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Interannual variability of surface energy exchange depends on stand age in a boreal forest fire chronosequence

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[1] Understanding how changes in the boreal fire regime will affect high latitude climate requires knowledge of the sensitivity of the surface energy budget to shifts in vegetation cover. We measured components of the surface energy budget in three ecosystems that were part of a fire chronosequence in interior Alaska for 3 years. Our sites were within the perimeter of stand-replacing fires that occurred in 1999, 1987, and ~1920 (hereafter referred to as the 1999-burn, the 1987-burn, and the control). Vegetation cover consisted primarily of sparse short grasses at the 1999-burn, aspen and willow (deciduous trees and shrubs) at the 1987-burn, and black spruce (evergreen conifer trees) at the control. Averaged over the 3 years of our study, annual net radiation decreased by approximately 25% at the 1999-burn and 30% at the 1987-burn, relative to the control. Sensible heat decreased by an even larger amount, by approximately 57% for the 1999-burn and 44% for the 1987-burn as compared with the control. Climate during spring and summer varied considerably among the 3 years. The three stands responded differently to this climate variability with consequences for surface energy exchange. As a result of earlier snow cover loss in 2003 and 2004, net radiation during spring increased substantially in the recently disturbed stands, but not in the control. In response to a sustained summer drought in 2004, latent heat decreased more in the 1987-burn during August than in the control. Our results imply that a shift in plant functional types expected to accompany increases in boreal fire activity may amplify interannual climate variability during both spring and summer.

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1. Introduction

[2] Fire disturbance plays an important role in shaping species composition and landscape diversity in boreal forest ecosystems [*Wein and MacLean*, 1983; *Kasischke and Stocks*, 2000; *Johnstone and Kasischke*, 2005]. Postfire succession and changes in soil properties, in turn, affect ecosystem processes that control surface energy exchange and hydrological and biogeochemical cycles in northern regions [*Kurz and Apps*, 1995; *Zimov et al.*, 1999; *Chapin et al.*, 2000; *Harden et al.*, 2000; *Kasischke and Stocks*, 2000; *Wirth et al.*, 2002; *O'Neill et al.*, 2003; *Bond-Lamberty et al.*, 2004; *Kasischke and Johnstone*, 2005; *Amiro et al.*, 2006; *Goulden et al.*, 2006]. The pathways by which boreal forest fire influences the surface energy budget are multiple. Removal of the canopy overstory following fire leads to an increase in ground heat flux [*Chambers and Chapin*, 2002; *Liu et al.*, 2005] and active layer depth [*Viereck*, 1982], a reduction in canopy roughness [*Chambers and Chapin*, 2002; *Chambers et al.*, 2005], and a reduction in the energy available to drive sensible and latent heat fluxes [*Chambers and Chapin*, 2002; *Liu et al.*, 2005].

[3] Albedo in postfire ecosystems increases during fall, winter, and spring because of loss of canopy cover and increased snow exposure. In the first few decades after fire, albedo ranges between 0.3–0.7, levels that are substantially higher than those observed in late succession evergreen conifer forests (typically ~0.2) [*Betts and Ball*, 1997; *Amiro et al.*, 2006; *Randerson et al.*, 2006]. During summer, the presence of black carbon on boles and on the soil surface decreases albedo below that typically found in late succession forests for the first 1–2 years [e.g., *Yoshikawa et al.*, 2002]. As this black carbon coating is lost and with establishment of highly reflective grasses and deciduous shrubs, albedo increases above prefire levels [*Randerson et al.*, 2006] (A. M. S. McMillan and M. L. Goulden, Age-dependent variations in the biophysical properties of boreal forests, submitted to *Global Biogeochemical Cycles*, 2008, hereinafter referred to as McMillan and Goulden, submitted manuscript, 2008). After 3–5 decades, albedo gradually decreases during both snow-covered and snow-free seasons [*Randerson et al.*, 2006; McMillan and Goulden, submitted manuscript, 2008], probably from an increase in evergreen conifer canopy cover and the mortality of broadleaf decid-

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uous trees. As compared with conifers, broadleaf deciduous trees tend to have higher leaf, branch, and canopy surface reflectance values [*Roberts et al.*, 2004].

[4] Removal of the overstory canopy after fire decreases surface roughness and leads to more absorption of shortwave radiation by the soil surface during summer. As a consequence of these changes, surface soil temperatures increase by several degrees in recently burned stands (e.g., up to 7°C) during summer [*Liu et al.*, 2005]. The increases in surface temperature cause outgoing longwave radiation to increase during the first few years after fire, and as a consequence, a decrease in net radiation [*Chambers and Chapin*, 2002]. During intermediate stages of succession, net radiation in summer may remain low as a consequence of higher surface albedos associated with broadleaf deciduous plant canopies [*Liu et al.*, 2005]. Fire-induced changes in snow cover, surface roughness, and species composition also have important consequences for the partitioning of net radiation into sensible and latent heat fluxes. During spring, increased snow exposure under deciduous tree and shrub species reduces available energy for both sensible and latent heat fluxes. During summer, these same species have higher stomatal and canopy conductance [*Dang et al.*, 1997; *Hogg et al.*, 2000], causing more of the available energy to flow into latent heat as compared with energy flow in evergreen conifers [*Eugster et al.*, 2000; *Chambers and Chapin*, 2002].

[5] Considering the changes in the surface energy budget described above, more boreal forest fire would probably lead to cooler northern air temperatures because of increases in surface albedo and decreases in net radiation and sensible heat fluxes [e.g., *Bonan et al.*, 1992; *Snyder et al.*, 2002; *Govindasamy et al.*, 2001; *Bala et al.*, 2007]. Concurrent fire effects on aerosols and greenhouse gases (including CO_2, CH_4, and O_3) probably offset some of this cooling at a global scale [*Randerson et al.*, 2006].

[6] While we have some understanding of how fire influences seasonal and annual surface energy budgets, we know less about how fire-induced changes in land cover may influence the year-to-year variability of surface energy exchange. If the interannual variability of surface energy exchange increases within terrestrial ecosystems, for example, it may cause weather and climate to become more variable [e.g., *Schubert et al.*, 2004]. Different types of land cover (e.g., grasslands, deciduous forests, and conifer forests) may interact differently with larger-scale climate anomalies, in some cases amplifying and in other cases damping these variations. Key ecosystem factors that regulate this interaction include the plasticity of leaf phenology to climate variability, the sensitivity of canopy conductance to drought, and the modulation of snow cover and surface albedo by plant height and vegetation structure [e.g., *Guillevic et al.*, 2002; *Koster et al.*, 2006; *Euskirchen et al.*, 2007]. All of these factors in boreal forests vary with postfire stand age.

[7] Eddy covariance measurements show that species composition within alpine and boreal forests modulates the response of gross primary production (GPP) and net ecosystem exchange (NEE) to interannual variability in climate [*Black et al.*, 2000; *Arain et al.*, 2002; *Barr et al.*, 2002; *Monson et al.*, 2005; *Welp et al.*, 2007]. Specifically, carbon fluxes from broadleaf deciduous-dominated ecosystems appear more sensitive to spring air temperatures and summer drought than evergreen conifer-dominated ecosystems. These different ecosystem sensitivities, in turn, may have implications for feedbacks to interannual climate variability because they are associated with different degrees of coupling between carbon and energy fluxes by means of stomatal regulation.

[8] In this study, we assessed interannual chances in the surface energy budget measured in three ecosystems that were part of a fire chronosequence in interior Alaska. Our objective was to quantify how surface energy exchange varied with stand age in response to interannual variability in climate over a three-year period.

2. Site Description

[9] Our three tower sites were a part of a fire chronosequence, with stand replacing fires that occurred in 1999, 1987, and ~1920 (hereafter referred to as the 1999-burn, the 1987-burn, and the control, respectively). All three sites were located near Delta Junction (63°54′N, 145°40′W), just to the north of the Alaska Range in interior Alaska. Climate information for 2002, 2003, and 2004, the period of our measurements, is provided in Figure 1. Over this three-year interval, summer temperatures increased and precipitation decreased during each successive year. Soils at these sites consisted of well-drained silty loams on top of glacial moraines [*Manies et al.*, 2004]. *Mack et al.* [2008] measured aboveground net primary production and standing biomass for the different species at these three sites.

