Influence of forest clear-cutting on the thermal and hydrological regime of the active layer near Yakutsk, eastern Siberia

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[1] Thermal and hydrological conditions in the active layer were investigated at a mature larch forest and an experimental cutover, to clarify the characteristics of heat and water budget in the active layer and to assess the influence of clear-cutting on permafrost and active layer conditions. Clear-cutting enhanced ground thawing and the difference in the active layer thickness between the forest and the cutover after 1 year was 14 cm. The soil water content drastically decreased at the forest, while that at the cutover was retained during the first thaw season after clear-cutting. Although the ground heat flux continued to increase at the cutover (the difference in the total amounts between sites from May to August were 44 and 69 MJ/m^2 in 1 year and 2 years after the clear-cutting, respectively) marked changes in the active layer conditions were limited only to the first thaw season. The correspondent differences in the active layer thickness between the sites were 16 and 14 cm in 2 years and 3 years, respectively. Further increases in the maximum thaw depth at the cutover site were inhibited by the thermal inertial effect of the larger amount of ice in the second spring after disturbance. This suggests a self-retention mechanism of the active layer thickness after forest disturbance in this continuous permafrost zone.

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1. Introduction

[2] About 50% of world's coniferous forest occurs in Russia, and about 30% of the forested area of Russia belongs to eastern Siberia, forming an extensive boreal coniferous forest called Taiga [*Kakisawa and Yamane*, 2003]. The average annual precipitation in Yakutsk, central Yakutia, eastern Siberia is about 240 mm, which is categorized climatically as a steppe region [*Takahashi*, 1994]. The distribution of Taiga over this extremely dry region could be explained by the existence of continuous, ice-rich permafrost which prevents deeper infiltration of the limited rainwater. The soil moisture in the active layer, defined as the layer of ground that is subject to annual thawing and freezing in areas underlain by permafrost [*Everdingen*, 2002], is effectively stored for use by vegetation. Forest vegetation acts as a thermal insulation buffer against heat exchange between the ground and the atmosphere, with the coexistence of permafrost and forest sustaining the Taiga ecosystem [*Fukuda*, 1984, 1997]. Increasing forest disturbances [*Kakisawa and Yamane*, 2003], such as forest fires and clear-cutting, destroy much of the permafrost-surface vegetation coexistence relationship by altering ground surface conditions and water flow. To model the atmosphere-vegetation-soil interaction in permafrost regions, it is first necessary to understand the thermal and hydrological dynamics in the active layer.

[3] Climatic, physical, and biotic factors control the active layer and upper permafrost thermal regime and a wide range of responses to fire have been found, depending on the site features [e.g., *Burn*, 1998; *Mackay*, 1995; *Swanson*, 1996; *Viereck*, 1982]. *Yoshikawa et al.* [2002] concluded that the impact of wildfires on permafrost and ground thermal regime is dependent on burn severity, such as the degree of surface organic mat loss by fire, but *Linell* [1973] showed that clear-cutting was effective for permafrost degradation in discontinuous permafrost zones. Forest disturbance also alters the hydrological regime of the active layer: In interior Alaska, *Sharratt* [1998] concluded that clearing forestland enhances soil drying and *Ping* [1987] noted that removal of vegetation cover changes the drainage of the active layer from poorly to moderately well drained. Comparisons of the surface energy balance at different sites representing different stages of vegetation recovery after forest disturbance have been performed in North America [e.g., *Rouse*, 1976; *Chambers and Chapin*, 2002]. In eastern Siberia, *Sherbakov* [1979] reported soil temperature

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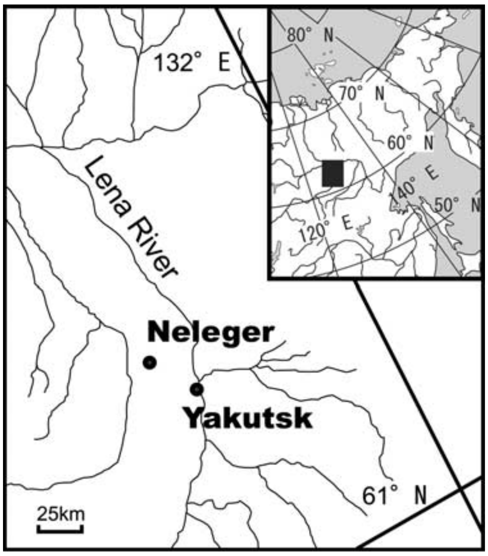

