day heat sum (0 °C) from continuous measurements in the middle of the mesic horizon.

Table 3.2. Long-term accumulation of Black Carbon in surface mineral soil (A horizons).

| South aspect[a] | Initial BD (kg m$^{-3}$) | Total OC (g C kg$^{-1}$ soil) | Acid Insoluble[b] (g C kg$^{-1}$ soil) | Black Carbon[c] (g C kg$^{-1}$ soil) | BC:Total OC (g kg$^{-1}$) | BC Stocks (g m$^{-2}$) |
|---|---|---|---|---|---|---|
| MM | 441 (19) | 165.2 (18.7) | 97.6 (5.6) | 9.2 (1.2) | 53.3 (5.0) | 68.9 (5.3) |
| MH | 412 (12) | 140.4 (5.8) | 74.6 (5.1) | 12.7 (3.6) | 90.3 (22.4) | 141.2 (14.8) |
| PH | 394 (7) | 178.9 (11.4) | 89.3 (8.1) | 24.7 (4.7) | 143.7 (29.4) | 223.8 (4.0) |
| TH | 473 (8) | 128.8 (4.9) | 62.7 (7.3) | 17.7 (4.4) | 143.3 (24.9) | 259.4 (4.4) |
| Average[d] | 430 (17) | 153.3* (11.42) | 81.0* (7.7) | 16.1 (3.4) | 107.7* (22.0) | 173.3 (42.7) |
| North & no aspect | | | | | | |
| EL | 478 (27) | 196.5 (16.4) | 105.0 (16.5) | 7.9 (1.3) | 38.9 (3.7) | 45.3 (6.2) |
| EH | 384 (12) | 201.8 (19.8) | 81.2 (11.9) | 8.0 (1.1) | 54.7 (14.8) | 40.0 (4.2) |
| EM | 285 (39) | 211.0 (16.2) | 169.0 (22.4) | 18.0 (2.9) | 57.4 (4.9) | 71.8 (8.3) |
| ML | 442 (6) | 198.6 (16.2) | 122.5 (21.3) | 8.2 (1.0) | 39.7 (3.0) | 86.9 (4.4) |
| PM | 466 (3) | 188.1 (14.8) | 84.3 (13.7) | 16.3 (1.7) | 94.3 (6.8) | 152.0 (7.9) |
| PL | 467 (31) | 244.8 (9.9) | 144.0 (19.8) | 20.6 (3.1) | 84.6 (10.3) | 221.2 (14.5) |
| TL | 297 (21) | 210.6 (5.3) | 125.7 (13.8) | 10.6 (1.0) | 49.5 (4.5) | 81.9 (3.0) |
| TM | 305 (17) | 226.3 (14.1) | 109.0 (5.5) | 17.8 (1.6) | 77.2 (9.8) | 195.6 (4.9) |
| Average | 391 (30) | 209.7* (6.5) | 117.6* (10.4) | 13.4 (1.9) | 62.0* (7.4) | 111.8 (24.4) |

Sample $n$ = 12 for Bulk Density determination and $n$ = 6 for subsequent analyses. Standard Errors of the mean are in parentheses.

[a] Site labels describe their location on the Elliott Hwy. (E), Taylor Hwy. (T), Parks Hwy. (P), and Murphy Dome (M).

[b] Fraction of OC remaining after sequential acid ($NO_3$) and base (NaOH) extraction, following Kuhlbusch (1995).

[c] Additional non-hydrolyzable OC remaining after combustion at 340 °C in pure $O_2$ for 2 h.

[d] Arithmetic mean; asterisks denote significant differences in mean values between aspects ($p$ <0.05; 1-way ANOVA).