[10] The 1999-burn was located south of Delta Junction (63°55′18″N, 145°44′44″W). The Donnelly Flats fire burned approximately 7600 ha of black spruce (*Picea mariana*) during 11–18 June 1999 [*Alaska Fire Service*, 2006]. By 2002, approximately 30% of the surface was covered by bunch grasses (*Festuca altaica*) and deciduous shrubs (that had a height less than 1 m). Standing dead black spruce boles had a density of 2691 ± 778 trees per hectare (M. C. Mack, personal communication, 2004) and a mean height of ~4 m. The soil was well drained, with approximately 30 cm of loess soil on top of a glacially deposited gravel layer [*Manies et al.*, 2004; *Neff et al.*, 2005]. A uniform tower fetch extended for more than 1 km in all directions within the burn perimeter. MODerate Resolution Imaging Spectroradiometer (MODIS) enhanced vegetation index (EVI) observations from Terra [*Huete et al.*, 2002] show that leaf area increased monotonically from 2000 through 2004 (Figure 2).

[11] The 1987-burn was located southeast of Delta Junction (63°55′4″N, 145°22′23″W). By 2002, heterogeneous aspen and willow dominated the overstory (*Populus tremuloides* and *Salix* spp.). The aspen had a mean canopy height of 5 m. The sparse understory vegetation included shrubs (*Salix* spp., *Ledum paustre*, *Rosa acicularis, Vaccinium uliginosum*, and *Vaccinium vitis-idaea*), black spruce (*Picea mariana*), and grasses (*Festuca* spp. and *Calamagrostis lapponica*), separated by patches of moss in open areas (*Polytrichum* spp.). Black spruce boles killed by the 1987 fire had a density of 3200 ± 1329 dead trees per hectare [*Chambers and Chapin*, 2002]. Approximately half of the dead boles remained upright in 2004 - the other half had fallen over or had become entangled with other boles.

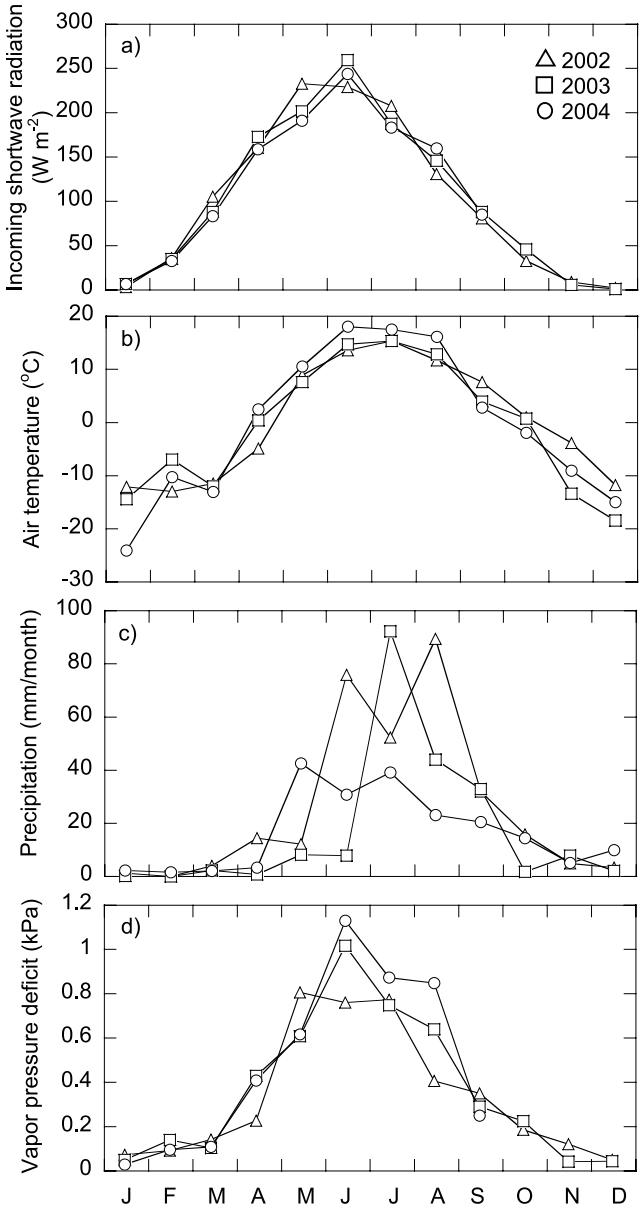

Figure 1. Seasonal patterns of (a) incoming shortwave radiation (W m^{-2}), (b) above-canopy air temperature (°C), (c) precipitation (mm/month), and (d) vapor pressure deficit (VPD; kPa) during 2002, 2003, and 2004. The incoming shortwave radiation represents the average of observations from upward looking Eppley pyranometers at the 1999-burn and the control. The above canopy air temperature time series represents the monthly average of measurements collected above the canopy at all three sites. The precipitation data shown here were measured at the climate monitoring station in nearby Big Delta, Alaska and include both rain and snow components [*WRCC*, 2007].

The burn scar from the tower extended for more than 1 km to the north, and approximately 500 m to the east, west, and south.

[12] The control was located approximately 5 km to the south of the 1999-burn (63°53′17″N, 145°44′22″W). The canopy overstory consisted of homogeneous stands of black spruce (*Picea mariana*) with a mean canopy height of 4 m, a stand density of 3744 ± 462 trees per hectare, and a mean age of approximately 80 years based on tree ring measurements [*Mack et al.*, 2008]. The sparse understory consisted primarily of shrubs (*Ledum palustre*, *Vaccinium uliginosum*, and *V. vitis-idaea*). The dominant ground cover species were feathermoss (*Pleurozium schreberi* and *Rhytidium rugosum*) and lichen (*Cladonia* spp. and *Stereocaulon spp.*). Moss and soil organic layers had a mean depth of approximately 11 cm to mineral soil [*Manies et al.*, 2004]. The site extended from the tower for more than 1 km to the south, west, and north, with the shortest fetch to the east (approximately 200 m).

3. Instruments and Methods

[13] Turbulent fluxes of sensible heat, latent heat, and carbon dioxide were measured using an eddy covariance system on a micrometeorological tower at each site. Details of the flux measurements and analysis methods are reported in *Liu et al.* [2005]. Briefly, wind velocity and sonic

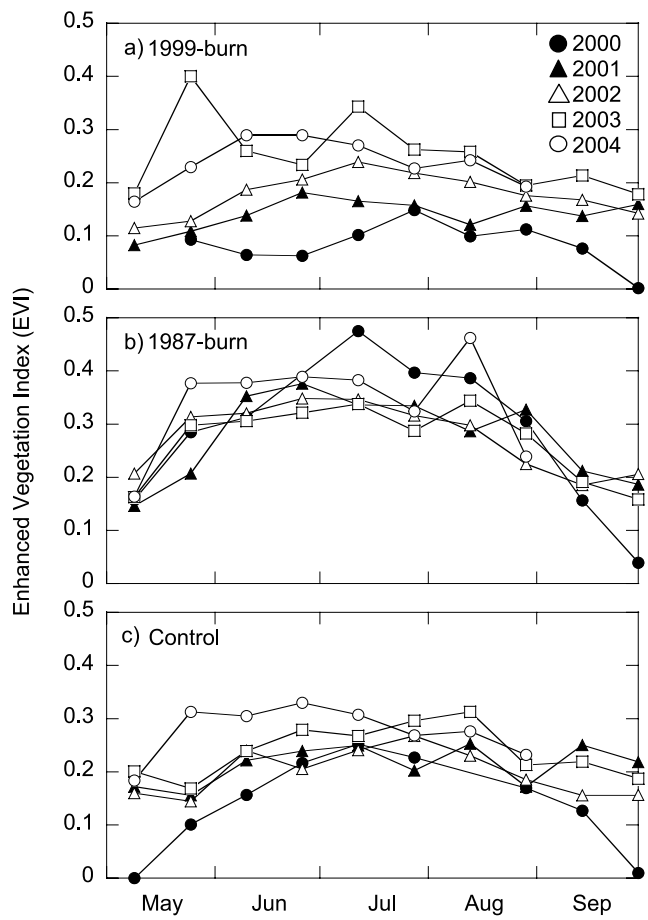

Figure 2. Variation of enhanced vegetation index (EVI) during 2000, 2001, 2002, 2003, and 2004 derived from MODIS satellite data for the three sites. The values presented here were the average of a 1 km × 1 km pixel centered at each tower. The data were obtained from the MODIS Subsetting and Visualization Tool for North America from Oak Ridge National Laboratory Distributed Active Archive Center [*ORNL DAAC*, 2006].

temperature were measured with a three-dimensional sonic anemometer/thermometer (model CSAT-3, Campbell Scientific, Inc.). H_2O and CO_2 densities were measured with an open-path infrared gas analyzer (model LICOR-7500, LiCor Inc.) at a 10-Hz acquisition frequency. Turbulent fluctuations were calculated as the difference between the instantaneous and the 30-min mean quantities. Sonic temperature was converted to air temperature following the procedure suggested by *Campbell Scientific, Inc. Instruction Manual* [2006]. Vertical fluxes of sensible (H) and latent heat (LE) were obtained via 30-min mean covariance between vertical velocity (w') and the respective air temperature (T') and water vapor density (ρ') fluctuations. We applied density corrections for latent heat fluxes following the approach described by *Webb et al.* [1980].