Figure 1. Location of the study area (Neleger) near Yakutsk.

increases at experimentally burnt sites and observed temporary increases in soil moisture at some sites, but concluded that the more general change in soil moisture content after forest fires was a decrease in soil moisture storage. The interaction between surface conditions and permafrost in eastern Siberia is still limited, despite the work of *Pavlov* [1975], *Gavrilova et al.* [1996] and *Varlamov et al.* [2002]. Although thermal and hydrological dynamics in the active layer at a single site have been analyzed in continuous permafrost zones [e.g., *Boike et al.*, 1998; *Roth and Boike*, 2001; *Yoshimoto et al.*, 1996], comparisons of detailed observations of the active layer between sites with different surface conditions have rarely been made.

[4] This paper focuses on forest clear-cutting as a surface disturbance in a permafrost region, by treatment as changes in the energy and water balance characteristics of the surface. Since ground heat flux is often the smallest component in energy balance equations of the surface where thick vegetation exists, it has attracted little attention in micrometeorological studies. In permafrost studies, ground heat flux plays an important role in the determination of the active layer and upper permafrost thermal conditions. Our objectives are to describe overall effects of the clear-cutting on the thermal and hydrological characteristics of the active layer in a continuous permafrost zone and to find changes in the energy partitioning of the ground heat flux immediately after the disturbance.

2. Site Descriptions

[5] The research area, Neleger, is located about 35 km northwest of Yakutsk (62°19′N, 129°31′E), Russia (Figure 1), 24 km from the Lena River and 14 km from the slope edge of the Lena valley. The permafrost, which impedes downward water drainage, is very thick (around 400–500 m) in this region [*Ivanov*, 1984]. The mean annual temperature at Yakutsk is approximately −9.9°C and annual precipitation is about 239 mm, based on data from 1951–2000. The study sites were located on the highest terrace of the left bank of the Lena River, 215–221 m above mean sea level. This area was covered with a thin layer (6–8 m) of Quaternary deposits consisting of alluvial deposits with ice complex of the Lena River or recent temporary streams and creeks. The terrace was almost flat and was covered with pure and mixed stands of larch, birch, pine, swamp, and thermokarst features called "Alas."

[6] Intact forest experimental cutover sites were established. Site F was a mature larch stand (*Larix gmelinii*; over 200 years old) with a mean tree height of 8 m, a mean canopy height of approximately 15 m and a density of 0.21 trees/m^2 based on measurements of tree specimens taller than 1.3 m [*Yajima et al.*, 1998] and 100–150 m from the nearest Alas. The underlying vegetation was mainly comprised of mountain cranberries (*Vaccinium vitis-idaea*), with some lichens, mosses, shrubs (e.g., *Betula platyphylla*), and herbaceous plants. Site C was a rectangle (70 × 170 m) located about 100 m from Site F that was cutover between 25 November and 10 December 2000. The trees were removed and the mountain cranberries were damaged during the cutover, but the surface organic mat (layer of vegetation and soil organic matter) was not compacted.

3. Methods

[7] Measurements in the active layer were conducted at sites F and C from 1 May to 4 October 2001 and from 20 April to 9 October 2002. Soil temperatures were measured at depths of 0.01, 0.05, 0.1, 0.15, 0.2, 0.25, 0.3, 0.4, 0.6, 0.8, 1.0, 1.2, 1.4, and 1.6 m at site F, and 0.01, 0.05, 0.1, 0.15, 0.2, 0.25, 0.4, 0.7, 1.1, and 1.6 m at site C using calibrated thermistors (104ET, Ishizuka Denshi, Tokyo, Japan) at site F and calibrated probes (107 probe, Campbell Scientific, Inc.) at site C. Soil temperature probes were calibrated with an ice-water bath with a precision of 0.02°C at 0°C and the overall probe accuracy for the temperature range of −20 to 30°C was calculated as less than ±0.09°C. Soil moisture was measured by the TDR method (CS615 probes, Campbell Scientific, Inc.) at depths of 0.05, 0.15, 0.25, 0.4, 0.6, and 0.92 m at site F, and 0.05, 0.15, 0.25, 0.4, 0.7, and 1.1 m at site C. Calibrations for TDR probes were conducted gravimetrically using 3 to 10 samples at each depth for each site.