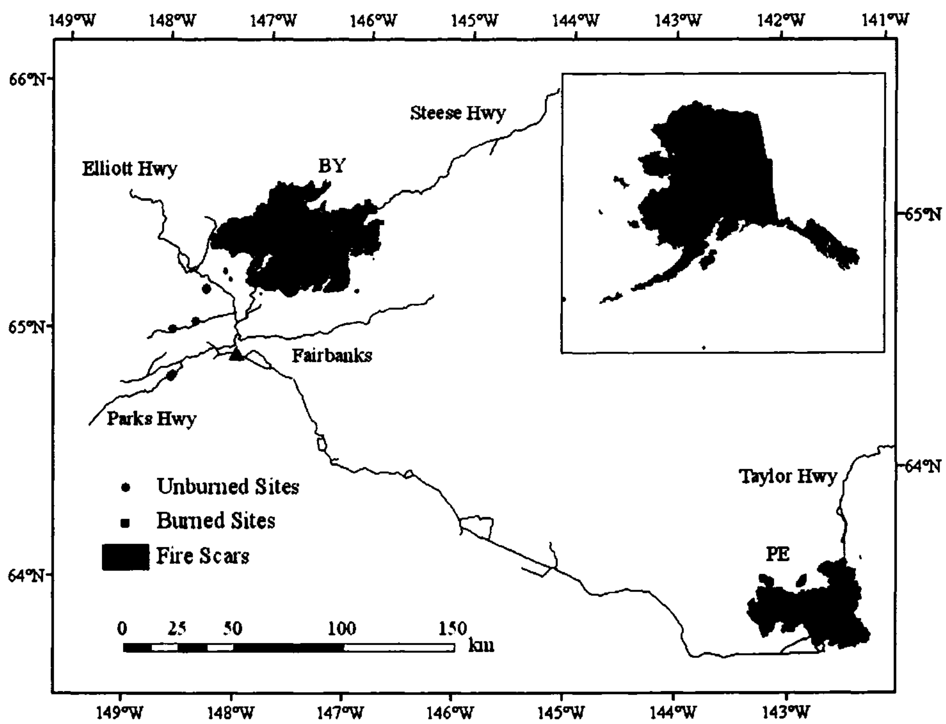


Figure 3.1. Distribution of unburned (~1860-1950) and burned (2004) sites. Burned sites studied were located in the Porcupine unit (PE; $n$=15) and Boundary unit (BY; $n$=16) fire scars. Inset figure shows the relative position of the study region in the state of Alaska, USA.

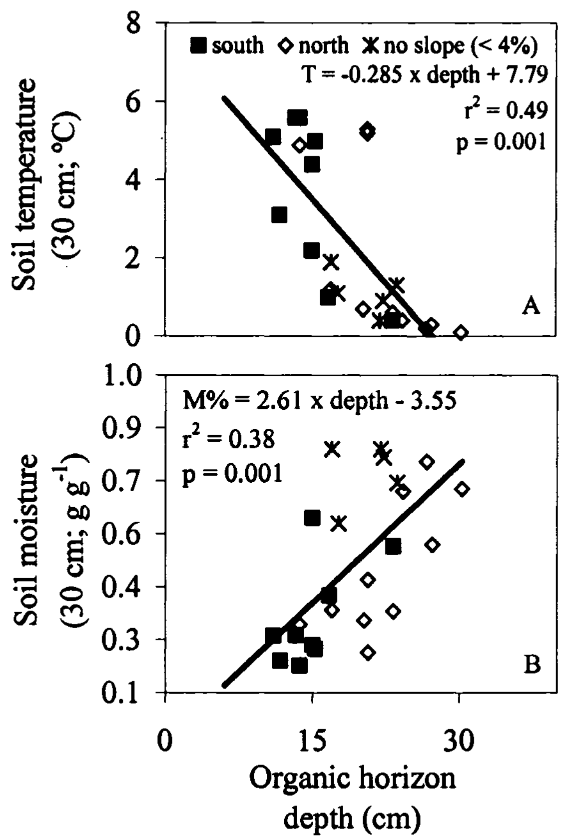


Figure 3.2. Organic layer depth vs. soil temperature (A) and soil moisture (B). Measurements were averaged from the upper mineral soil profile (30 cm) and incorporated locations with the most and least organic matter accumulation at each plot (n=3). All measurements were obtained from unburned black spruce stands (60-160 years old) in September.

