Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska

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[1] The impact to the permafrost during and after wildfire was studied using 11 boreal forest fire sites including two controlled burns. Heat transfer by conduction to the permafrost was not significant during fire. Immediately following fire, ground thermal conductivity may increase 10-fold and the surface albedo can decrease by 50% depending on the extent of burning of the surficial organic soil. The thickness of the remaining organic layer strongly affects permafrost degradation and aggradation. If the organic layer thickness was not reduced during the burn, then the active layer (the layer of soil above permafrost that annually freezes and thaws) did not change after the burn in spite of the surface albedo decrease. Any significant disturbance to the surface organic layer will increase heat flow through the active layer into the permafrost. Approximately 3–5 years after severe disturbance and depending on site conditions, the active layer will increase to a thickness that does not completely refreeze the following winter. This results in formation of a talik (an unfrozen layer below the seasonally frozen soil and above the permafrost). A thawed layer (4.15 m thick) was observed at the 1983 burned site. Model studies suggest that if an organic layer of more than 7–12 cm remains following a wildfire then the thermal impact to the permafrost will be minimal in the boreal forests of Interior Alaska. *INDEX TERMS:* 1823 Hydrology: Frozen ground; 0614 Electromagnetics: Biological effects; 1866 Hydrology: Soil moisture; *KEYWORDS:* forest fire, boreal forest, permafrost, frozen ground, ground temperature

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1. Introduction

[2] The discontinuous permafrost zone is the one of the most sensitive areas to climate warming in the world. Throughout the circumpolar north, the boreal forest widely overlaps the area of discontinuous permafrost [*Péwé*, 1975; *Brown et al.*, 1997; *Osterkamp et al.*, 2000]. The thermal condition of permafrost in this region is quite unstable, as it is very close to thawing, often $-1°C$ or warmer. The distribution of the permafrost is strongly influenced by local factors such as landscape, soil type, and vegetation cover [*Viereck*, 1982; *Haugen et al.*, 1982]. The presence and thickness of the surface organic layer are the most important factors controlling degradation or aggradation of permafrost [*Viereck*, 1982]. When the organic layer is removed, the surface albedo decreases and the soil thermal conductivity of the surface soil layer increases from about 0.2 to about 1.0 W/m K [*Hinzman et al.*, 1991]. Wildfire is one of the most important agents controlling the thickness of organic layer in the boreal forest.

[3] Wildfires have been a natural part of the boreal forest ecosystems, burning an average of 1 million ha/yr in 1950 increasing to almost 3 million ha/yr in 2000 in the North American boreal forests (B. Stocks et al., The changing fire regime of Western North America submitted to *Journal of Geophysical Research*, 2002) [*Kasischke and Stocks*, 2000]. Fires in boreal forests have both immediate and long-term impacts on the ecosystem due to their effects on surface energy, water balance, and underlying permafrost. The return period for wildfire in the boreal forest is about 29–300 years [*Yarie*, 1981; *Kasischke et al.*, 2000; *Dyrness et al.*, 1986] and is strongly influenced by climate and human activities (both as a source of ignition and as an agent of fire control) [*Burn*, 1998a; *Brown and Grave*, 1979; *Fastie and Mann*, 1993].

[4] Short-term studies of the effects of fire on the soil moisture and ground thermal regime in more temperate regions as well as cold regions have been well documented. Soil moisture content increases immediately following fires due to a decrease in evapotranspiration [*Tiedemann et al.*, 1979; *Klock and Helvey*, 1976; *Moore and Keeley*, 2000]. *Klock and Helvey* [1976] also noted a decrease in soil moisture content to prefire levels after a 5 year period. However, *Liang et al.* [1991] reported lower soil moisture contents in a burned area 2 years after a fire. Numerous

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other studies show increasing soil moisture contents following logging operations [*Klock and Helvey*, 1976; *Croft and Monninger*, 1953; *Zierner*, 1964].

[5] Although the net surface energy balance may not change significantly after fire, the ways in which the incoming energy is partitioned does change substantially. For example, following wildfire, the surface albedo is significantly reduced. The reduced albedo means that soils can absorb more incoming shortwave radiation than before the fire, which is then converted to sensible heat, resulting in higher ground surface temperatures. The offset of this is significantly increased outgoing longwave radiation due to these higher ground surface temperatures. The effect of this is an actual decrease in the net radiation by about 10%, but increased energy conducted into the ground [*Rouse and Mills*, 1977].

[6] Changes in depth to permafrost and active layer thickness are well documented and have been observed in numerous experimental studies. Changes in the surface material and thermal properties allow for increased heat flow. *MacKay* [1970] reported an average 24.1 cm (149%) increase in active layer thickness at the end of the first year and 34.8 cm increase (171%) in the second autumn. *Burn* [1998b] documented more rapid thawing and delayed freeze-back of the active layer compared to unburned areas. *Brown* [1983] noted that each of these effects is directly proportional to the fire severity. In cases of more severe fires, the surface organic layer is entirely combusted exposing the mineral soil beneath. Any disturbance to the surface layer will increase heat flow through the active layer into the permafrost. After approximately 3–5 years (depending on site conditions) the active layer will increase to a thickness that does not completely refreeze the following winter forming a talik. *Viereck* [1982] reported there was no significant difference in the active layer between the burned site and unburned sites in the first summer following the 1971 Wickersham Dome fire, located near Fairbanks, Alaska. However, the time required for the active layer to become completely frozen was delayed by 1 month compared to the unburned site in the first winter after the fire (12 December to 15 January 1971). In the following summer, the active layer was 161% deeper in the burned site as compared to the unburned control area. *Viereck* [1982] also demonstrated that fire lines (mechanically removing vegetation with a bulldozer) had more impact to the active layer than the severely burned site. *Heginbottom* [1971] reported that by the summer after a fire in Northwest Canada, thaw was 9 cm deeper on a burned site as compared to a control. *Lotspeich et al.* [1970] reported that the 1966 Dennison River fire, Eastern Alaska, showed no significant difference in thaw depth between burned and unburned sites. The burn was not severe enough at any site to remove the organic layer, which remained about 20 cm thick at study burned site. *Wein* [1971] reported the conditions of site burned in 1969 in Interior Alaska. He described the active layer as being 35–50% deeper the next spring following the fire, and 15–20% greater in fall. *Brown et al.* [1969] observed several fire sites along the Alaskan Taylor Highway. The thaw depth was increased 140–160%. *Kryuchkov* [1968] observed the active layer was actually thinner several years after a fire in case of Siberian tundra. Before the fire, the active layer was 50–70 cm deep however it was only 40–45 cm a few years after the fire. This decrease in summer thawing was due to higher soil moisture (ice) contents resulting from decreased transpiration. Controlled field experiments achieved more uniform results as compared with observations following wildfires. *Dyrness* [1982] studied differences in active layer thickness under the different ground surface treatments (control, lightly burned, heavily burned, half the forest floor organic layer removed, and entire organic layer removed). After the fourth summer, the only treatment that had a statistically significant effect on soil temperature and permafrost depth was the mechanical removal of all or a portion of the forest floor. *Esch* [1982] tested the impact on permafrost of removing vegetation for road construction in Alaska. Three years after treatment he observed the removal of vegetation had increased the depth to permafrost by 300–600%.

[7] Wildfires also impact geomorphological features on small and larger scales. In the first and second years after a fire, mass wasting or landslides frequently occur on hillslopes, being more prone to failure due to increased soil moisture content [*Brown and Grave*, 1979; *Brown*, 1983; *Tiedemann et al.*, 1979]. Several decades later, thermokarst formation may occur as a result of thawing of ice-rich permafrost [*Brown*, 1983; *Viereck*, 1973].

[8] This investigation focused on postfire impacts to the permafrost to quantify the impacts of fire on (1) direct heat conduction and convection to the ground, (2) removing moss as an insulating material, (3) heat budget, (4) soil moisture characteristics, and (5) active layer thickness and talik formation. These agents are the major components of permafrost dynamics. The evaluation of these agents will provide a better understanding of the relationship between burn severity and permafrost response and the effects on the hydrologic regime following wildfire.