[14] In addition to the eddy covariance measurements, we measured net radiation (REBS Q-7.1, Radiation and Energy Balance Systems [REBS], Inc.) at all three sites. Incoming and outgoing shortwave radiation fluxes were measured only at the 1999-burn and control sites (Precision Spectral Pyranometers, Eppley Lab., Inc.). At all three sites we measured air temperature and humidity (HMP45C, Vaisala, Inc.), wind speed (models 03001 and 03101, RM Young, Inc.), and soil temperature. Soil temperature profiles were measured by placing thermocouples at 0, 2.5, 5, 10, and 20 cm depths below the surface at each site. We also measured the soil heat flux (G_{10}) at a depth of 10 cm using soil heat flux plates (model HFT3, REBS, Inc.). The soil heat flux (G_0) at the surface was estimated using a thermal conductivity equation with the thermocouple temperature profiles. The method for calculating soil thermal conductivity was obtained from National Center for Atmospheric Research Common Land Model 2.0, which is based on the work by *Farouki* [1981]. We measured volumetric soil water content at each site with water content reflectometers (model CS615-L, Campbell Scientific, Inc.) at depths of 2, 4, 11, and 37 cm at the 1987-burn, at depths of 2, 4, 25, and 40 cm at the 1999-burn, and at depths of 2, 5, 22, and 27 cm at the control. In our analysis, we used precipitation data measured at the Big Delta climate monitoring station, approximately 15 km north of Delta Junction [*Western Regional Climate Center* (*WRCC*), 2007].

[15] Half-hourly data of net radiation and sensible and latent heat fluxes in 2002, 2003, and 2004 are presented in Figure 3. Based on these data, we analyzed the closure of the surface energy balance for three years using the 30-min data available during each season. We calculated linear regression coefficients (slope and intercept) from an ordinary least squares fit of H + LE versus Rn − G_0, which accounts for errors in both coordinates, to evaluate the surface energy closure. We did not find a relationship between the closure of surface energy balance and wind direction, and so we did not use wind direction as a criteria for excluding data from our analysis. Seasonal variations and site differences in closure for 2002 are summarized in *Liu et al.* [2005]. During 2002–2004, seasonal mean values of the slopes and intercepts of H + LE versus Rn − G_0 ranged from 0.69 to 0.81 for the 1999-burn, from 0.75 to 0.86 for the 1987-burn, and from 0.71 to 0.87 for the control. Closure during summer was higher than during fall, winter, or spring at each of the three sites. Closure of the surface energy budget during summer was highest at the control followed by the 1987-burn and then by the 1999-burn (Table 1). These closure estimates are within the range of those reported by the FLUXNET community [*Wilson et al.*, 2002].

4. Results

4.1. Three-Year Mean Surface Energy Budget

[16] Changes in canopy structure and species composition along the chronosequence caused changes in midday surface albedo (Figure 4 and Table 2). The three-year mean midday spring albedo (March, April, and May) was 0.44 for the 1999-burn and 0.16 for the control. As a result, outgoing shortwave radiation at the surface was 33 W m^{-2} higher during spring at the 1999-burn than at the control (Figure 5). Differences in albedo persisted during summer (June, July, and August) with midday surface albedo at the 1999-burn (0.13) higher than at the control (0.08) (Table 2 and Figure 4). This corresponded to an increase of 11 W m^{-2} in outgoing shortwave radiation at the 1999-burn, relative to the control (Figure 5).

[17] Figure 6 shows the three-year mean seasonal cycle of net radiation, sensible heat, and latent heat for the three stands. In the spring (March, April, and May), net radiation was lower by about 32% (31 W m^{-2}) at the 1999-burn and 34% (32 W m^{-2}) at the 1987-burn as compared with the control (95 W m^{-2}) (Table 2 and Figure 6), primarily as a result of the increased albedo [*Liu et al.*, 2005]. Soil heat flux at the depth of 10 cm increased at the recently disturbed stands as compared with the control (Table 2). As a consequence, there was less available energy for warming the atmosphere through sensible and latent heating. Sensible heat fluxes decreased by 61% (31 W m^{-2}) and 39% (19 W m^{-2}) at the 1999-burn and the 1987-burn, respectively, as compared with the control (50 W m^{-2}). Latent heat fluxes decreased in parallel, by 18% (4 W m^{-2}) and 21% (5 W m^{-2}) at the 1999-burn and 1987-burn, respectively, as compared with the control (23 W m^{-2}) (Table 2). More of the available energy in postfire ecosystems during spring was dissipated as sensible heat (H/Rn = 30%, 49%, and 53% for the 1999-burn, 1987-burn, and control, respectively) than as latent heat (LE/Rn = 29%, 28%, and 24% for the 1999-burn, 1987-burn, and control, respectively).

[18] During summer (June, July, and August), net radiation remained lower in the recently disturbed ecosystems than in the control, although the differences were smaller and probably were caused by different factors at the 1999-burn and 1987-burn. Specifically, decreased surface roughness and increased soil surface temperatures at the 1999-burn probably increased outgoing longwave radiation whereas higher albedo at the 1987-burn probably increased outgoing shortwave radiation (e.g., Table 4 of *Liu et al.* [2005]). Net radiation was reduced by 14% (19 W m^{-2}) and by 17% (24 W m^{-2}) at the 1999-burn and 1987-burn relative to the control (139 W m^{-2}) (Table 2). Sensible heat fluxes were reduced both at the 1999-burn (24%; 13 W m^{-2}) and at the 1987-burn (31%; 17 W m^{-2}). Although latent heat fluxes declined by 23% (12 W m^{-2}) at the 1999-burn, there was 13% (7 W m^{-2}) increase at the 1987-burn as compared with the control (51 W m^{-2}). Much of the increase in latent heat at the 1987-burn (above that mea-