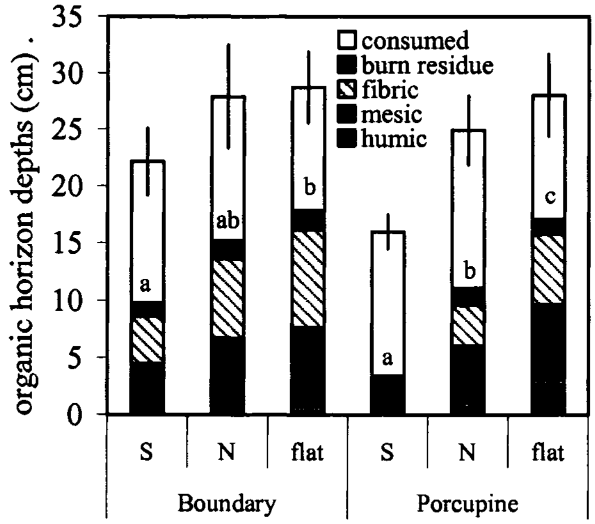


Figure 3.3. Residual organic horizon depths in stands from the fires of 2004. Residual organic horizon depths measured along opposed north and south facing toposequences in the Boundary and Porcupine Unit fires of 2004. Total organic matter depths (consumed) were reconstructed from the root collars of residual trees. Toposequences were measured along north (N) and south (S) and flat (< 4% slope) areas. Error bars represent combined standard errors of the mean values for residual and consumed organic matter depths. Letters denote significant differences in post-fire organic matter within a burn unit (ANOVA pairwise comparisons; $\alpha = 0.05$).

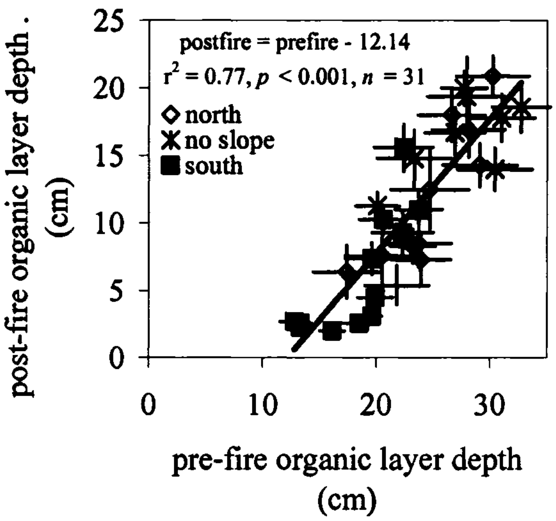


Figure 3.4. Post-fire vs. pre-fire organic horizon depths. The depth of organic horizons remaining after a fire increases with increasing depth of organic horizons prior to burning, as estimated from residual tree root collars. The relationship between pre- and post-fire organic matter depths did not vary with landscape position (ANCOVA; F = 1.50, p = 0.24). Error bars are standard errors of the mean.

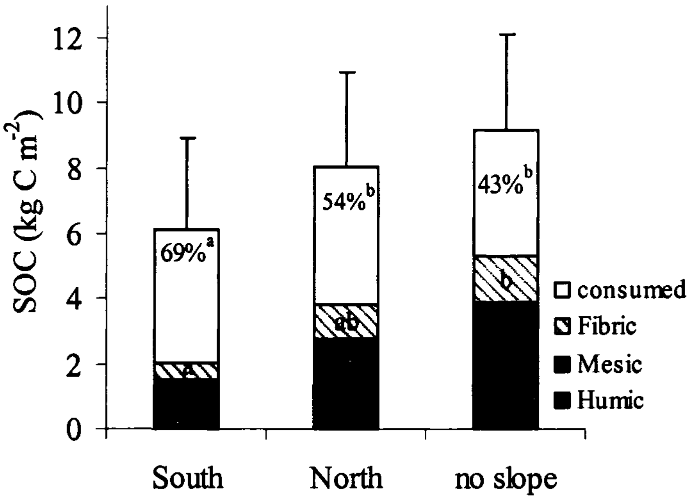


Figure 3.5. Residual soil organic carbon stocks. Soil organic carbon stocks from the Boundary and Porcupine unit fires are averaged for a given landscape position. Inset values are percent of total pre-fire SOC consumed in fire. Pre-fire organic matter depths were estimated from residual tree root collars. An empirical relationship between OM depth and SOC was used to determine C consumed in fire and also to estimate error associated with depth measurements. Letters denote significant differences in C stocks across landscape positions for a given horizon (1-way ANOVA pairwise comparisons; $\alpha = 0.05$). Error bars represent combined standard errors of the mean values for residual and consumed SOC.