2. Methods

[9] Eleven wildfire sites throughout Interior Alaska, with dates of ignition ranging from 1924 to 2000, were selected for this study (Figure 1). At each of these sites, active layer depth, ground temperature, soil moisture content, thermal conductivity and air temperature were measured. Several historical fires are located near Caribou-Poker Creeks Research Watershed (CPCRW) 48 km of north of the Fairbanks, Alaska [*Fastie*, 2000; *Fastie and Mann*, 1993]. Four of these wildfire burn areas within or immediately adjacent to CPCRW were selected for study and occurred in 1999 (site 1, control, moderate, severe), 1996 (site 2, moderate, control), 1990 (site 3, moderate, control), and 1924 (site 4, severe, moderate). The July 1999 controlled burn experiment (FROSTFIRE) took place in a 9.7 km^2 subbasin underlain by discontinuous permafrost [*Hinzman*, 2000]. At the 1996, 1990, and 1924 burn areas, two small pits were dug: one in a burned area and one control. In the 1924 site, one pit was located inside the severely burned area and the other inside the moderately burned area. Locations of the pits were selected to be as physically similar to each other as possible. The total depth of the pits varied between 60 and 80 cm. In each pit, the temperature and soil moisture content were measured. TDR probe and gravimetric samples were collected in the surface duff layer, in the organic soil, and at regular intervals through the

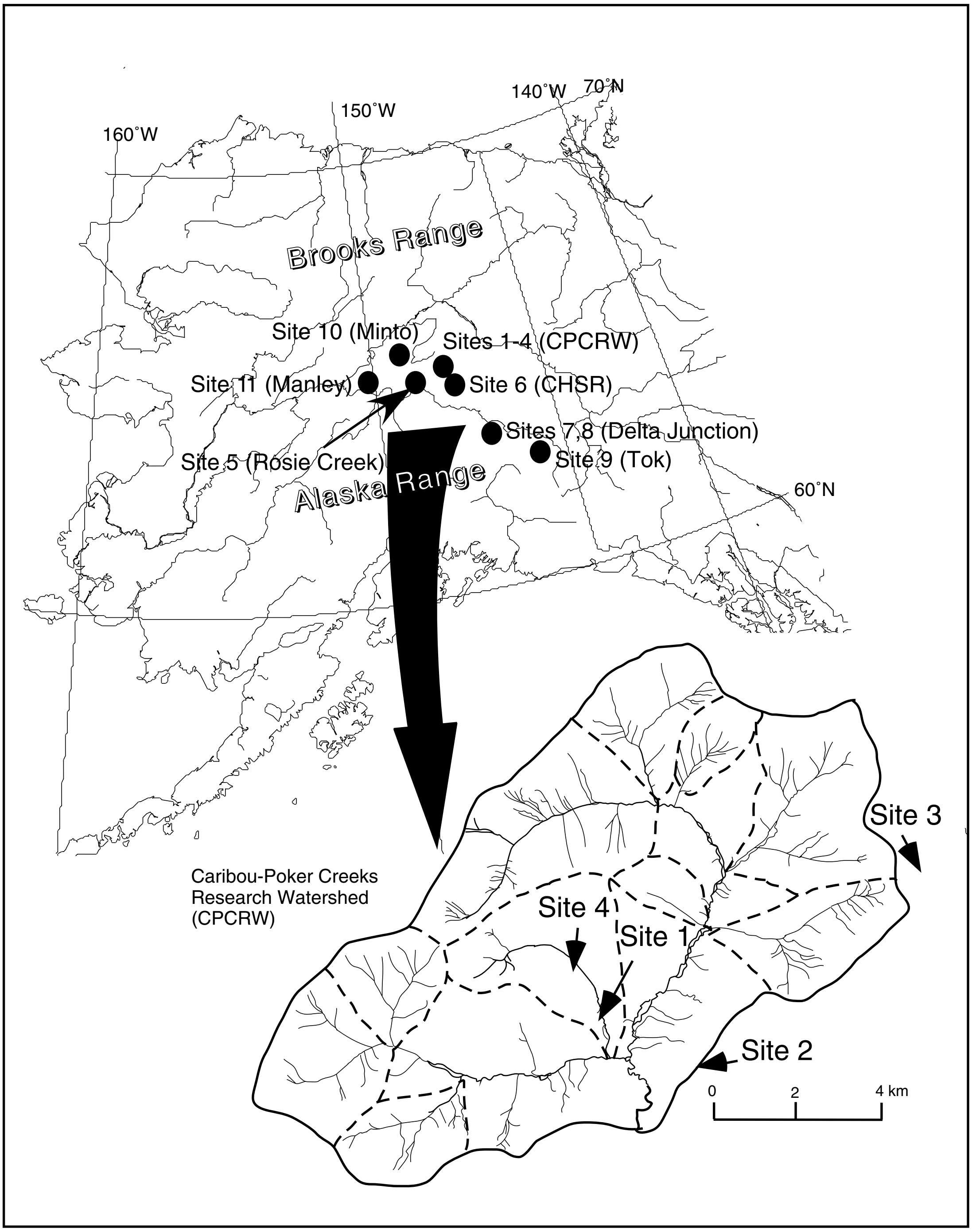

Figure 1. Map of study sites. Eleven sites are located in Interior Alaska between the Alaska Range and the Brooks Range. Four of the 11 sites were located near CPCRW (insert map).

mineral soil. A ground temperature transect was established between the burned and unburned sites. Shallow surface (15–20 cm depth) temperature measurements were collected at 3 m intervals using a thermocouple. A series of thermistors were used to measure the (0–100 cm depth) ground profile temperatures along the transect. Measurements were collected at 10–20 m intervals with additional measurements made near the boundary of the burned and unburned areas. Measurements were made every other month at the 1996 site and in August 1998 at the 1990 and 1926 sites. Seasonal frost depths were also measured. At the 1999 sites (site 1, FROSTFIRE), active layer depth, ground temperature, thermal conductivity, soil moisture content, and meteorological observations were collected.

[10] The Rosie Creek site (Bonanza Creek LTER) was located in a burn that occurred in 1983 (severe, control) located approximately 10 km southwest of Fairbanks, Alaska. The Chena Hot Springs Road site was located at a controlled burn that was conducted in August 2000, approximately 25 km east of Fairbanks, Alaska. This con-

Table 1. Summary of Site Information and Instrumentation (K: Type K Thermocouple, T: Type T Thermocouple, Th: Thermistor (Alpha Thermistor 14A5001C2), Pt: 100 Ohm Platinum Resistor, CS: Campbell Scientific CS615 Probe, Vt: Hydra Soil Moisture Capacitance Probe by Vitel Inc., MRC: MRC Thermistor Ground Temperature Probe, TDR: Time Domain Refractometry)

Site	Location	Latitude	Longitude	Year of ignition	Ground temperature	Soil moisture	Active layer	Profile method	Meteorological observation
Site 1	CPCRW	65.1°N	147.3°W	1999	K, Th, Pt	CS, Vt	Thaw probe, MRC	trench, borehole (26 m)	10 m tower
Site 2	CPCRW	65.1°N	147.3°W	1996	K, Th	Vt	Thaw probe, MRC	trench	no
Site 3	CPCRW	65.1°N	147.2°W	1990	K	Vt	Thaw probe	trench	no
Site 4	CPCRW	65.1°N	147.3°W	1924	K, Th	Vt	Thaw probe	trench	no
Site 5	Rosie Creek (LTER)	64.7°N	148.1°W	1983	K, Th	CS, Vt	Thaw probe, MRC	trench, borehole (18 m)	10 m tower
Site 6	25 km east of Fairbanks	64.8°N	147.1°W	2000	K, T, Th	CS, Vt, TDR	Thaw probe, T	trench	no
Site 7	Donnely Flats	64.0°N	145.7°W	1994	K, Th	CS, Vt	Thaw probe, th	trench	no
Site 8	Hajdukovich Creek	64.0°N	145.7°W	1999	K, Th	CS, Vt	Thaw probe, th	trench	10 m tower
Site 9	Tok	63.3°N	143.0°W	1990	K, Th	CS, Vt	Thaw probe	trench	no
Site 10	Minto	65.2°N	149.3°W	1999	K	CS, Vt	Thaw probe	trench	no
Site 11	Manley	65.0°N	150.6°W	1963	K	CS, Vt	Thaw probe	trench	no

trolled burn lasted about 3 hours, during which direct measurements were collected. The purpose of this experiment was to determine the thermal impact from direct heating by the fire. A black spruce stand, 16 m^2 × 5 m high, with woody debris on the surface was ignited on a surface of feathermoss (*Hylocomium* spp.) and *Sphagnum*. Soil moisture contents and temperatures were measured below the surface to a depth of 32 cm in 2 cm increments. Type T and type K thermocouples were installed in the organic layer. Type K thermocouples with silfa silica insulation were used at the surface and in the moss (8 cm). The deeper layers were monitored using copper-constantan (type T) thermocouples with regular polyvinyl chloride (PVC) insulation. Temperature measurements were collected every minute. Soil moisture content was measured using balanced TDR (time domain reflectometery) probes and a Tektronix model 1502B cable tester on 10 min intervals.