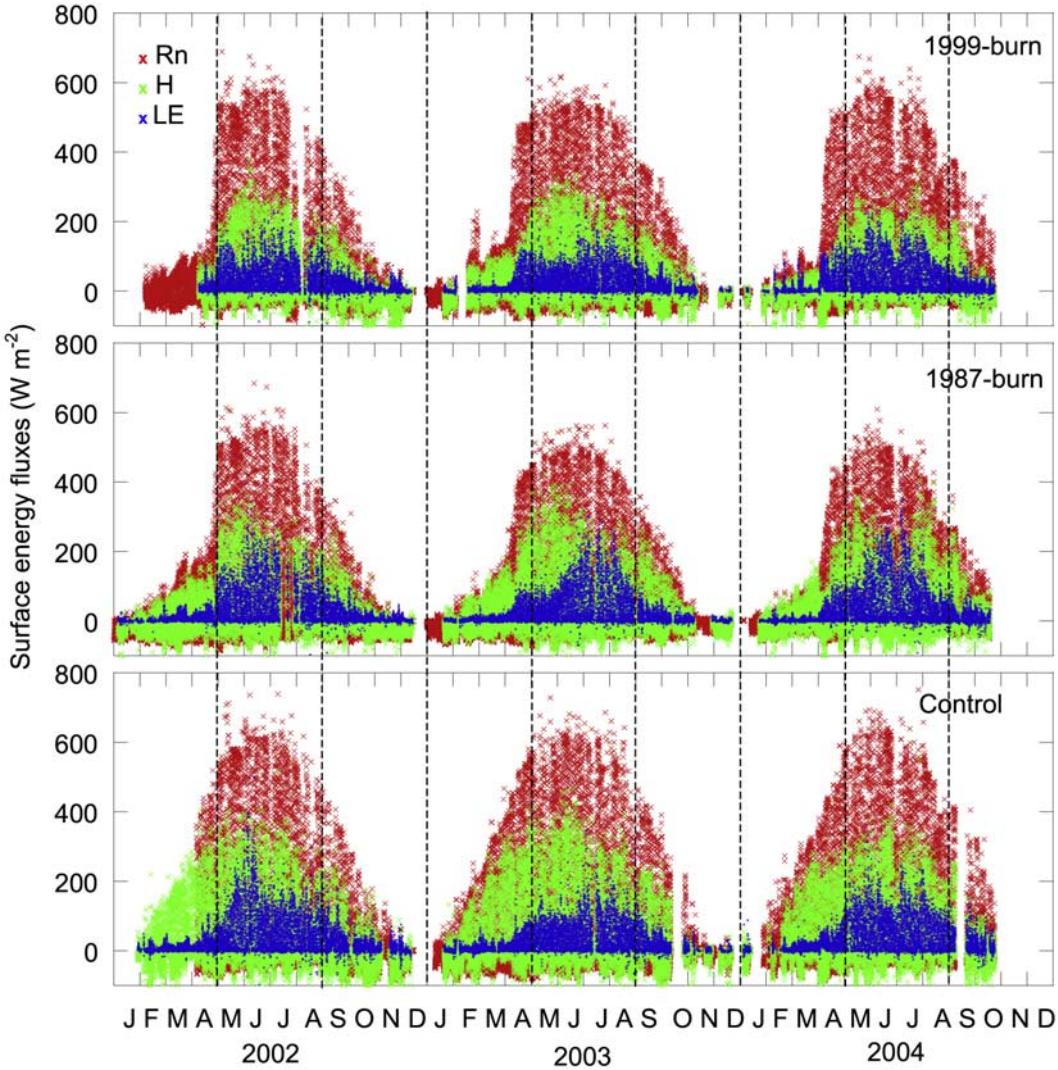

Figure 3. Half-hour time series data of net radiation, sensible heat flux, and latent heat flux measured at the three sites for 2002, 2003, and 2004. Many of the data gaps were caused by power interruptions and instrument failures.

sured at the control) occurred during the months of June and July (Figure 6).

[19] In fall and winter, all of the components of the surface energy budget were small, reflecting low levels of incoming shortwave radiation, and making it challenging to identify stand-specific differences. Throughout both fall and winter, net radiation was smaller (more negative) in the recently disturbed sites than in the control (Table 2 and Figure 6). During Nov–Feb, net radiation was negative, reflecting a loss of longwave radiation from the surface that was partly compensated by negative sensible heat fluxes (a flow of heat from the atmosphere to the land surface) and a flow of heat from subsurface soils towards the surface.

[20] For the three-year mean annual budget, the averaged annual net radiation decreased by 25% (15 W m^{-2}) for the 1999-burn, and by 30% (18 W m^{-2}) for the 1987-burn as compared with the control (60 W m^{-2}) (Table 2). Postfire changes in canopy structure and physiology caused substantial changes in partitioning of available energy into sensible and latent heat fluxes. The three-year averaged

Table 1. Energy Balance Closure During Summer (JJA)

	1999-Burn				1987-Burn				Control			
Year	N	Slope	Intercept	r^2	N	Slope	Intercept	r^2	N	Slope	Intercept	r^2
2002	3780	0.81	8.7	0.82	3217	0.81	7.9	0.81	4164	0.86	9.7	0.88
2003	2407	0.78	7.9	0.88	3857	0.84	5.1	0.85	4031	0.84	7.2	0.87
2004	4164	0.79	4.2	0.87	4201	0.86	−1.4	0.86	4299	0.87	5.2	0.89

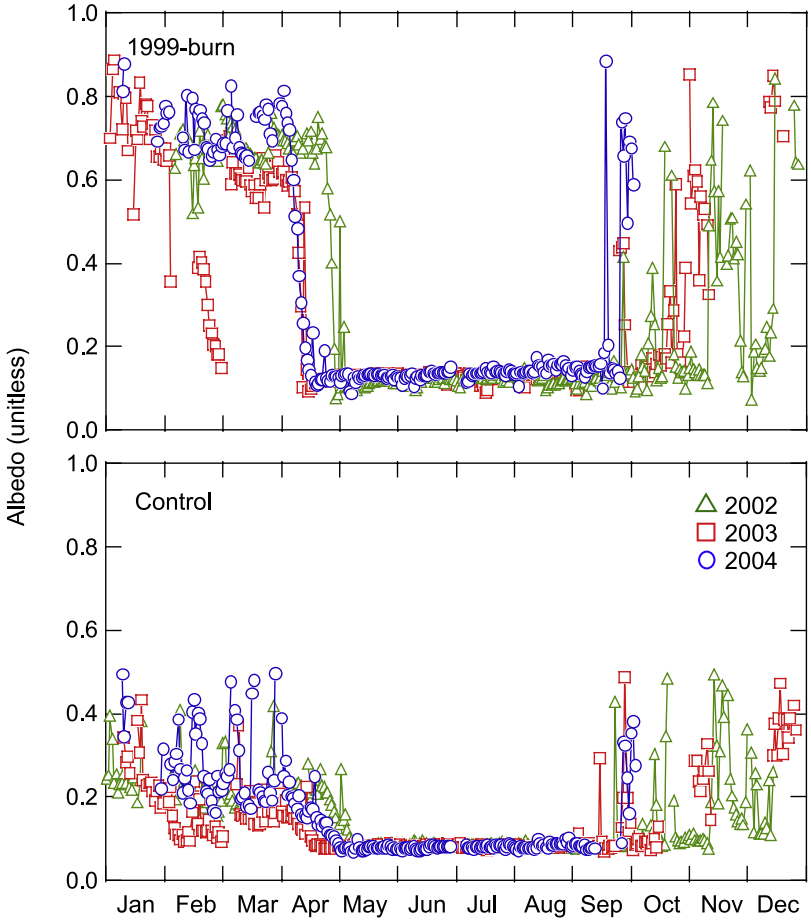

Figure 4. Seasonal cycle of midday albedo measured at both the 1999-burn and the control for 2002, 2003, and 2004. During fall, winter, and spring, large variations in midday albedo were caused by snowstorms. Values represent averages between the hours of 11:00 and 13:00 local time.

annual sensible heat fluxes were reduced by 57% (13 W m^{-2}) for the 1999-burn and by 44% (10 W m^{-2}) for the 1987-burn, relative to the control (23 W m^{-2}). The latent heat fluxes had a more complex pattern. The annual mean latent heat flux at the 1999-burn declined by 23% (5 W m^{-2}) as compared with the control (22 W m^{-2}). In contrast, there was no discernable difference in annual mean at the 1987-burn and the control: 21 W m^{-2} vs. 22 W m^{-2}.

4.2. Interannual Variability in Surface Energy Budget

4.2.1. Variations in Surface Albedo

[21] Interannual variations in snow cover and the timing of snowmelt in spring had important consequences for albedo, with larger absolute (and relative) interannual differences at the 1999-burn than at the control (Figure 4). The springs of 2002 through 2004 had progressively warmer air and soil temperatures (Figure 1 and Table 3). In parallel, snowmelt during 2003 and 2004 occurred 16 days and 18 days earlier than in 2002 in the 1999-burn (Table 4). Snowmelt at the control was delayed relative to the 1999-burn. Snowmelt at the control during 2003 and 2004 occurred 18 days and 20 days earlier than in 2002 (Table 4). The difference between average presnowmelt and postsnowmelt midday albedo was 0.52, 0.44, and 0.53 in 2002, 2003, and 2004, respectively, for the 1999-burn, and 0.10, 0.06, and 0.11 in 2002, 2003, and 2004, respectively, for the control (Table 4). The timing of snowmelt influenced the mean albedo during spring. The average spring albedo for the 1999-burn decreased by about 0.14 and 0.09 in the 2003 and 2004 springs, respectively, as compared with 2002. In contrast, the average albedo for the control changed by a smaller amount - decreasing by 0.03 in 2003 and remaining unchanged in 2004, relative to 2002.