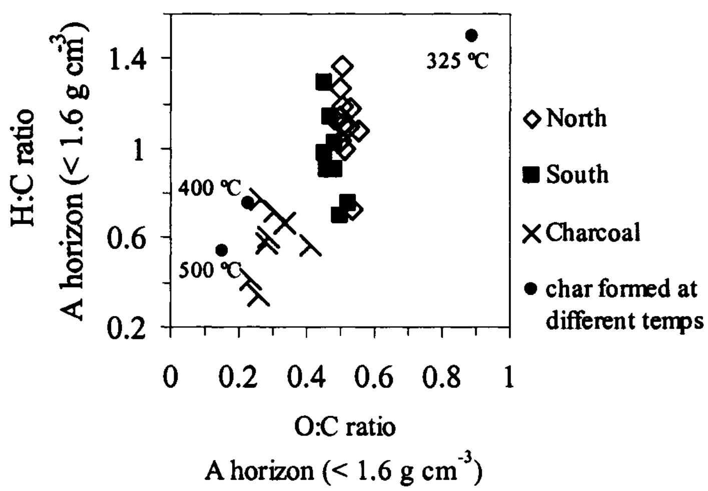


Figure 3.6. Atomic H:C vs. atomic O:C diagram of light fraction (< 1.6 g cm$^{-3}$) A horizon mineral soils sampled on north and south facing aspects (burned 60-160 years ago). Values for macroscopic soil charcoal (X's) obtained from the organic/mineral soil interface at each sites are also plotted. Circles (•) are values for cellulose and cellulose chars obtained by heating at several temperatures (data from Sekiguchi et al. [1983], plotted for comparison).

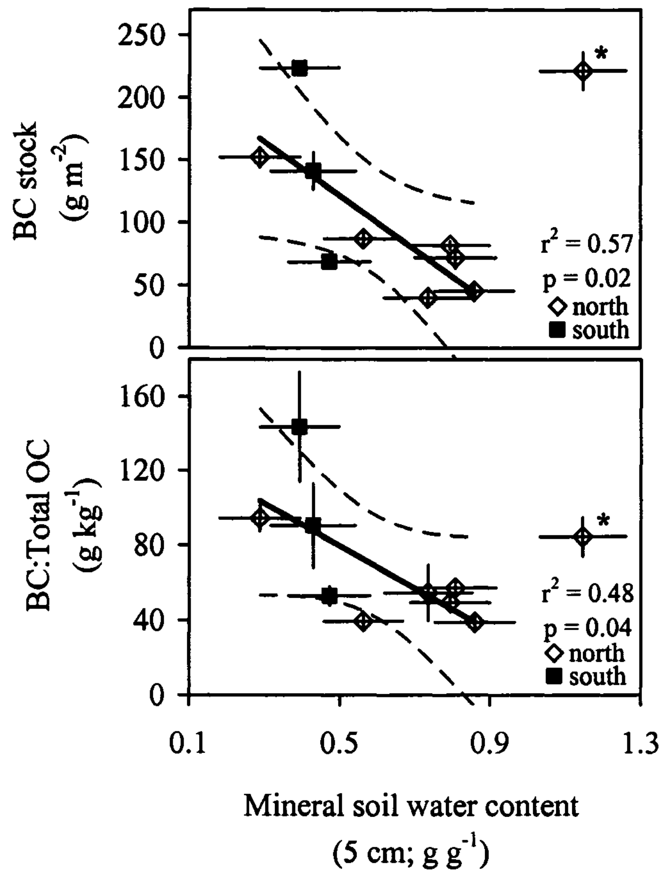


Figure 3.7. BC stocks and the ratio of BC to soil OC vs. soil moisture. Surface mineral soil BC stocks (top) and the ratio of BC to soil OC (bottom) both decrease with increasing mineral soil moisture measured ($n$=6) in September at each site. Dashed lines are a 95% confidence interval about the values and error bars represent standard errors of the mean. Asterisk denotes an outlier not included in the linear regression (PL site).