[11] Two sites (sites 7 and 8) were located near Delta Junction, Alaska. The 1994 fire at Hajdukovich Creek (severe, control) and the 1999 fire at Donnely Flats (severe, moderate, control). Both fires had a variety of severities in black spruce (*Picea mariana*) with a feathermoss forest floor. The fire at site 9 occurred near Tok, Alaska in 1990 and was classified as a moderate burn with some severe areas. The final two sites are located north of Fairbanks, at Minto (site 10, 1999, moderate, control) and at Manley (site 11, 1963, severe, moderate).

[12] At all sites, the remaining moss thickness, soil moisture content, ground temperature and thermal conductivity were measured (Table 2). Soil moisture contents were measured using the Hydra soil moisture capacitance probe by Vitel Inc. [*Atkins et al.*, 1998]. The Campbell Scientific CS615 probe [*Bilskie*, 1997] was used to detect changing dielectric permittivity (constant). Electronic measurements were verified by gravimetric sampling and oven drying for 72 hours at 65°C. Classification of organic soils and degree of humidification was based on the works of the *Soil Classification Working Group Staff* [1998] and of *Pritchett and Fisher* [1987].

[13] Ground temperatures were measured using three different methods. Routine monitoring of active layer temperatures was accomplished using NTC thermistors (Alpha thermistor 14A5001C2) and recorded on Campbell Scientific CR10X data loggers. Type K and T thermocouples with CR10X loggers were used to measure near surface temperatures directly during fire. Three shallow (20–30 m) boreholes were installed at the CPCRW 1999 site (site 1) and Rosie Creek 1983 site (site 5) to periodically monitor permafrost temperature profiles. A precision thermistor was lowered slowly down the boreholes to accurately (<0.001°C error) record permafrost temperature changes and geothermal gradients.

[14] Thermal conductivity was measured using the heat probe method [*Shiozawa and Campbell*, 1990] made by Kona systems and Thermal Logic. Thermal conductivity was also measured in the laboratory as a function of the water content and temperature on undisturbed samples from all sites.

[15] Incoming and reflected shortwave radiation and incoming and emitted longwave radiation were measured using a pair of calibrated pyranometers (Eppley Black and White Pyranometer Model 8-48) and a calibrated pair of pygeometers (Eppley Precision Infrared Radiometer Model PIR), respectively. An independent estimation of net radiation was obtained using a calibrated Frischen type net radiometer (REBS Model Q7.1). A wind speed (Met One 014) dependent dome cooling correction was applied to the results. Table 1 provides a summary of measurements and instruments for each site.

[16] The radiation budget before and after the CPCRW controlled burn was measured using shortwave, longwave and net radiometers. Sensors measuring shortwave and longwave incoming and reflected (emitted) were installed on a 3 m tower (unburned site) and a 1.5 m tower (burned site), both were higher than the black spruce canopy. The radiation balance (Q) is expressed in terms of incoming solar radiation K↓, surface albedo α, incoming longwave radiation L↓ and emitted longwave radiation L↑ in the form:

$$Q = \mathbf{K}\downarrow(\mathbf{1}-\boldsymbol{\alpha}) + \mathbf{L}\downarrow - \mathbf{L}\uparrow$$

[17] Talik formation was simulated using data from Rosie Creek (site 5). Ground temperature observations and other

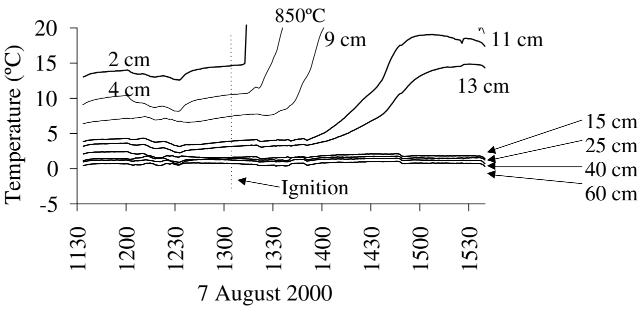

Figure 2. Ground and surface temperatures during the fire at 25 km Chena Hot Spring Road (site 6) showing the increase in the near surface temperatures shortly after ignition. The ground temperature at this site rose only at the shallow depths (<15 cm). Most of the heat from the wildfire transfers to the ground by conduction, which does not penetrate deeply.

meteorological, vegetation and soils data were obtained from the data archive of the Bonanza Creek Long-Term Ecological Research (LTER) (http://www.lter.alaska.edu/). The thermal models used for these studies were described by *Osterkamp and Romanovsky* [1996] and *Romanovsky et al.* [1997].

3. Results

[18] Five categories of the impacts to the permafrost were evaluated: (1) direct fire affects, (2) removing (burned) moss as an insulating material, (3) heat budget, (4) soil moisture characteristics, and (5) active layer thickness and talik formation. All of these impacts are interrelated and cannot be absolutely separated; however, the following is a discussion of the separate mechanisms and the factors that influence them.

3.1. Impact 1: Direct Fire Affects

[19] Figure 2 shows the ground temperature at several depths during the fire at 25 km Chena Hot Springs Road (site 6). The increase in temperature at the forest floor during the fire is quite rapid. The temperature of the forest floor (surface of the moss itself) was more than 800°C during this experimental burn by using type K thermocouple. Temperature started to rise 2 cm below the surface about 10 min after the ignition. No significant increase in temperature was recorded below 15 cm at the feathermoss (*Hylocomium* spp.) and *Sphagnum* sites.

[20] Many natural wildfires occur in forests where the floor is dominated by feathermoss (*Hylocomium* spp.) while fires in areas dominated by *Sphagnum* moss are relatively rare. The differences are due primarily to the characteristic moisture conditions both species prefer. Feathermoss (*Hylocomium* spp.) has a much lower moisture field capacity (about 20%) (C. M. Mack, Earth System Science, University of California, Irvine, personal communication, 2000) and the typical water content is around 10% by volume in summer months, as compared to *Sphagnum* (40% moisture field capacity and typical levels of 30–40% by volume in summer months). The type of moss is not only an important factor in characterizing burn potential and severity, but also affects the surface thermal properties. About 70% of the volume of the live feathermoss is air and 10% is solid; consequently the thermal conductivity is strongly controlled by the water content. Increasing the water content by 40% increases the thermal conductivity by 10-fold (Figure 3). However, the typical condition of the moisture content of feathermoss is somewhat dry (usually less than 15% by volume). The thermal conductivity of the dry moss is almost the same as air (0.02 W/m K). As a result, under these conditions, fires do not transfer an immediate thermal impact to deeper soil layers during a burn.

3.2. Impact 2: Removing (Burned) Moss as an Insulating Material

[21] Removal of the organic layer exerts a serious impact to the ground thermal regime. In the first year following a fire, *Viereck* [1982] reported the thawing index of the soil surface may reach 1900 days°C in severely burned sites, as compared to 1500 days°C in unburned sites. The Mean Annual Surface Temperature (MAST) can increase by 2–3°C. The thawing index at 10 cm depth was 847 days°C (unburned site), 940 days°C (fire line) and 722 days°C (burned site). These suggest that the burn site is slightly cooler at the ground surface during the summer. However, because of higher thermal conductivity of the surface soils of the burned site, the thawing index will be larger throughout the soil profile at the burned site as compared to similar depths in an undisturbed site.