[22] The year-to-year variations in spring albedo had a large impact on outgoing shortwave radiation, and may have implications for climate feedbacks. Decreases in spring albedo at the 1999-burn led to decreases in the outgoing shortwave radiation by 26 W m^{-2} in 2003 and by 22 W m^{-2} in 2004 as compared with 2002. The magnitude of interannual variability at the 1999-burn (i.e., 26 W m^{-2} in 2003) was substantial relative to the spring mean (i.e., 54 W m^{-2}). In contrast, the magnitude of interannual variability of outgoing shortwave radiation at the control was substantially smaller, decreasing by 7 W m^{-2} during 2003 and 4 W m^{-2} during 2004 as compared with 2002 and relative to a spring mean of 21 W m^{-2}. These results imply that during years with greater snow cover, postfire ecosystems may have a greater potential to cool

Table 2. The 2002–2004 Mean Radiation Budget and Energy Fluxes

Site	$K_{\downarrow}$	Rn	G_{10}	H	LE	α
		Spring (Mar–Apr–May)[a]				
1999-burn		64.7	4.6	19.4	18.8	0.44
1987-burn	155.0	63.3	3.6	30.9	18.0	–
Control		95.2	0.7	50.2	22.8	0.16
		Summer (Jun–Jul–Aug)				
1999-burn		119.4	5.4	42.3	38.7	0.13
1987-burn	194.2	115.1	7.7	38.2	57.0	–
Control		138.7	6.5	55.5	50.5	0.08
		Autumn (Sep–Oct–Nov)				
1999-burn		10.9	−3.0	−6.2	9.3	0.24
1987-burn	43.8	9.4	−2.3	−1.0	8.7	–
Control		17.5	−1.1	−5.4	13.0	0.14
		Winter (Dec–Jan–Feb)				
1999-burn		−14.3	−4.2	−16.6	1.9	–
1987-burn	14.2	−19.0	−4.3	−17.5	1.8	–
Control		−10.1	−5.9	−10.6	2.5	–
		All Year (Jan–Dec)				
1999-burn		45.2	0.7	9.7	17.2	–
1987-burn	102.9	42.2	1.2	12.6	21.4	–
Control		60.3	0.1	22.5	22.2	–

[a]Monthly mean diurnal cycles were constructed using all available data during 2002–2004. In a second step, the three monthly mean values were averaged to obtain the seasonal estimate. $K_{\downarrow}$: incoming shortwave radiation (W m^{-2}) averaged from the 1999-burn and the control; Rn: net radiation (W m^{-2}); G_{10}: soil heat flux (W m^{-2}) measured at a depth of 10 cm; H: sensible heat flux (W m^{-2}); LE: latent heat flux (W m^{-2}); α: midday albedo averaged over the hours of 11:00 LT and 13:00 LT.

regional air temperatures and prolong the duration of the snow-covered period.

[23] Summer albedo increased monotonically from 2002 through 2004 at the 1999-burn, from 0.12 in 2002 to 0.13 in 2003 and 0.14 in 2004, probably from increased grass and shrub cover and a loss of black carbon within the burn perimeter [*Randerson et al.*, 2006]. The increases in the summer albedo in the 1999-burn caused outgoing shortwave radiation to increase by 4 W m^{-2} in 2003 and 5 W m^{-2} in 2004. In contrast, albedo in the control remained the same at about 0.08 during each of the three years of the study (Figure 4).

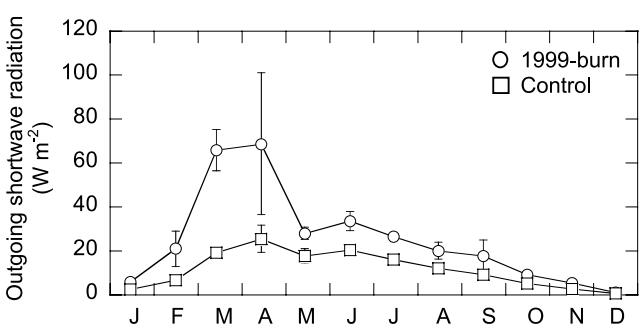

Figure 5. Observations of outgoing shortwave radiation from the 1999-burn (circles) and the control (squares), measured using downward looking Eppley pyranometers above the canopy and averaged over 2002–2004. Error bars denote standard deviations of the year-to-year differences. These data are the same as those shown in Figure S6 of *Randerson et al.* [2006].

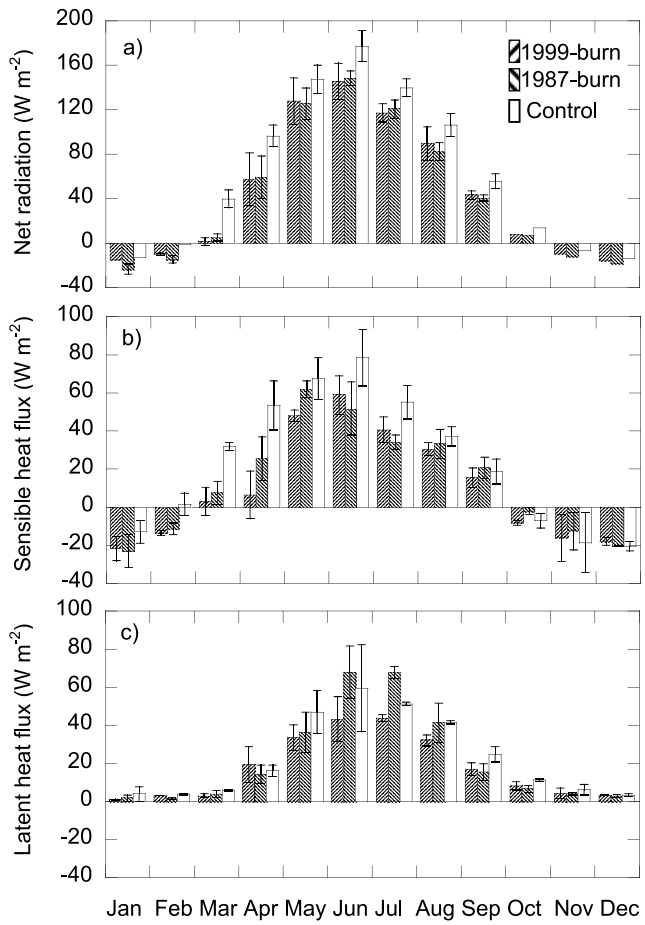

Figure 6. Three-year mean (a) monthly net radiation, (b) monthly sensible heat flux, and (c) latent heat flux at the surface for the three sites. Error bars denote standard deviations of the year-to-year differences.

4.2.2. Year-to-Year Changes in the Surface Energy Budget Spring Warming

[24] Two factors appeared to contribute to the observed differences among sites in the year-to-year levels of net radiation during spring. In the recently disturbed stands, year-to-year differences in net radiation tracked changes in the timing of snowmelt, whereas at the control, year-to-year differences followed incoming shortwave radiation (that was at a maximum in 2002). Net radiation at the 1999-burn

Table 3. Year-to-Year Climate Differences During Spring and Summer

	Air Temperature[a]		Soil Temperature		Precipitation	
Year	Spring	Summer	Spring	Summer	Spring	Summer
2002	5.9	13.5	−4.3	8.9	31	218
2003	7.6	14.3	−3.5	8.2	11	144
2004	10.3	17.2	−1.1	10.3	48	93

[a]Seasonal averaged above-canopy air temperature (°C), soil temperature at 10 cm (°C) averaged across all three sites, and total precipitation (mm) in spring (March, April, and May) and summer (June, July, and August) in 2002, 2003, and 2004. Precipitation data are from the Big Delta climate station [*WRCC*, 2007].