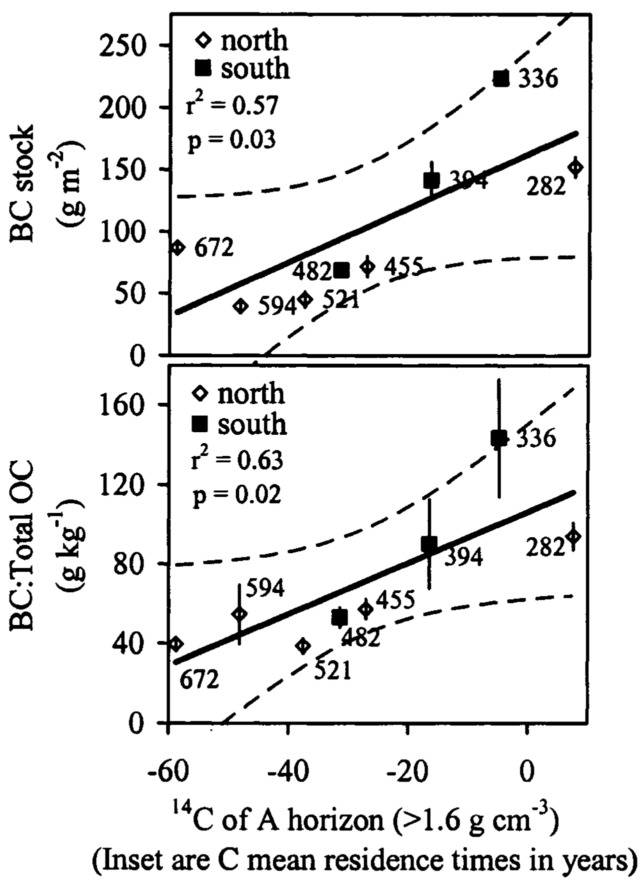


Figure 3.8. BC stocks and the ratio of BC to soil OC vs. $\Delta^{14}C$ (‰). Surface mineral soil black carbon stocks (top) and the ratio of BC to soil OC (bottom) both increase with the enrichment of $\Delta^{14}C$ (‰) measured in dense fraction (>1.6 g $cm^{-3}$) A horizon soils. Inset C mean residence times (years) were estimated based on the degree of $\Delta^{14}C$ incorporation. Dashed lines are a 95% confidence interval about the values and error bars represent standard errors of the mean.

# Chapter 4: Integration and concluding remarks

## 4.1: Integration and conclusion

In chapter 1 we demonstrated that near surface soil organic carbon (SOC) decreased with increased stand productivity (and hence, probable increases in C mineralization) across four replicated gradients in black spruce stand production and climate in interior Alaska. The distributions of pools within the mineral soil (a decrease in total SOC and proportion of light-fraction SOM) and within the organic soil (a decrease in Oi horizon depth and SOC) were apparently driven by interactions between increased soil temperature and stand productivity across all sites. Moreover, variation in total SOC across all sites was best explained by the difference in aboveground and belowground heat sums, which we suggest is largely mediated by the insulative properties of surface organic horizons. If organic horizon depths decrease in a warmer future climate, barring other limitations to mineralization, the thermal buffering capacity of the soil will also decline, leading to probable changes in the distribution and quantity of SOC pools. While total SOC pools declined with enhanced aboveground production, the changes in net C balance between above and belowground pools was not directly explored.

In agreement with these findings, in chapter 2 we demonstrated that total soil C efflux ($CO_2$ and water-soluble organic C, or WSOC) increased with soil temperature and stand productivity. Since these warmer and more productive stands also had shallower organic horizons, they also had a lower capacity to store water above the mineral soil, which was frozen during the period of peak water flush through the soil early in the spring. Water movement through organic horizons was higher in warmer and more productive forests, and peak outflows occurred early in the spring with snowmelt. Therefore, more SOC was leached during the spring runoff from warmer, more productive forests. Moreover, the more degraded products of microbial activity (fulvic acids) were higher in warmer, more productive stands whereas relatively labile C fractions (hydrophilic OM and low-molecular weigh fulvic acids) were highest in a cold stand with low productivity. These data suggest that both mineralization and lateral transfer of WSOC may be mechanisms by which more C could be lost from soils in warmer and more productive forests, especially if the magnitude of snowmelt flush increases (owing to increased winter precipitation).