[22] At nearly every site we monitored, the summer ground temperatures of the burned areas were substantially warmer (by 1–20°C) than the adjacent complementary unburned control sites. The only exception to this trend was site 1 in the 1999 FROSTFIRE burn where the thermistors in the burned site were more shaded from the Sun as compared to the control site (Figure 4a). In cases of fires that burned within the last 10 years, the ground surface temperature was warmer at the burned sites as compared to control sites during the early freezing period in autumn. This difference was maintained until the active layer became completely frozen. The moisture content of recent fires was higher in the burned sites, probably due to lower

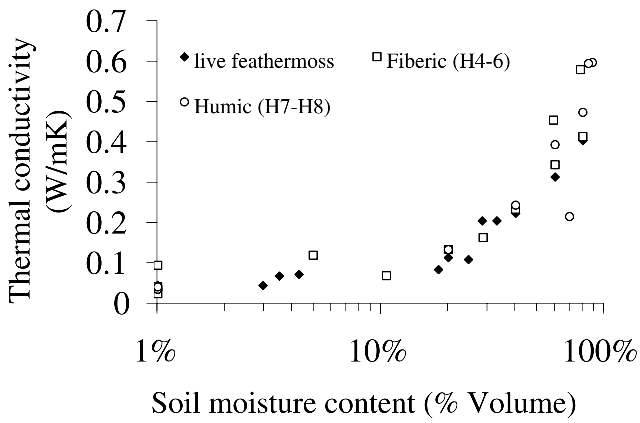

Figure 3. The thermal conductivity of live feathermoss, fibric, and humic layers are strongly affected by moisture content. The classification of the organic layer was based on the works of the *Soil Classification Working Group Staff* [1998] and of *Pritchett and Fisher* [1987].

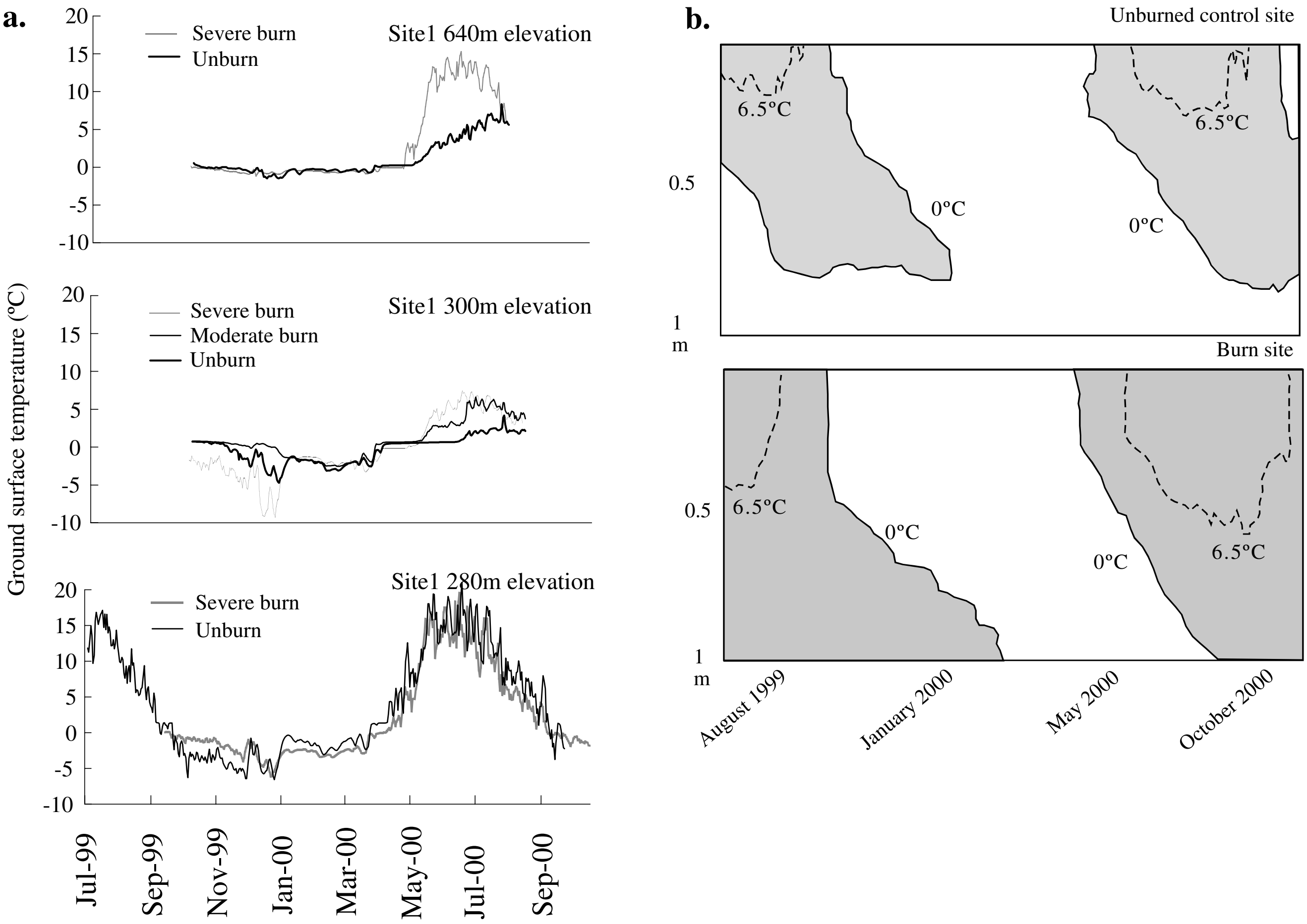

Figure 4. Ground temperature dynamics at the FROSTFIRE site 1. (a) Annual variation of the near surface temperature in three separate locations showing warmer surface temperatures and delayed freezing in the burned sites. (b) Development of active layer in the unburned site is limited to 80 cm while the active layer in the burned site exceeds 1 m.

transpiration rates, so freezing would take longer due to larger latent heat of fusion requirements. However, in the older fires, the soils of the burned sites were somewhat drier and they would freeze faster in the fall. The depth of snow was not significantly different within sites (except in the Delta Junction area, sites 7 and 8) and is not considered to be a controlling factor for these differences. Sites where the burn severity was classified as moderate or severe also demonstrated an increased thickness in the active layer (Figure 4b).

[23] The most important factor controlling the active layer thickness is the thermal conductivity of the organic layer. The active layer thickness is not substantially impacted when the thickness of the organic layer is not significantly reduced during burning, even though the surface albedo is lowered. Table 2 shows field observations of the thermal conductivity of each organic layer. In general, thermal conductivity is mainly a function of density, moisture content and temperature; water content was particularly important in the postfire environment. The type of moss and extent of decomposition of the organic material are important factors in controlling the soil thermal regime (Table 2). A surficial layer of feathermoss has a lower thermal conductivity than *Sphagnum*, but after burning, feathermoss has a fibric layer that has a higher (or equal) thermal conductivity than *Sphagnum*. The thermal conductivity of the surface layer of *Sphagnum* does not change appreciably with burning, thus feathermoss has potentially larger thermal impacts after burning.

[24] The thickness of the surface organic layer remaining after the fire and the resulting active layer were measured in CPCRW and Delta Junction areas. The active layer thickness is strongly controlled by the moss thickness, its thermal conductivity and moisture content and can be predicted based upon a relatively simple relationship. Sites that were severely burned (all or nearly all organic layer burned) always had deeper active layers than their control or adjacent lightly burned sites. A wildfire did not always affect the underlying permafrost. An organic layer with a thickness of 10 cm provided adequate thermal resistance to protect the frozen mineral soil.