[8] The soil temperature measurements were conducted every 10 s and stored every 10 min as averages. Soil moisture measurements were conducted every hour and recorded. All data were logged using a CR10X data logger (Campbell Scientific, Inc.).

[9] Soil samples were taken from layers every 0.05 to 0.1 m down to the maximum thaw depth at each site from 6 to 11 September 2000. The thickness of the surface organic mat, soil grain size characteristics of the active layer mineral soil, bulk density profile, and organic content profile (only for site F) were determined. The relations between negative temperature and volumetric unfrozen water content were determined for the mineral layer as an average of 21 samples from laboratory experiments by the NMR method [*Ishizaki et al.*, 1990]. Considerable amounts of liquid water can exist at negative temperatures [*Farouki*, 1981], and the dependence of the unfrozen water content on temperature in soils found in this study was similar to that for silty-clay soil reported by *Yershov* [1998].

[10] Thaw depth was determined by pounding a steel rod into the ground until frost was encountered. *Mackay* [1995] discussed the unreliability of this method when the active layer is deep and with high unfrozen water content but the upper layer of permafrost at the study sites was a silty-clay deposit with high ice content, resistance of the permafrost was very clear during probing even when the active layer was deep and had a high water content. The spatial distribution of thaw depth was measured at the end of August 2000 (before the cutover) on a 150 × 100 m grid set (856 points), including all of site C and part of site F. The measurements were repeated on a 50 × 50 m grid set (152 points) at site F and on a 150 × 50 m grid set (295 points) at site C at the end of August 2001, at site F (35 points) and site C (239 points) at the beginning of September 2002, and at the end of August in 2003 at site F (121 points) and at site C (250 points).

[11] Snow depth was measured at each site within the grids of the thaw depth survey on 10 April 2002, with average values calculated for 50 samples from each site. Temporal changes in snow depths were observed from six surveys during the winter taking six to seven samples each time and at each site. The data were taken at a similar forest to site F located about 400 m south of site F in 2000-2001 winter.

[12] Volumetric ice content in the ground was measured with an augur (JEA-50, Tanaka Kogyo Co., Ltd., Japan) with the samplers (φ 63 and 51 mm). The weight of the sampled frozen cores was measured in air and in a kerosene bath. Net radiation was measured at each site using a four-component net radiometer (CNR1, Kipp and Zonen, Saskatoon, Saskatchewan, Canada). Calibrations were performed after the measurements and found to differ from each other by less than 1%. Radiometers were mounted at heights of 21 m at site F and 9 m at sites C. Net radiation (R_n) is defined as $R_n = S_i - S_o + L_i - L_o$, where S_i, S_o, L_i, and L_o are solar radiation, reflected solar radiation, longwave radiation emitted from the surface and atmospheric longwave radiation, respectively. Air temperature and relative humidity were measured at a height of 1.5 m with a thermistor and capacitive relative humidity sensor probe (HMP45C, Vaisala, Helsinki, Finland) in ventilated shelters at each site. All supporting meteorological and ground data, except for precipitation, were measured every 10 s and stored as 10-min averages using data loggers (CR10X, Campbell Scientific Inc., Edmonton, Alberta, Canada). Precipitation was measured using a tipping bucket rain gauge with sensitivity of 0.1 mm (52203, R. M. Young Co.) and logged as the sum of 10 min in an open area (about 200 m south of site C) in 2000 and at site C from 2001 to 2003.

4. Analysis

[13] To analyze the ground thermal and hydrological dynamics, we considered a compartment of the soil layer from the upper limit of the surface organic mat to a depth of 1.5 m as a reference layer and applied a simple energy and water balance model to this compartment. At a depth of 1.5 m, which was deeper than the active layer thickness at both sites, we assumed heat conduction was the predominant heat transfer process and moisture migration was negligibly small. We also assumed that heat transfer processes in the compartment were instantaneous and that soil thermal, physical, and moisture conditions were horizontally homogeneous and thus that heat and water flow occurred only in the vertical plane. For flat areas, such as the sites examined in this study, the assumption of no lateral water flow is probably valid. However, if the permafrost table were not flat, water potential gradients would be a potential driving force of lateral water migration. In this study, we refer to ground heat flux G as energy flux through the upper limit of the surface organic mat.