Table 4. Snowmelt Timing at the 1999-Burn and the Control

	Timing of Snowmelt[a]			Presnowmelt Albedo[b]			Postsnowmelt Albedo[b]		
Sites	2002	2003	2004	2002	2003	2004	2002	2003	2004
1999-burn	4/27	4/11	4/9	0.68	0.60	0.74	0.16	0.16	0.21
Control	5/2	4/14	4/12	0.19	0.15	0.20	0.09	0.09	0.09

[a]Timing of snowmelt was identified based on the midday albedo time series from February 1 to July 31, using the following procedure. Individual daily data points were replaced through linear interpolation when their departures relative to the 9-point mean exceeded 1.5σ of the 9 points (where σ denotes standard deviation). A 9-point boxcar filter was then applied to obtain a smoothed time series data. The timing of snowmelt was defined as the day when albedo decreased by over 50% of the difference between presnowmelt and postsnowmelt albedo. With this algorithm, the presnowmelt and postsnowmelt albedo levels were defined as the average midday albedo during February and March and in June and July, respectively, using the smoothed data set.

[b]The presnowmelt albedo and postsnowmelt albedo values were obtained by averaging 14-day midday albedo before and after the timing of snowmelt, respectively.

increased by 8 W m^{-2} in 2003 and by 4 W m^{-2} in 2004, relative to the spring of 2002 (Figure 7a and Table 5). Net radiation also increased at the 1987-burn, by 6 W m^{-2} in both 2003 and 2004 (Figure 7d and Table 5). In contrast, net radiation at the control decreased by 1 W m^{-2} in 2003 and by 3 W m^{-2} in 2004 (Figure 7g and Table 5). At the control, the evergreen spruce canopy masked the interannual variability in snow cover, and as a consequence, variations in net radiation were more sensitive to interannual differences in total incoming shortwave radiation.

[25] We estimated that a one-day advance of snowmelt during spring, on average, increases net radiation by 1.0 W m^{-2} (1.6%) per spring at the 1999-burn and by 0.4 W m^{-2} (0.4%) per spring at the control (Table 6). This information helps to quantify how the strength of the snow-climate feedback varies with vegetation cover [*Euskirchen et al.*, 2007]. Similarly, a one-day advance of snowmelt increases sensible heat flux by 0.3 W m^{-2} (1.4%) per spring at the 1999-burn and by 0.2 W m^{-2} (0.4%) per spring at the control (Table 6). These results provide additional evidence that recently disturbed ecosystems have more potential to feedback positively with year-to-year variations in climate and snow cover than mature evergreen conifer stands.

[26] In the warmer and drier spring of 2003, sensible heat fluxes increased by 9, 10, and 1 W m^{-2}, and latent heat fluxes decreased by −1, 5, and 6 W m^{-2} at the 1999-burn, 1987-burn, and control, respectively, as compared with the 2002 spring. In contrast, in the warmer and moister spring of 2004, sensible heat flux decreased by 4.5, −0.9, and 13.1 W m^{-2}, and latent heat fluxes increased by 10, 1, and 3 W m^{-2} at the 1999-burn, 1987-burn, and control, respectively, as compared with the 2002 spring (Table 5).

[27] In general, climate-induced changes in the surface energy budget from year-to-year were smaller than those caused by fire disturbance (Figure 6). Our results do suggest, however, that spring warming will weaken the net cooling effect caused by fire-induced changes in surface energy exchange. For example, with consecutive increases in spring air temperatures between 2002 and 2004 (Table 3), differences in net radiation between the 1987-burn and the control showed a decreasing trend, with differences of −42, −30, and −28 W m^{-2} during 2002, 2003, and 2004 springs, respectively. Spring sensible heat flux differences between the two stands showed a similar trend, decreasing from −27 in 2002 to −13 W m^{-2} during 2004.

4.3. Summer Drought

[28] Summers were progressively warmer and drier from 2002 to 2004 (Figure 1 and Table 3). In 2004, severe

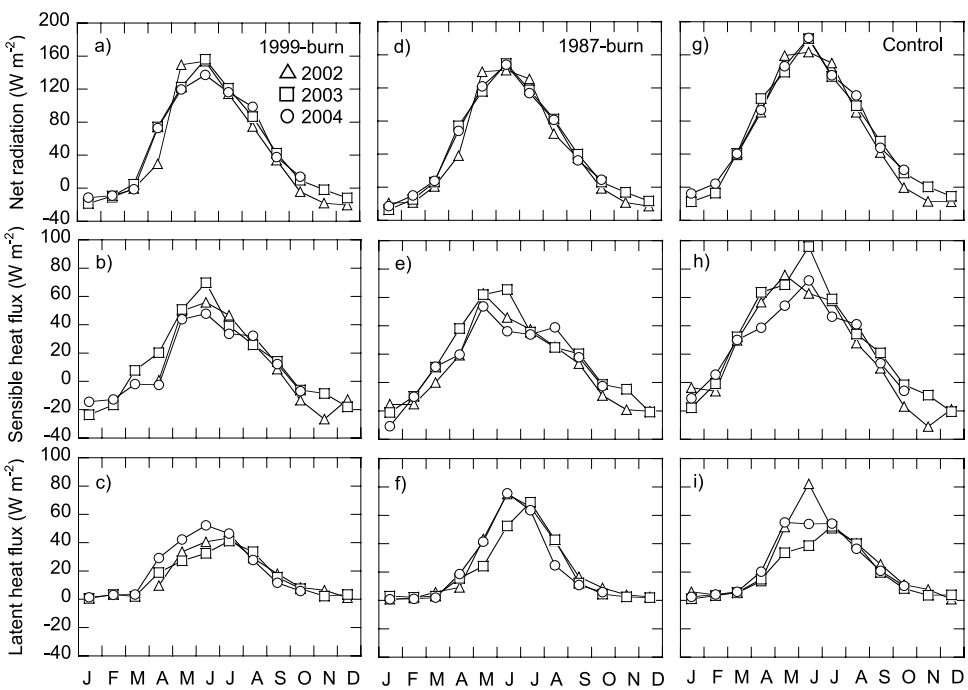

Figure 7. Monthly averaged net radiation, sensible heat flux, and latent heat flux for the three stands during 2002–2004.

Table 5. Seasonally Averaged Radiation Budget and Energy Fluxes During 2002–2004[a]

Sites	$K_\downarrow$[b]	Rn	G_{10}	H	LE	$K_\downarrow$	Rn	G_{10}	H	LE	$K_\downarrow$	Rn	G_{10}	H	LE
	2002					2003					2004				
	Spring (Mar–Apr–May)					*Spring (Mar–Apr–May)*					*Spring (Mar–Apr–May)*				
1999-burn		59.5	4.0	17.7	15.3		67.3	4.5	26.3	16.1		63.8	5.3	13.2	25.0
1987-burn	166.3	59.7	3.0	27.3	19.5	154.2	65.6	3.6	37.0	14.3	144.4	65.9	4.2	28.2	20.7
Control		96.3	−0.5	54.1	23.7		95.8	0.2	55.0	18.1		93.5	2.2	41.0	27.1
	Summer (Jun–Jul–Aug)					*Summer (Jun–Jul–Aug)*					*Summer (Jun–Jul–Aug)*				
1999-burn		114.4	5.7	43.1	37.8		121.3	5.1	45.0	35.8		117.3	5.5	38.0	42.2
1987-burn	189.2	112.6	7.9	36.3	61.0	197.5	116.6	7.7	42.1	55.0	195.8	114.6	7.4	36.3	54.7
Control		134.6	7.0	49.4	57.9		137.4	5.9	63.0	43.8		142.1	6.5	53.2	48.2
	Autumn (Sep–Oct–Nov)					*Autumn (Sep–Oct–Nov)*					*Autumn (Sep–Oct–Nov)*				
1999-burn		3.8	−3.0	−10.4	11.0		16.83.0	−0.2	8.4	–	–	–	–	–	
1987-burn	40.8	5.2	−2.3	−5.0	10.0	45.8	13.42.3	4.7	6.5	–	–	–	–	–	
Control		8.2	−1.1	−12.7	14.8		24.61.1	3.4	10.6	–	–	–	–	–	
	Winter (Dec–Jan–Feb)					*Winter (Dec–Jan–Feb)*					*Winter (Dec–Jan–Feb)*				
1999-burn		−16.1	−5.5	−17.6	1.7		−10.6	−2.8	−15.0	2.6	–	–	–	–	–
1987-burn	14.6	−21.7	−5.4	−17.0	2.6	13.8	−16.4	−3.3	−20.9	1.3	–	–	–	–	–
Control		−14.1	−7.6	−12.5	1.8		−4.6	−4.2	−8.7	3.6	–	–	–	–	–
	All Year (Jan–Dec)					*All Year (Jan–Dec)*					*All Year (Jan–Dec)*				
1999-burn		41.9	0.3	8.4	16.9		48.7	1.0	14.0	15.8	–	–	–	–	–
1987-burn	102.7	38.9	0.8	10.4	23.3	103.1	44.8	1.4	15.8	19.2	–	–	–	–	–
Control		56.3	−0.5	19.6	24.6		63.3	0.2	28.2	19.0	–	–	–	–	–

[a]Since there were no H and LE data available for March of 2002 at the 1999-burn, we filled in the gaps using the means of corresponding data from March of 2003 and 2004. Since there were no Rn data for March of 2002 at the control, we first obtained the mean ratio of Rn/(H + LE) = 1.107 from the data in March of 2003 and 2004, and we then estimated Rn for March of 2002 using the above ratio and H and LE data in March of 2002.