3.3. Impact 3: Heat Budget

[25] After the prescribed burn at the FROSTFIRE site 1, the total net radiation decreased compared to an unburned site (97.7 at a moderate burn site compared to 160.2 MJ/m^2 at the control site between 23 July and 10 August 1999) as a result of the increase in the longwave emission (424–438

Table 2. Physical and Thermal Properties of the Organic Layer at Select Study Site

Site	Year of ignition	Vegetation	Observation date	Density (g/cm^3)	SMC (%vol.)	Thermal conductivity (W/m K)	Ground temperature (°C)	Type of layer	Degree of humification	Dielectric constant (100 MHz)
Site 1 Birch Stand (3 cm)	1999	*Pleurozium schrebrii*	10 July 2000	NA	NA	0.09	15.55	live moss	H1	3.1166
Site 1 Birch Stand (10 cm)	1999		10 July 2000	NA	NA	0.15	7.84	Dead moss, litter	NA	3.5529
Site 1 Birch Stand (15 cm)	1999		10 July 2000	NA	NA	NA	NA	mineral, litter	NA	12.9701
Site 1 Open Spruce (3 cm)	1999	*Sphagnum,*	10 July 2000	NA	NA	0.10	13.73	live moss	H1	3.4405
Site 1 Open Spruce (12 cm)	1999	*Pleurozium schrebrii*	10 July 2000	0.08	36.20	0.16	5.78	Fibric	H2	2.7769
Site 1 Open Spruce (20 cm)	1999		10 July 2000	NA	NA	0.24	1.12	Humic	NA	28.4742
Site 5 UB (3 cm)	>1900	*Sphagnum*	7 July 2000	0.05	2.96	0.03	21.98	live moss	H1	2.1634
Site 5 UB (10 cm)	>1900	*Pleurozium schrebrii*	7 July 2000	0.05	10.56	0.06	7.72	Dead moss		4.7433
Site 5 UB (20 cm)	>1900		7 July 2000	0.06	19.47	0.41	0.96	Humic		11.9779
Site 5 B (3 cm)	1983	*Hylocomium splendens*	7 July 2000	0.05	4.96	0.11	19.53	Dead moss	H1	2.253
Site 5 B (8 cm)	1983		7 July 2000	0.36	41.96	NA	NA	Humic	NA	16.7177
Site 5 B (13 cm)	1983		7 July 2000	0.67	60.16	0.91	19.56	Mineral, charcoal	NA	36.4338
Site 7 S (3 cm)	1994	*Ceratodon purpureum*	20 June 2000	0.79	30.58	0.49	15.0	live moss, charcoal	NA	15.94
Site 7 UB (3 cm)	>1860	*Hylocomium splendens*	20 June 2000	0.09	3.27	0.05	13.3	Dead moss	H1	NA
Site 7 UB (12–18 cm)	>1860		20 June 2000	0.09	3.67	0.03		Fibric	H3	2.78
Site 7 UB (25 cm)	>1860		20 June 2000	0.71		0.20	0.6	Humic	H7	12.60
Site 8 S (3 cm)	1999		20 June 2000	0.88	23.76	0.39	17.8	Charcoal	H3	5.92
Site 8 UB (10 cm)	>1860	*Polytrichium, lichen*	20 June 2000	0.32	24.60	0.10	10.4	live moss	H1	11.68
Site 8 UB (12 cm)	>1860		20 June 2000	1.02	71.00	0.57	1.6	Humic	H6	57.48
Site 8 M (7 cm)	1999		20 June 2000	0.09	3.27	0.04	14.8	Fibric	H2	4.22
Site 8 M (15 cm)	1999		20 June 2000	0.50	26.65	0.15	1.5	Humic	H4	19.20
Site 9_1 (8 cm)	1990	*Polytrichium commune.*	20 June 2000	0.45	28.34	0.13		Fibric	H1	18.92
Site 9_2 (8 cm)	1990	*Polytrichium spp.*	20 June 2000	1.16	35.87	0.99		Humic	H7	16.72
Site 10 B (3 cm)	1999		21 June 2000	0.20	6.54	0.05	22.0	dead moss charcoal	H3	3.13
Site 10 B (15 cm)	1999		21 June 2000	0.86	65.19	0.58	3.1	fibric	H8	55.79
Site 10 UB (5 cm)	1999	*Hylocomium splendens*	21 June 2000	0.04	1.88	0.06	15.3	live moss	H1	2.21
Site 10 UB (15 cm)	1999	*Pleurozium schrebrii*	21 June 2000	0.09	4.50	NA	1.4	fibric	H4	3.34
Site 11 M (3 cm)	1963	*Drepanocladus sp.*	21 June 2000	0.23	16.73	0.22	6.9	live moss	H1	10.11
Site 11 M (15 cm)	1963	*Polytrichium*	21 June 2000	0.44	34.23	0.44	0.3	Humic	H5	41.03
Site 11 S (5 cm)	1963	*Polytrichium*	21 June 2000	0.08	4.29	0.06	6.3	live moss	H1	2.50
Site 11 S (15 cm)	1963		21 June 2000	0.07	3.89	0.11	2.0	Fibric	H2	3.85
Site 11 S (23 cm)	1963		21 June 2000	0.97	58.10	0.58	0.5	Humic	H7	20.67

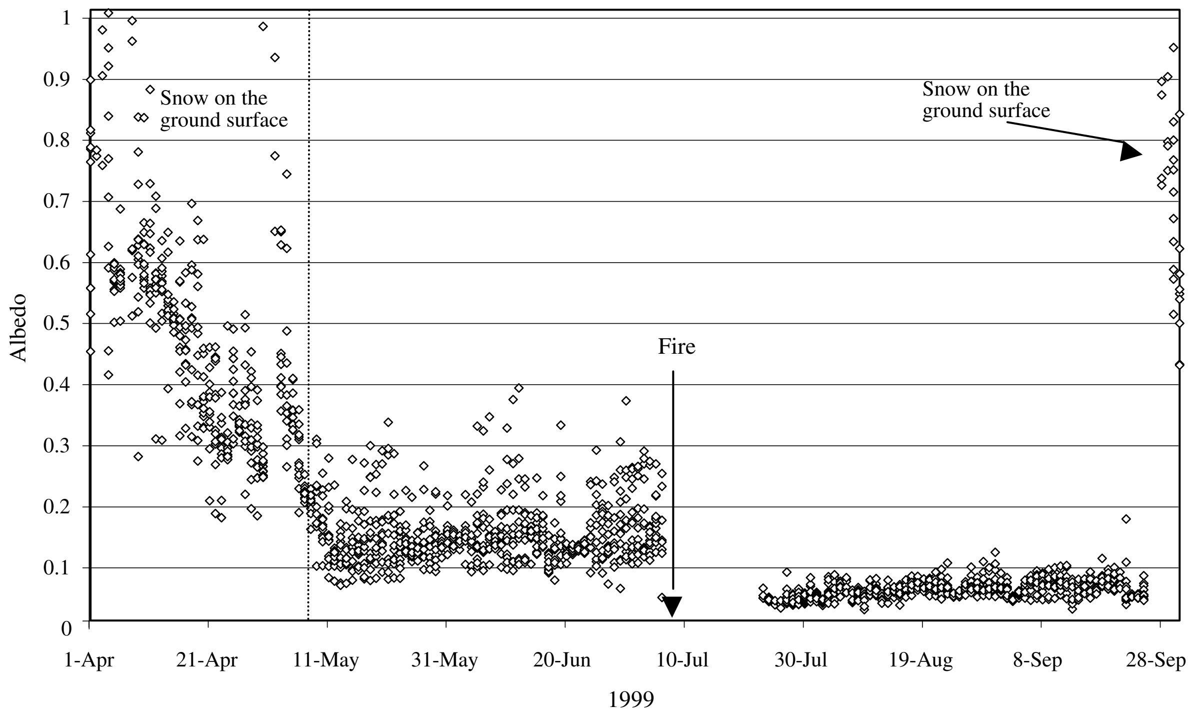

Figure 5. There are strong differences in albedo before and after wildfire. During snowmelt, the albedo ranges from 0.2 to 0.9 or more, decreasing to about 0.14 on a feathermoss surface prior to the fire. The albedo drops to 0.07 at a moderately burned site after the fire. Plotted data are the daytime (0600–1700 AST) averages.

W/m^2). On the other hand, the albedo dramatically decreased from ~0.14 to ~0.05 (Figure 5) yielding greater absorbtion of shortwave radiation at the ground surface. The daily variation of the net radiation (Q) was more variable at the unburned site (max 537 W/m^2, min −48.7 W/m^2) than the burn site (max 458 W/m^2, min −37.6 W/m^2). The radiation balance components are given in Table 3 in term of 100 units of incoming shortwave.