4.1. Energy Balance

[14] G can be calculated as the sum of the storage of sensible heat (Q_S) and the storage of latent heat (Q_L) [*Rouse*, 1984]. Assuming there is no lateral heat or water flux within the soil and transfer processes within the unfrozen part of the soil are instantaneous, the ground heat flux (G) through the soil surface may be formulated [from *Rouse*, 1984; *Boike et al.*, 1998] as

$$G = Q_S + Q_L, \tag{1}$$

where Q_S is the sum of the storage of sensible heat in the reference layer in the balance model Q_A and the conductive heat flux at the depth of 1.5 m Q_P, which is the lower limit of the compartment. Q_S can be expressed as

$$Q_S = Q_A + Q_P, \tag{2}$$

$$Q_A = \sum_j C_j \frac{\Delta T_j}{\Delta t} h_j, \tag{3}$$

$$Q_P = k_f \left(\frac{dT}{dz}\right)_f. \tag{4}$$

The soil compartment was divided into five to seven layers based on the depth at which temperature and water content probes were installed at each site. Q_A was calculated as in equation (3) where Δt is calculation interval in which soil thermal properties are stable and we used 3600 s. $\Delta T_j/\Delta t$ and h_j represent the rate of temperature change in the time interval Δt and layer thickness of the soil layer j, respectively. Layer j has soil with heat capacity C_j, which may be estimated as the sum of heat capacity of components α, giving

$$C_j = \sum_{\alpha \in \{s,w,i\}} c_\alpha \rho_\alpha \theta_{\alpha j}, \tag{5}$$

where the subscripts s, w, and i refer to soil particles, unfrozen water, and ice, respectively, and c_α, ρ_α, and θ_α indicate the specific heat, the bulk density, and the volume fraction of the component α, respectively. The ice content or unfrozen water content in the frozen soil was estimated using the relationship between negative temperature and volumetric unfrozen water content measured by the NMR method. Q_P was calculated by equation (4) where k_f is the thermal conductivity of the soil and $(dT/dz)_f$ is the vertical temperature gradient at a depth of 1.5 m. We used the experimental equation of Kersten for frozen silty-clay to estimate k_f [*Kersten*, 1949], which has two potential errors. Although the temperatures at a depth of 1.5 m remained negative, there is a potential error associated with the temperature dependence of the thermal conductivity of

Table 1. Specific Heat and Density of the Active Layer Materials at 0°C used in the Energy Budget Analysis[a]

	Specific Heat, J/g/K	Density, Mg/m^3
Water	4.22	1.00
Live plant layer	1.92	1.30
Mineral soil	0.90	2.65
Ice	2.11	0.92

[a]Latent heat of fusion: 0.333 MJ/kg.

frozen soil varies from 1.2 to 1.6 W/m/K [*Yershov*, 1998]. There is also an error in the estimation of temperature gradient by linear interpolation which increases in the middle of summer, when the temperature difference between the upper layer and permafrost widened. However, these errors did not influence the structure of the energy balance in the active layer because of their relatively small magnitude.

[15] Q_L was calculated from the rate of thaw depth change $\Delta D/\Delta t$ and volumetric ice content of soil that thaws in the time interval Δt. This gives

$$Q_L = L_f \theta_i \frac{\Delta D}{\Delta t}, \tag{6}$$

in which L_f is the specific latent heat of fusion for ice. It was assumed that the water content profile measured just before freezing of the active layer did not change at negative temperatures. Although water migration in frozen soil induced by temperature gradient could be expected as *Roth and Boike* [2001] discussed, the amount of water in the active layer will not change significantly. The properties of soil components and value of L_f used in this analysis are given in Table 1 [*De Vries*, 1963].

[16] Using a heat flux plate, the combination method [*Oke*, 1987] has commonly been used to estimate ground heat flux, but the thermal conductivity of the soil at the depth at which the heat flux plate is installed significantly affects the measurements [*Schwerdtfeger*, 1976]. In this study, large differences in thermal properties between the heat flux plate and surrounding soil were expected owing to the high organic content and seasonal variations in volumetric water content in the upper layer. Furthermore, *Rouse* [1984] and *Boike et al.* [1998] noted that heat flux plates are an unreliable form of measurement in permafrost terrain because of sensible heat flux in the form of liquid water and latent heat flux in the form of water vapor at the depth at which the heat flux plate is installed.