[b]$K_\downarrow$: incoming shortwave radiation (W m^{-2}) averaged from the 1999-burn and the control; Rn: net radiation (W m^{-2}); G_{10}: soil heat flux (W m^{-2}) measured at a depth of 10 cm; H: sensible heat flux (W m^{-2}); LE: latent heat flux (W m^{-2}). There were no data available for the 2004 autumn and winter since the towers were shut down in October of 2004.

drought triggered large-scale fires that burned over 2.6 Mha [*Alaska Fire Service*, 2006]. Precipitation levels during June, July, and August of 2004 were 47% of the 1937 to 2005 mean at the climate monitoring station in nearby Big Delta, Alaska [*WRCC*, 2007]. Temperatures were 2°C warmer during the summer of 2004 than the long-term mean (14.3°C). The combined effect of higher air temperatures and lower than normal precipitation substantially increased vapor pressure deficit (VPD) and decreased soil moisture content in both soil top layer and deep layer during 2004 (Figures 1 and 8). The drought was particularly intense during August of 2004 - with precipitation levels only 46% of the long-term mean and air temperatures more than 3.2°C above normal [*WRCC*, 2007].

[29] During August 2004, incoming shortwave radiation, net radiation, and sensible heat increased at all three sites relative to that observed in 2002. At the control, net radiation increased by 24%, sensible heat increased by 48%, and latent heat decreased by 7% during August of 2004, relative to that observed in 2002 (Figure 7). The response of the 1987-burn was even larger: net radiation increased by 28%, sensible heat increased by 55%, and latent heat decreased by 38%, relative to that observed during August of 2002 (Figure 7). Thus, the Bowen ratio increased from 0.6 to 1.6 from 2002 to 2004 at the 1987-burn, as compared with a 0.7 to 1.2 increase at the control. We excluded a detailed analysis of the 1999-burn because the strong secular increases in leaf area at this site immediately following fire (e.g., Figure 2) probably offset the impacts of drought on energy partitioning.

5. Discussion

[30] Interactions between boreal forest ecosystems and the climate system have received considerable attention since warming may trigger carbon loss from large, physically protected soil carbon pools via several different pathways [e.g., *Smith and Shugart*, 1993; *Chapin et al.*, 2000; *McGuire et al.*, 2006]. These pathways include melting of permafrost and the creation of thermokarst features that

Table 6. Changes in the Outgoing Shortwave Radiation ($\Delta K\uparrow$; W m^{-2}), Net Radiation (ΔRn; W m^{-2}), and Sensible Heat Flux (ΔH; W m^{-2}) for a 1 Day Advance in Spring Snowmelt at the 1999-Burn and Control Sites

Site	2002			2003			2004		
	$\Delta K\uparrow$[a]	ΔRn	ΔH	$\Delta K\uparrow$	ΔRn	ΔH	$\Delta K\uparrow$	ΔRn	ΔH
1999-burn	−85.2/−1.0	100.8/1.1	28.3/0.3	−63.5/−0.7	75.8/0.8	5.3/0.1	−65.4/−0.7	90.2/1.0	28.8/0.4
Control	10.5/0.1	42.5/0.5	25.4/0.3	7.7/0.1	27.5/0.3	11.2/0.1	6.7/0.1	36.5/0.4	19.4/0.2

[a]We calculated presnowmelt and postsnowmelt means for each variable (i.e., K↑, Rn, and H) from the average diurnal cycles using two-week data before and after snowmelt, respectively. $\Delta K\uparrow$, ΔRn, and ΔH were then calculated as differences between the presnowmelt and postsnowmelt means. The number before the sign "/" denotes the differences in the energy fluxes for one day, which was then divided by 90 (days) to obtain the averaged difference for spring (the number after the sign "/") due to an advanced snowmelt of 1 day earlier.

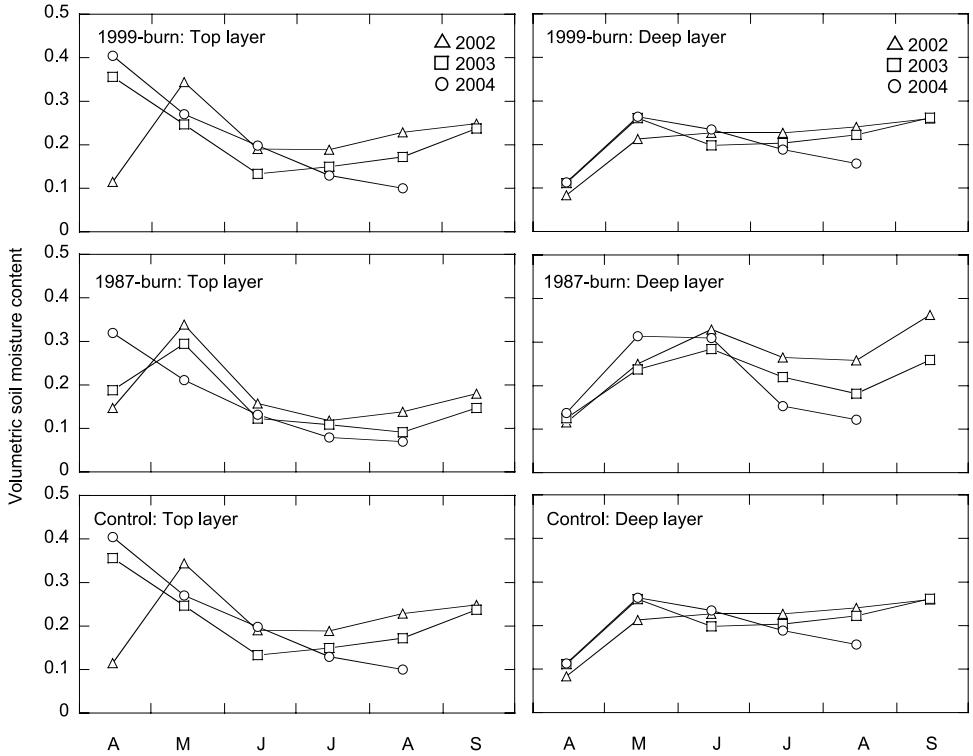

Figure 8. Monthly averaged volumetric soil moisture content in the top layer (left panel; 0–4 cm for the 1999-burn and 1987-burn, 0–5 cm for the control) and deep layer (right panel; 25–40 cm for the 1999-burn, 11–37 cm for the 1987-burn, 22–27 cm for the control) in 2002, 2003, and 2004.

increase methane emissions, and increase rates of decomposition of soil carbon that was previously protected in frozen soils. Subsequent increases in greenhouse gases may amplify warming in a positive feedback loop. Warming-induced drought may further contribute to a positive feedback by reducing midsummer rates of photosynthesis, although these losses may be offset by an earlier onset of growing season (and more carbon uptake during spring) [*McGuire et al.*, 2006]. Changes in the northern carbon cycle will occur simultaneously with changes in species composition and soils that have consequences for surface energy exchange. Feedbacks caused by changes in surface biophysics in northern ecosystems may be as important as, or even more important than, those associated with changes in the biogeochemistry [*Bonan et al.*, 1992, 1995; *Snyder et al.*, 2002; *Bala et al.*, 2007].