[26] Radiation efficiency (Re) was calculated as the ratio of incoming shortwave radiation to net radiation, Re = Q/K↓. The radiation efficiency of the site 1 (moderately burned) was 50% (at noon) while the control unburned site had a Re of 75%. This indicates that the burned site was 25% less efficient at retaining incoming radiation. *Chambers and Chapin* [1999] reported somewhat similar results of radiation efficiency being 75% of efficiency at moderate burn site and 78% at an unburned site. The energy absorbed at the ground surface divides into sensible, latent, ground heat fluxes and energy storage. The daytime sensible heat flux is large and comprises at least 20% of K↓ (38% according to *Chambers and Chapin* [1999]). The large sensible heat flux from the fresh burn manifests itself visibly on a sunny day [*Rouse and Mills*, 1977]. The Bowen ratio of a freshly burned site in the FROSTFIRE study was reported at $\beta = 4.5 \pm 3.3$ [*Chambers and Chapin*, 1999]. Using this value of β the ground heat flux and energy storage at our site 1 increased from an unburned area (0.07Q: 18.5 W/m^2) at the burned site (0.16Q: 29 W/m^2 or more). The ratio of ground heat flux to net radiation increased following fire. The distribution of the energy balance components shifted before and after fire. The increasing sensible heat and ground heat flux was balanced by decreasing the latent heat flux. As the result, ground temperatures were increased and wetter conditions became established despite the total net radiation decrease.

3.4. Impact 4: Soil Moisture Characteristics

[27] During burning, the soil moisture content of the soils becomes drier than an adjacent unburned area. Burned soils can develop a near surface hydrophobic layer [*DeBano*, 2000] that can resist surface water infiltration. Precipitation occurred during the 2 days following the CHSR Fire (site 6). In the control area, a significant increase in the soil moisture content was observed; while in the burn area, little

Table 3. Radiation Balance of Burned and Control Areas at Site 1 (Net Radiation Q, Incoming Solar Radiation K↓, Reflecting Shortwave Radiation K↑, Surface Albedo (%) α, Incoming Longwave Radiation L↓ and Emitting Longwave Radiation L↑, Radiation Efficiency Re, Total Net Radiation ΣQ)

Radiation	K↓	K↑	α (%)	L↓	L↑	Re	Q	ΣQ (MJ/μ^2)	Max. (W/m^2)	Min. (W/m^2)
Burned (site 1)	100	4.6	0.05	105	122	0.47	47.43	97.7	458	−37.6
Unburned (site 1)	100	13.5	0.14	105	118	0.74	73.59	160.2	537	−48.7

From 23 July to 10 August 1999 (sensor height 1 m at burned, 1.5 m unburned).

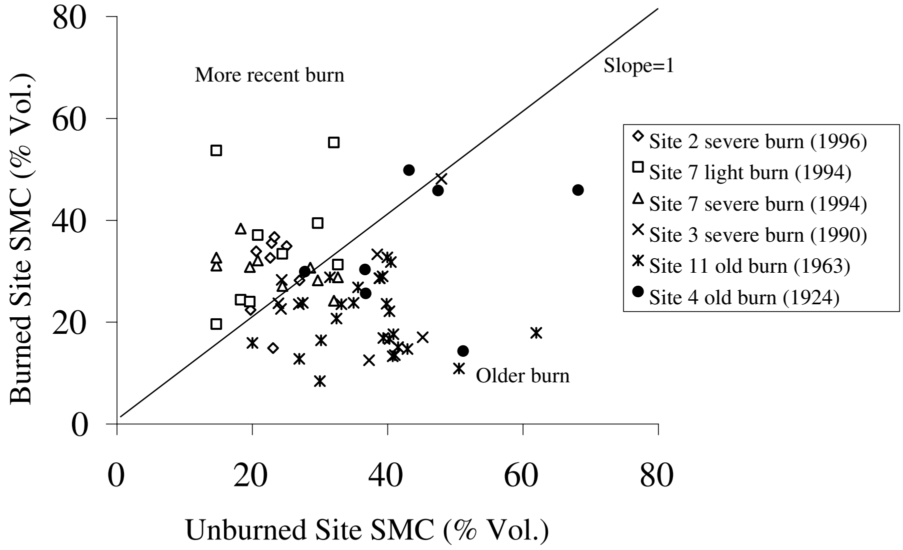

Figure 6. Comparison of the soil moisture contents (SMC) between burned areas and adjacent unburned control areas at fires of various ages (sites 2, 3, 4, 7, and 11) shows that in general more recent fires have higher mineral soil moisture contents in burned areas while older fires have higher mineral soil moisture contents in unburned areas.

change was observed. At the end of the observation period, the apparent hydrophobic layer was decreasing and the soil moisture content in the burned soils began to increase. It is unclear if the hydrophobicity effect is very important in the Alaskan soils that are usually somewhat moist to wet.

[28] Prior to the CPCRW fire (site 1), the soil moisture contents at two adjacent sites were nearly the same. However, following the fire, the burned area displayed a noticeable increase in soil moisture content compared to the unburned control area. This is due to a lower transpiration rate in the burned area. However, all of the older fire sites (>10 years) tend to display drier soils as compared to the unburned (or lightly burned) control sites. At several sites, the soil moisture content was measured in burned areas and adjacent unburned areas at the same depths. Figure 6 shows that the burned sites were always wetter at corresponding depths (active layer 0–50 cm) in the 1996 and 1994 burns. On the other hand, older burn sites were always drier at any depth. At the 1990 burn (site 3) the moisture content was about the same in the burned and unburned sites. The results of this study indicate the time required for severely burned soils to transition from having higher soil moisture contents to lower soil moisture contents to be about one decade.

3.5. Impact 5: Active Layer Thickness and Talik Formation

[29] Both increased soil moisture contents as well as soil temperatures affect the active layer at the severe burn sites. Figure 4b shows the active layer temperature profile for two thaw cycles in the FROSTFIRE severe burn area. There appears to be clear evidence of a change due to the fire in the first summer. The active layer was greater in locations where most of the organic material was burned to mineral soil. At the 1999 Minto fire (site 10), 50 random active layer depths were measured in both the moderate burned area and unburned areas. The average active layer thickness was virtually identical (29.9 cm burned area versus 29.2 cm unburned area on 21 June 2000). The permafrost has degraded to a depth 4.15 m from the surface at the severe burned area of Rosie Creek fire (site 5) since 1983. The maximum annual depth of seasonally frozen ground is about 2 m, so the layer of unfrozen soil between the seasonally frozen ground and the permafrost is a talik. The thawing of the permafrost seems to have slowed or ceased as evidenced by the year 2000 ground temperature profile being nearly identical (below 4 m) to the year 1996 ground temperature profile (Figure 7). There are many permafrost-free areas (or deep taliks) in areas where one would expect thick permafrost. Some of these permafrost-free anomalies correspond with fire scars indicating previous disturbances impart long lasting impacts to the permafrost. In light of the recent climatic conditions, we hypothesize that it will be difficult for permafrost to recover once thawed following a severe fire in this area.

4. Discussion

[30] The permafrost distribution and ground surface heat balance are closely related. It is possible to determine a simple heat balance based upon the freezing index (I_f) and thawing index (I_t) at the ground surface. The freezing and thawing indices are calculated based upon an accumulation of freezing or thawing degree days throughout the positive and negative daily mean temperature periods. $I_t = \Sigma t_s$ when ($t_s > 0$) and $I_f = \Sigma t_s$ when ($t_s < 0$) where t_s is the mean daily ground surface temperature. I_f and I_t were calculated at each burn and control site based on hourly data. The boundary conditions of permafrost presence is calculated using the following formula [*Romanovsky and Osterkamp*, 1995; *Lunardini*, 1981]:

$$I_t \leq (\lambda_f/\lambda_t) I_f$$

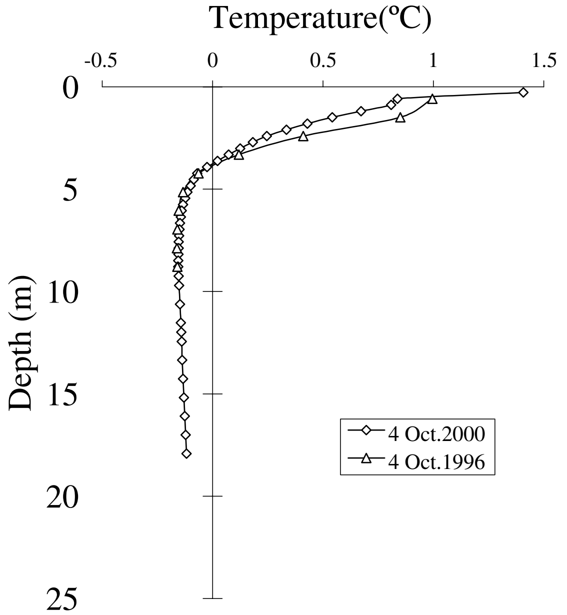

Figure 7. Ground temperature profile at site 5 borehole. The permafrost starts 4.15 m below the surface (4 October 2000). The permafrost is very warm and close to thawing with the coldest temperature only −0.16°C at 8.6 m.

where λ_f and λ_t are frozen and thawed thermal conductivities. The ground surface I_f (freezing degree days) ranges between 500 (on south facing slopes) and 3000 (at the base of the north facing slopes) in CPCRW. On the other hand, I_t (thawing degree days) does not vary greatly (Ca. 1400) compared to the freezing degree days [*Yoshikawa et al.*, 1998]. Thus I_f is a key factor in determining the presence or absence of permafrost. *Van Cleve and Viereck* [1983] suggest that I_t measured at 10 cm ground depth is a good indicator of permafrost presence. On a black spruce forest floor, I_t averages about 500–800 °C days. Less than 5 years after a fire, I_t increased to 1000–1250. If the organic layer and original trees recover, the active layer may return to its original thickness within 25–50 years [*Van Cleve and Viereck*, 1983].