4.2. Water Balance

[17] The total amount of water storage S in the compartment down to a depth of 1.5 m including liquid water and ice per unit area can be defined as

$$S = \sum_{\alpha \in \{i,w\}} \sum_j \theta_{\alpha j} h_j. \tag{7}$$

[18] Assuming no lateral runoff and no downward infiltration, the water balance in the soil for a given period can be formulated as

$$P = E + \Delta S, \tag{8}$$

where P, E, and ΔS are precipitation, evapotranspiration at the height of the vegetation top, and rate of water storage change for a given period. In this study, we estimated E as a residual of the balance equation (8).

5. Results

5.1. Climate and Soil Physical Properties

[19] Seasonal variations in daily rainfall, daily mean solar radiation (S_i), air temperature (T_a), and relative humidity

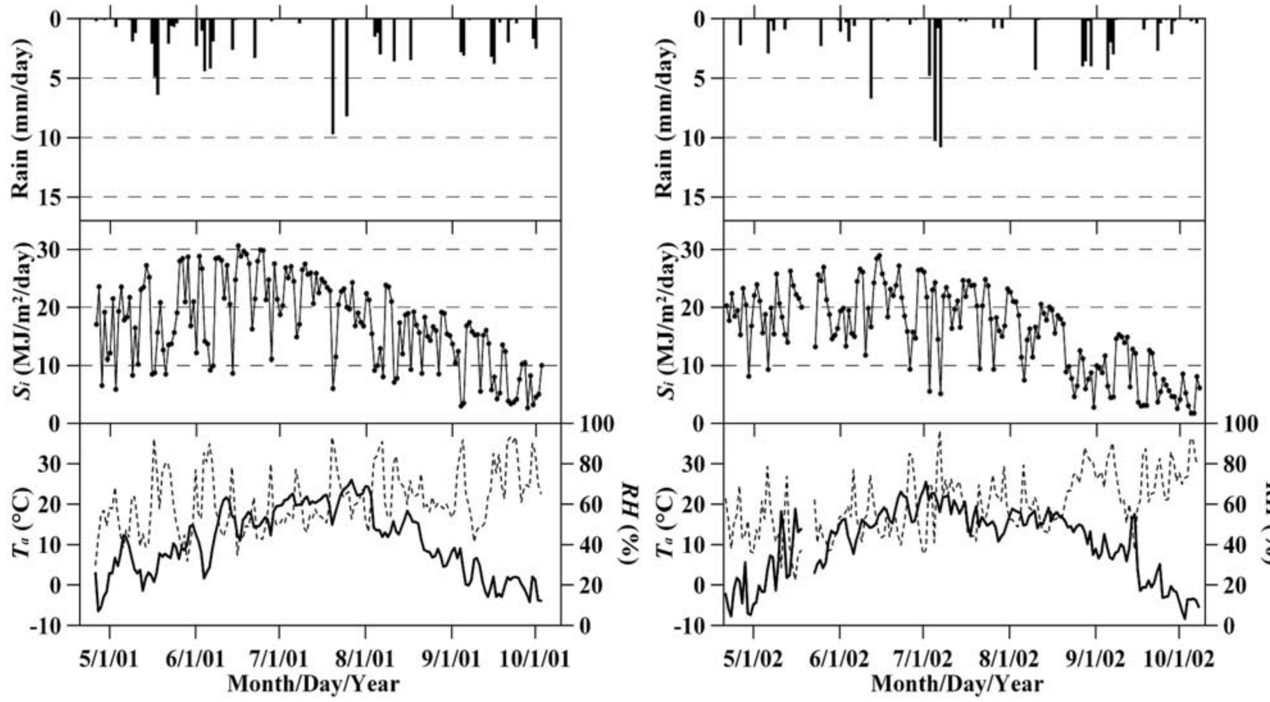

Figure 2. Seasonal variations in daily rainfall, solar radiation (S_i), mean air temperature (T_a; solid line) and mean relative humidity (RH; dashed line) at site C in (left) 2001 and (right) 2002.