While the first two chapters show significant changes in the distributions and fluxes of soil C with soil temperatures and stand characteristics, total C in above + belowground components (Table 1) did not significantly vary with soil temperatures (GDD >0°C, Oe horizons) across the 12 sites in this study ($p = 0.58$). Total belowground C pools (roots and soil) decreased as aboveground C (moss, forbs, and trees) increased (belowground C = -1832 x ln (aboveground C) + 23562; $r^2 = 0.35$, $p = 0.04$; see Table 1). Belowground C pools were approximately 10x and 4x greater than aboveground C pools in cold, low productivity stands and warmer, high productivity stands, respectively (Figure 1). Therefore, the distribution of C in belowground pools relative to above ground pools decreased with increasing temperatures across all sites (Figure 1), even though total C did not change across the climate and stand productivity gradients studied.

A redistribution of ecosystem C from belowground to aboveground pools with increasing temperature has important implications for the vulnerability and susceptibility of C loss through disturbances, such as wildfire. In chapter 3, we demonstrated that warmer and drier (south facing) forests lost a greater proportion of total SOC in wildfire than did their cooler, north facing counterparts within a burn unit. This was determined largely by antecedent organic horizon depths. Cooler and wetter north-facing forests had deeper pre-fire organic horizons, and therefore had greater post-fire residual organic matter depths than did drier, southerly slopes.

Stable isotope analysis revealed that most mineral soil pyrogenic C accumulated in the light-fraction ($< 1.6$ g $cm^{-3}$). Analysis of atomic O:C ratios of the light fraction mineral soil revealed that drier (south facing) forests harbored more pyrogenic C than did north-facing stands. Moreover, black carbon (BC, as defined by Kuhlbusch 1995) stocks and concentrations decreased with increased soil moisture content, with the driest stands with the shallowest organic horizons harboring the most BC. Increasing BC stocks accumulated at the organic/mineral soil interface with decreasing soil moisture supports the expectation that more biomass burned over time in the warmer and more productive south-facing forests.

Beginning in the mid-1980s, the frequency of larger fires years (those which affect >1% of the land surface) doubled in the Alaskan boreal forest region, resulting in a doubling of the annual burned area compared to the period of 1950 to 1984 (Kasischke and Turretsky 2006). A continuation of this trend coupled with the continued warming and drying predicted to occur (e.g., ACIA 2004) are likely to alter the distribution of C pools within black spruce forests of interior Alaska. The findings herein described suggest that, with climatic warming, soil C pools are likely to decline. While this decline in SOC probably does not constitute a net loss of C from the system due to co-occurring increases in aboveground production, the shift in C distribution from below- to above-ground pools as temperatures increase also increases the vulnerability of C to burning in wildfire.

**4.2: References**

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Kasischke ES, Turetsky MR. 2006. Recent changes in the fire regime across the North American boreal region - Spatial and temporal patterns of burning across Canada and Alaska. Geophys. Res. Lett. 33, doi:10.1029/2006Gl025677.

Kuhlbusch, TAJ. 1995. Method for determining black carbon in residues of vegetation fires. Environ. Sci. Technol., 29, 2695-2702.

Yarie J, Kane ES, Mack M. 2006. Biomass Equation for the Tree Species Present in interior Alaska. SNRAS misc. publ.

Table 4.1: Carbon pool sizes located belowground and aboveground. Soil and root stocks (B) and living phytomass (A) were measured along the four productivity gradients studied in the previous chapters.