[31] Wildfire affects frozen ground systems primarily through the removal of vegetation and the surface organic layer. The loss of vegetation increases the soil moisture content due to reduced evapotranspiration. The thermal conductivity of an organic layer is a function of the moisture content, temperature, and density. If the organic layer remaining after a fire is adequately thick, the active layer thickness will not be depressed, in spite of dramatically lowered albedo. As a result, thickness and type of organic layer are the most important agents influencing the permafrost response to wildfire. Figure 8 shows the relationship between active layer thickness and a function of the thermal

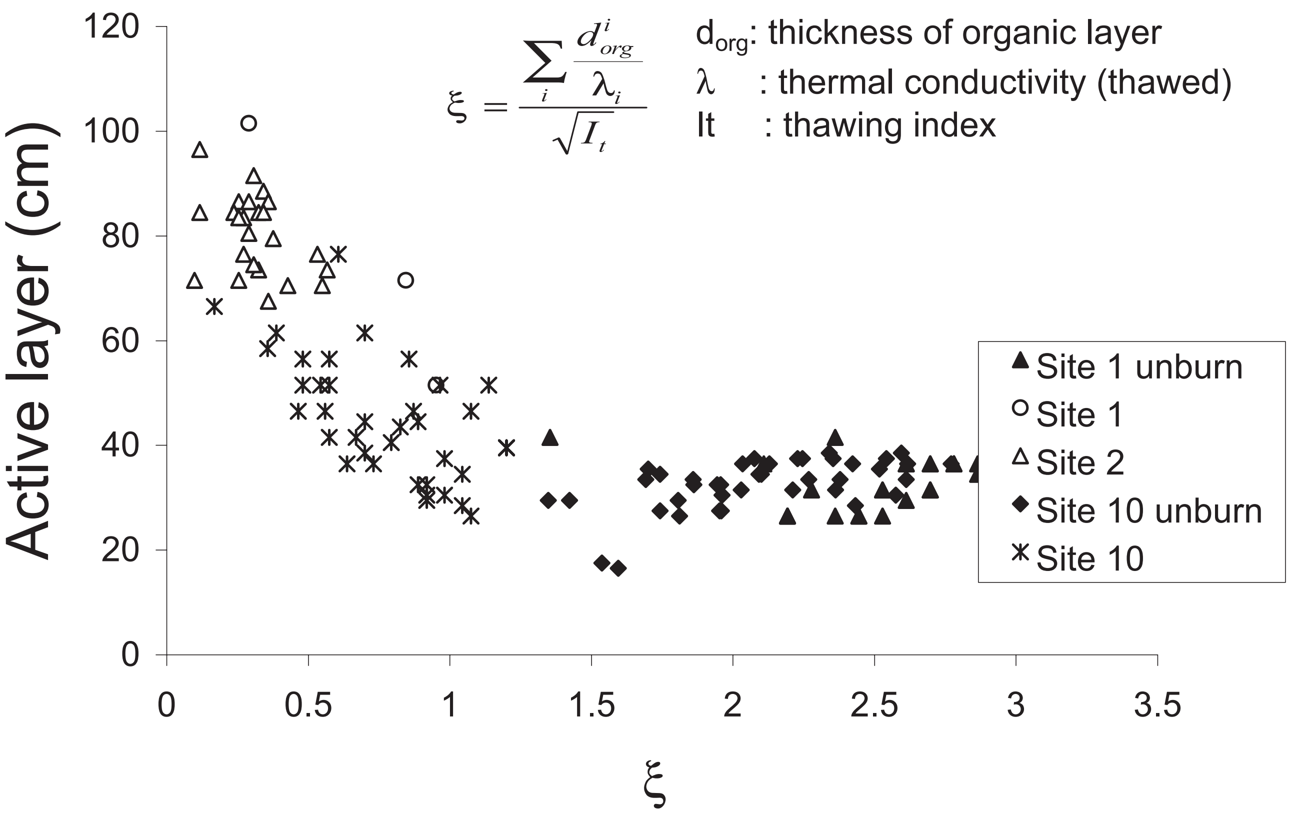

Figure 8. Active layer thickness is affected by the organic (living and dead moss, fibric, humic) layer thickness and composition, its thermal conductivity, and the thawing index. Data from sites 1, 2, and 10 in July 1999 demonstrate increasing active layer thickness with thinner organic layers.

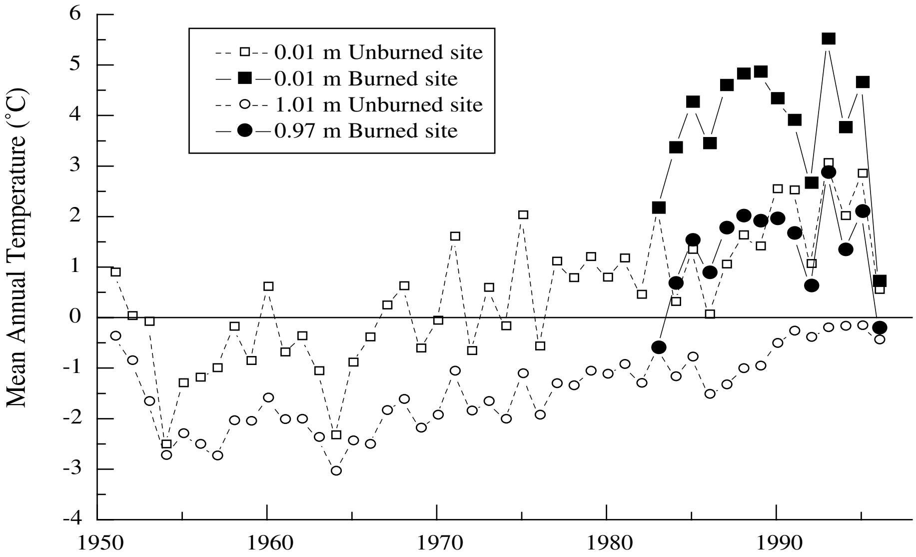

Figure 9. Modeled mean annual temperature at the ground surface (open and filled squares) and at 1 m depth (open and filled circles) at an unburned site (open symbols) and at burned site 5 (filled symbols).

properties of organic soil (ξ). This function is thermal resistance divided by the local thawing index:

$$\xi = \frac{\sum_i \frac{d_{org}^i}{\lambda_i}}{\sqrt{I_t}}.$$

where d_{org}^i is the thickness of organic layers (moss and peat, cm), λ_i is thermal conductivity of organic layers (W/m K) and I_t is thawing index of the ground surface (°C days). Thermal conductivity is strongly affected by the moisture content of moss (Figure 3). The thermal conductivity of moss and peat was determined by field and laboratory measurements [*Burwash*, 1972; *Andersland and Anderson*, 1978; *Farouki*, 1981]. In an unburned forest where permafrost is present, the function of thermal properties of organic soil (ξ) is usually greater than one. In areas where the fire severity was rated as low or moderate (organic soil was only lightly burned) such as site 10, the function of thermal properties of organic soils (ξ) will be between 0.7 and 1. As the severity of the fire increases, the function of thermal properties of organic soils (ξ) decreases. A ξ value of 0.7 or above appears to be a threshold as to when the permafrost will be adversely impacted by a wildfire. In Interior Alaska, the typical patterns of organic soils will yield thermal properties in the range of $\lambda = 0.25-0.45$ and $I_t = 1400$. Thus, after wildfire, if more than 7–12 cm of organic soil remain, the active layer thickness in the burned area will not be greater than that in the adjacent unburned area.