[31] High-latitude warming may cause large-scale shifts in species abundance, including increased shrub expansion and abundance [*Sturm et al.*, 2001, 2005], northward tree invasion into tundra [*Esper and Schweingruber*, 2004], and increases in boreal forest leaf area [*Myneni et al.*, 1997; *Nemani et al.*, 2003]. In turn, these vegetation changes may cause more shortwave radiation to be absorbed by the surface during both spring and summer, which is likely to further amplify climate warming [*Eugster et al.*, 2000; *Serreze et al.*, 2000; *Chapin et al.*, 2005; *Euskirchen et al.*, 2007].

[32] Fire may contribute to another set of important high latitude climate-vegetation feedbacks. Burned area has increased over the last several decades and boreal North America [*Kasischke and Turetsky*, 2006] and is projected to increase even more in the future as a result of warming and drying in the continental interior [*Flannigan et al.*, 2005].

[33] More fire increases the relative abundance of early to mid successional deciduous stands, as compared with older evergreen conifer stands, leading to changes in radiative exchange and the surface energy budget at a regional scale [*Chapin et al.*, 2000; *Chambers and Chapin*, 2002]. Our three-year study shows that fire substantially decreases both net radiation and sensible heat fluxes. Changes in biogeochemistry probably only partly cancel the reduced energy flow to the atmosphere at a regional scale, so that boreal fire may be one of the only known negative feedbacks to climate warming at high northern latitudes [*Randerson et al.*, 2006]. A diagram synthesizing fire-induced changes in surface energy exchange as a function of postfire stand age is shown in Figure 9. Fire creates a sudden change in land cover and surface properties, leading to a decrease in net radiation and sensible heat fluxes for the first 4–5 decades (Figure 9). An important question for future research is whether a shift from evergreen conifer to deciduous broadleaf plant functional types in boreal regions contributes to additional cooling via an increase in midsummer cloud formation (and an increase in planetary albedo). Our measurements show that deciduous broadleaf forests have higher evapotranspiration rates during early summer (June and July), and thus the potential to increase atmospheric water vapor content during this time (Figure 6). A concurrent decrease in sensible heat from these forests would decrease

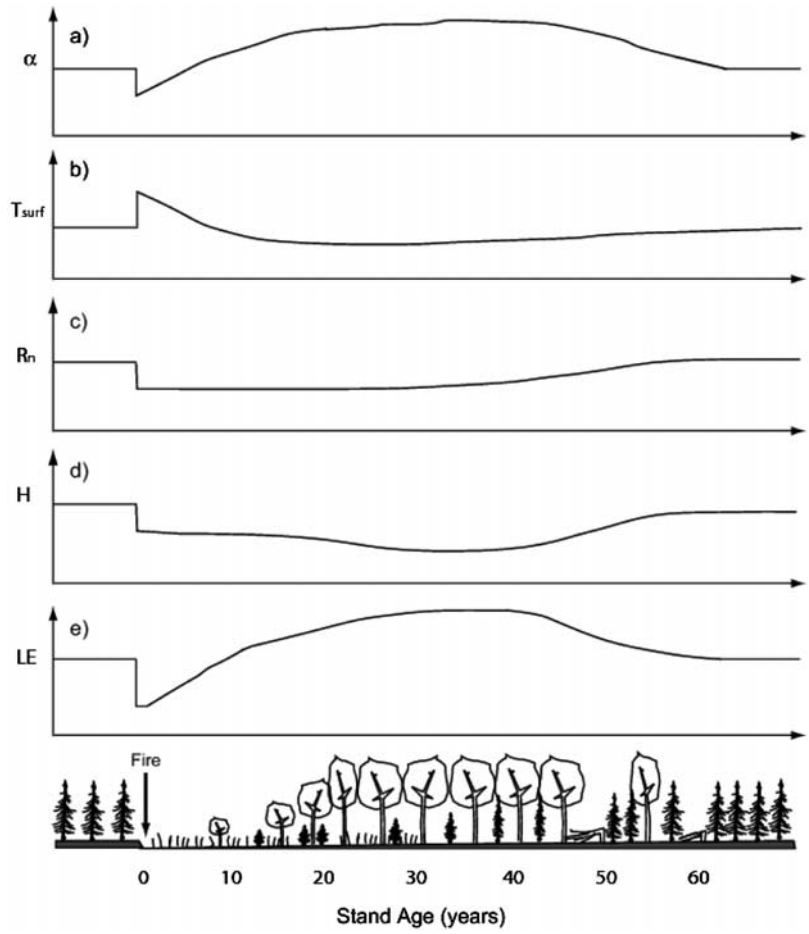

Figure 9. Schematic diagram for postfire changes in (a) summer albedo, (b) summer soil surface temperature, (c) midsummer net radiation, (d) midsummer sensible heat flux, and (e) midsummer latent heat flux. Albedo is low immediately after fire as a consequence of black carbon covering the boles of dead black spruce and soil surfaces. Concurrent decreases in surface roughness causes surface temperatures to increase, outgoing longwave radiation to increase, and net radiation to decrease. Sensible and latent heat fluxes are low immediately after fire as a result of the decrease in net radiation (and thus the available energy to drive these fluxes). During intermediate stages of succession (~20 to 40 year stands in black spruce successional trajectories), increased albedo associated with a deciduous broadleaf tree canopy causes net radiation to remain low as canopy roughness increases. High leaf area, canopy conductance, and transpiration causes more available energy to flow into latent heat, and a consequence, sensible heat fluxes remain low during this stage as compared with prefire levels. With the development of the canopy overstory and increasing abundance and thickness of the moss layer, the soil organic layer thickens and mineral soil layers cool. Forest trajectory drawing adapted from Figure 5–17 of *Kimmins* [2004] and *Hinzman et al.* [2003].

air temperatures in the lower troposphere, making it easier for water vapor to condense.

[34] Multiple lines of evidence indicate that the timing of snowmelt and rates of soil heating have changed in recent decades in Alaska and Canada [*Dye*, 2002; *Smith et al.*, 2004; *Euskirchen et al.*, 2006]. Our work implies that an increase in spring warming (and thus an earlier snow melt) would weaken the albedo-driven cooling effect of boreal forest fire. Specifically, earlier snowmelt leads to enhanced absorption of solar radiation and consequently an increase in sensible heating of the atmosphere for both prefire and postfire stands, with a greater rate of heating for the postfire stands. Consequently, earlier snowmelt reduces the difference in the absorbed solar radiation between postfire and mature conifer stands during spring.

[35] Our work also suggests that the surface energy budget in recently disturbed ecosystems may be more sensitive to larger-scale variations in climate than what may occur in mature evergreen conifer stands. The impact of variable snowmelt for atmospheric heating is larger, for example, in recently disturbed stands because the snowpack is more exposed and consequently leads to larger differences between presnowmelt and postsnowmelt albedo. Energy partitioning in the recently disturbed stands also appears more sensitive to midsummer drought. In response to the severe drought during August of 2004, evapotranspiration at the 1987-burn decreased substantially (by 38%)

and thus showed a greater potential to amplify regional drying than surface energy exchange in the control- where evapotranspiration decreased by only 8%. Taking into account both the spring and summer responses, our results suggest that fire-induced shifts in distribution of stand ages (and plant functional types) may amplify interannual climate variability by means of feedbacks associated with surface energy exchange.

6. Conclusions

[36] We measured components of the surface energy budget during 2002–2004 in three ecosystems that were part of a fire chronosequence in interior Alaska. Annual net radiation decreased by approximately 25% for the 1999-burn and 30% for the 1987-burn, relative to the control, averaged over the three years of our study. Sensible heat decreased by an even larger amount, by approximately 57% and 44% as compared with the control. Our data provide evidence that fire-induced changes in the surface energy budget probably contribute to regional cooling at high latitudes through an increase in surface albedo during spring and summer and a decrease in the Bowen ratio in intermediate-aged stands.

[37] In response to interannual variability in the timing of spring snowmelt and summer drought, energy fluxes changed by larger amount at the two postfire stands than at the control. During years with earlier spring snowmelt more shortwave radiation was absorbed by postfire ecosystems as compared with energy absorption by the control. These results imply that future reductions in snow cover in northern ecosystems may weaken the negative feedbacks caused by fire. An increase in the disturbance regime may also have consequences for the magnitude of interannual climate variability. Surface energy exchange was more sensitive to changes in the timing of spring snowmelt and summer drought at the deciduous broadleaf forest than at the mature conifer forest. Therefore, a shift in plant functional types expected to accompany an increase in boreal forest burned area has the potential to amplify climate variability during spring and summer.

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