| | Belowground (B) | | | | Aboveground (A) | | | | | | B : A |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Site | Total SOC[1] ($g\ m^{-2}$) | ± SE | Oi & Oe roots (>2 mm) ($g\ C\ m^{-2}$) | ± SE | Live moss ($g\ C\ m^{-2}$) | ± SE | Live forbs,grass ($g\ C\ m^{-2}$) | ± SE | Stand biomass[2] ($g\ C\ m^{-2}$) | ± SE | C pools ($g\ g^{-1}$) |
| EH | 9580.8 | 2113.4 | 514.1 | 152.8 | 166.1 | 17.2 | 32.4 | 4.9 | 2121.2 | 203.2 | 4.4 |
| EL | 8944.2 | 1296.6 | 264.0 | 89.2 | 285.5 | 27.0 | 59.5 | 15.0 | 2290.8 | 425.4 | 3.5 |
| EM | 10543.9 | 1505.0 | 254.2 | 46.6 | 212.8 | 21.1 | 22.5 | 0.8 | 1917.6 | 344.4 | 5.0 |
| MH | 7634.0 | 910.1 | 175.8 | 27.9 | 289.4 | 61.7 | 9.7 | 2.4 | 5491.0 | 639.7 | 1.3 |
| ML | 9296.4 | 791.6 | 496.0 | 260.2 | 267.4 | 52.6 | 154.0 | 48.7 | 688.6 | 118.1 | 8.8 |
| MM | 8134.6 | 1172.1 | 270.9 | 90.3 | 242.8 | 40.3 | 27.0 | 6.3 | 3463.8 | 467.1 | 2.3 |
| PH | 7467.4 | 430.7 | 166.9 | 46.7 | 225.8 | 29.9 | 6.8 | 0.2 | 2275.7 | 369.2 | 3.0 |
| PL | 12884.8 | 2154.0 | 383.7 | 89.2 | 170.1 | 16.9 | 143.8 | 46.9 | 1114.5 | 176.2 | 9.3 |
| PM | 9127.2 | 854.6 | 230.2 | 35.4 | 152.0 | 20.4 | 30.6 | 7.5 | 3806.0 | 436.6 | 2.3 |
| TH[3] | 6890.1 | 656.9 | 247.5 | 36.1 | | | | | 3249.7 | 266.2 | 2.2 |
| TL | 10811.9 | 1545.5 | 320.5 | 69.8 | 188.6 | 12.3 | 276.2 | 76.6 | 390.8 | 58.3 | 13.0 |
| TM[3] | 7945.8 | 907.5 | 329.0 | 102.2 | | | | | 1472.9 | 141.7 | 5.6 |

*note* : all root, moss, forb, and stand biomass data were assumed to be 45% organic carbon. Samples and *n* were explained in Chapter 1. Ground layer biomass was measured through 1 $m^2$ harvests of everything to brown moss (0.2 $m^2$ for moss plots) at each site; $n$ = 3.

[1] Total soil organic carbon includes Oi, Oe+Oa, A, and mineral soil (5 cm) horizons.

[2] Stand biomass was estimated from an allometric relationship developed from whole-tree harvests at sites including ML & PL, among others (Yarie et al. 2006; *SNRAS misc. publ.* ).

[3] Data not available after wildfires occurred in 2004.

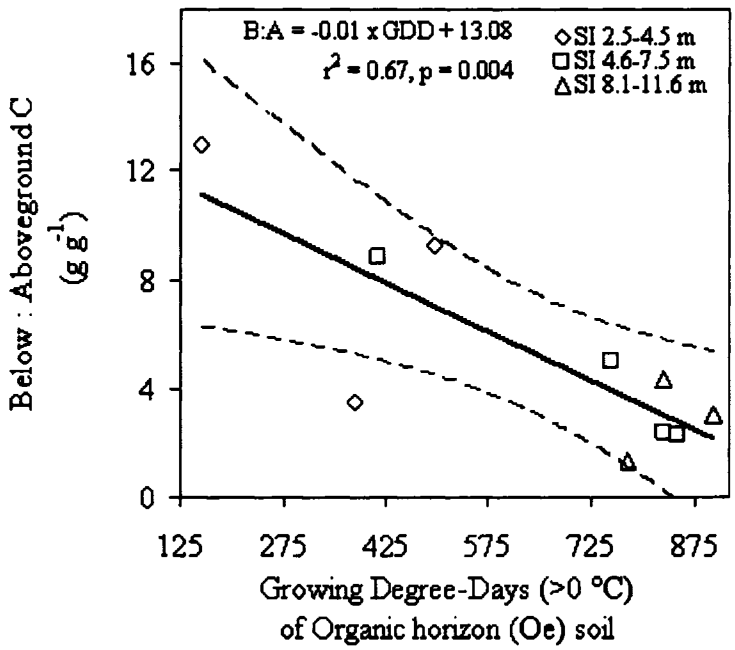


Figure 4.1: Belowground C pools decreased relative to aboveground C pools. Carbon allocation in belowground pools (Oi, Oe+Oa, A, and 5 cm B horizons, and coarse (>2 mm) roots) decreased relative to aboveground C pools (stand, vegetation, and moss) as soil temperatures increased across the four productivity gradients studied. Sites were grouped by their relative level of productivity (Site Index; stand height (m) at 50 y). Dashed lines are a 95% confidence interval about the values. B:A values are those presented in Table 1.