[32] Heat transfer processes other than by conduction do occur and have been documented in many studies [*Hinkel and Outcalt*, 1994; *Kane et al.*, 2001; *Outcalt et al.*, 1990, 1997; *Woo*, 1982]. Nonconductive heat transfer processes occur in soils primarily in association with water movement, either in the vapor or liquid phase. The processes increase in importance in soils with large pore sizes and large thermal or moisture gradients. Some heat movement coupled with vapor transport may have occurred in the active layer, decreasing the effective thermal conductivity [*de Vries*, 1974] and lowering the amount of heat transferred by conduction. It is difficult to quantify heat transfer by convection from measurements of soil moisture because the latent heat of vaporization is so large (approx. 2.5 MJ/kg), relatively small changes in moisture may accompany significant heat flux. However, we do not believe that convective transport played a major role in accounting for the differences observed among these field sites, except for brief periods when the near surface soils were quite wet and the near surface temperature gradient was quite large (such as on sunny spring days when the active layer was still frozen near the surface). In general the pore size of these soils is quite small following a fire, greatly limiting free convection [*Russell*, 1935]. The most noticeable nonconductive thermal effect is produced by snowmelt water, which during the spring can infiltrate into the upper moss layer and into the frozen soils using the thermal cracks as flow paths. However, the annually averaged effect of this process on permafrost temperatures is minimal because the infiltrating water cannot transport any significant amount of sensible heat due to small differences in the temperatures of this water and surrounding material. The latent heat of refreezing water can noticeably increase soil temperatures at this time but the same amount of latent heat will be consumed during the melt of this additional ice in the active layer during its thawing. So, the annual heat balance in the active layer will not be changed and the thermal effect on permafrost will be negligible.

5. Modeling

[33] To investigate the effects of forest fire on permafrost temperature regime, we applied numerical modeling of the active layer and permafrost temperature field dynamics at the 1983 Rosie Creek fire (Bonanza Creek LTER, site 5). The models used for these studies were described by *Osterkamp and Romanovsky* [1996] and *Romanovsky et al.* [1997]. Daily air temperatures and snow depth records

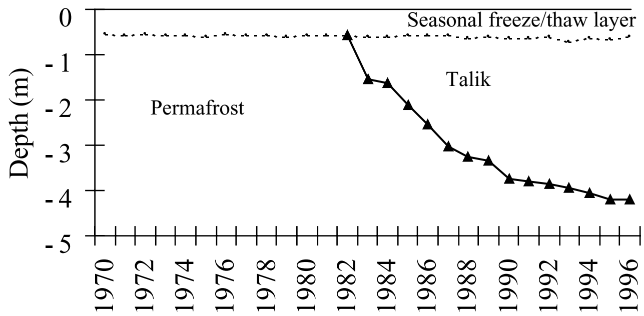

Figure 10. Simulated active layer dynamics at an undisturbed site (dashed line) and talik formation at a burned site (solid line) at site 5 from 1970 to 1996.

for the last 46 years from the Fairbanks International Airport meteorological station were used for the upper boundary conditions. The lower boundary was placed at the approximate base of the permafrost at this site (54 m). A constant heat flux of 0.04 W/m^2 was used for the lower boundary condition. The models include the effects of latent heat and unfrozen water in the active layer and permafrost since it was found that the thermal response of the active layer and permafrost could not be successfully modeled without doing so [*Riseborough*, 1990; *Burn*, 1992; *Romanovsky and Osterkamp*, 2000]. Time steps in these calculations were 15 min, while the 200 vertical steps were changed from 1 cm within the upper 1 m of soils to 1 m at the lower boundary of the spatial domain (54 m). Site specific calibration of the models was accomplished using annually measured temperature profiles at this site and daily mean temperatures measured at the ground surface and at several depths in the upper meter of soil. Daily air temperatures and snow cover thickness from the Fairbanks International Airport were used to complete the model calibration and to extend the calculations back in time. Drilling records were used to determine the lithology and the initial approximate thermal properties of the soils in the thawed and frozen states. The thermal properties (including unfrozen water content curves) were refined using a trial and error method [*Osterkamp and Romanovsky*, 1997]. The results of modeling for the undisturbed site were discussed by *Osterkamp and Romanovsky* [1999]. Here we will compare these results with calculations for the site within the LTER Bonanza Creek research area where a severe forest fire occurred in the summer of 1983 (Rosie Creek fire).

[34] As a result of this fire, the entire moss layer and most of the peat layer were destroyed. The model was modified to reflect these changes in soil properties. Other parameters of the model and driving variables such as air temperature and snow cover thickness were kept the same as in simulations of the undisturbed site. Calculations were started in the summer 1983 and the temperature profile from the no-disturbance simulation at the time of fire ignition were used as initial conditions. The calculation results are shown in Figure 9. Immediately after the fire, the ground surface and active layer temperatures increased significantly and the long-term permafrost thawing began. As is shown in Figure 9, the mean annual temperatures at the ground surface and at 1 m depth increased by 3–3.5°C during the first 5 years after the fire. This difference then decreased to 1.5–2°C during the 1990s.

[35] The modeled permafrost thawing progression and a talik development at site 5 are shown in Figure 10. The thawing of the permafrost was especially rapid during the first 5 years. By the end of this period, the depth of talik was 3.4 m. During the last 8 years of calculations, the talik increased by only 0.8 m, totaling 4.15 m in depth by 1996. Recent measurements (Figure 7) show that during the rest of 1990s, the talik did not increase with the permafrost table position was practically stationary at the 4.15 m depth.

[36] The organic soil was completely burned at site 5 (Table 2). In the case of a light burn (simulation of a 16 cm moss layer, λ = 0.7 W/m K, ξ = 1.2), this model study suggests that maximum active layer reaches only 98 cm instead of 4.15 m observed and simulated following a severe burn (simulation of a 6 cm moss layer, ξ = 0.75). The active layer depth of the area around site 5 is usually 60–70 cm at the unburned control area, while the burned area was about 20 cm deeper. Simulation results indicate at the end of the 1989–1990 winter, the seasonal frost depth did not completely reach the top of the permafrost forming a 10–20 cm thick talik. Model simulations indicate the talik continues to increase in thickness for 5 years. The modeling study suggests that severity of burn (e.g., thickness of remaining organic layer) is an important factor controlling active layer thickness and talik formation. Observations and simulation results of the site 5 area indicate permafrost degradation due to fire is about 10 years in both the lightly and severely burned sites under the current climatic conditions.

6. Conclusions

[37] Depending on burn severity, wildfires result in an immediate impact upon the permafrost and ground thermal regime. In permafrost regions, the soils in a burned site will be warmer than in an adjacent unburned site for many decades. In the short term, wildfires will decrease rates of transpiration causing increasing soil moisture contents. In the long-term (more than a decade), the increased thickness of the active layer and recovery of vegetation will cause decreases in soil moisture content in burned sites as compared to control sites. However, after a light burn, where most of the fibric layer and some of dead moss remains, the permafrost is not significantly impacted in spite of a decrease in surface albedo to less than half that of an undisturbed forest. The thermal conductivity of the soils is greatly increased by increasing in soil moisture content. There appears to be a threshold of organic material remaining after a fire that determines the degree of influence of the wildfire on frozen grounds. This threshold value is a function of the thickness of the moss, thermal conductivity of moss and thawing index at the ground surface.

[38] In the case of severe fire, the active layer begins to increase immediately following the fire, but heat from the fire itself does not affect the active layer. Formation of talik depends on the thickness of the organic layer. Entirely removing an organic layer would very quickly initiate talik formation in this region. After a talik has formed, it is unlikely that the permafrost will recover under present climate condition. The impact of the wild-

fires may influence the permafrost on a more global scale as well. With increased global warming (increasing ground surface thawing index), fire frequencies and severities may increase. Since permafrost temperatures in Interior Alaska are at or near 0°C, the influence of wildfires may have a far-reaching impact on the landscapes, ecological impacts on the vegetation types, and enhanced feedback to global warming.

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