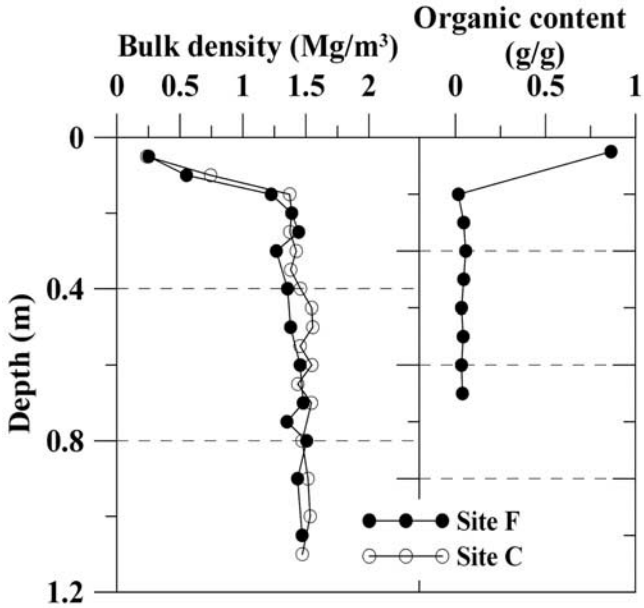

Figure 3. Bulk density and organic matter content profiles of the active layer at sites F and B. Values are averages from four to six samples at each depth.

(*RH*) at site C from May to October, 2001 and 2002, are shown in Figure 2. S_i reached its peak value of about 30 MJ/m^2/day at the end of June and the peak of T_a was delayed by about half a month from the S_i peak and *RH* tended to increase from spring to fall. Precipitation from May to September in 2001 and 2002 (90 and 79 mm, respectively) were smaller than in 2000 (128 mm), 2003 (223 mm) and the 50 years average (159 mm), and *RH* was low (59%) during June and July in 2001 and 2002 compared to 2000 (68%). Average air temperatures during June and July were higher in 2001 and 2002 (17.7° and 17.2°C, respectively) and lower in 2000 (15.7°C) or in 2003 (16.3°C). We treat year 2000 as dry and cold conditions, and years 2001 and 2002 as very dry and hot. There were no marked differences between forest and disturbed sites in daily mean values through the thawing season.

[20] The thicknesses (0.1 to 0.2 m) of the surface organic mat were similar at the forest and disturbed sites. The average grain size characteristics in the active layer mineral soil were 21, 60 and 19% at site F and 40, 47 and 13% at site C for sand, silt, and clay, respectively, falling into the silty-clay texture class. Bulk density profiles were similar between the two sites (Figure 3). It was assumed that there were few spatial differences in the soil properties between sites F and C, because they were located only 100 m apart and site C was initially the same forest as site F before clear-cutting. From the results of this section and considerations regarding weather conditions at each site, we concluded that with the exception of the moisture conditions, weather and physical conditions in the active layer were almost similar at both sites.

5.2. Impacts on Seasonal Variations in Soil Temperature and Moisture

[21] The maximum thaw depth deepened dominantly at site C in the first thawing season after the clear-cutting treatment. Active layer thawing commenced at the beginning of May in 2001 when the snow had melted completely (Figure 4). Although temperature data were missing at site F, we can infer that thawing started on the same day as at site C, on the basis of snow disappearance on the same day. The temperatures at a depth of 1.6 m were −4.9° and −6.7°C on 11 May 2001 at sites F and C, respectively. We presume the soil temperature profiles were similar at both sites before clear-cutting, and this difference is probably due to intensive cooling at site C during the 2000−2001 winter after experimental clear-cutting. The surface vegetation was completely removed and the accumulated snow was com-

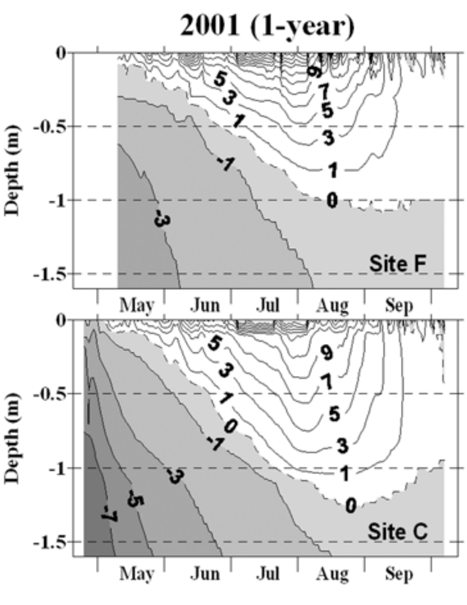

Figure 4. Seasonal changes in the soil temperature (°C) profiles at sites F and C in 2001 and 2002 based on daily mean measurements. Solid isotherms are drawn with increments of 2°C, and the dashed line indicates the 0°C isotherm. See color version of this figure in the HTML.

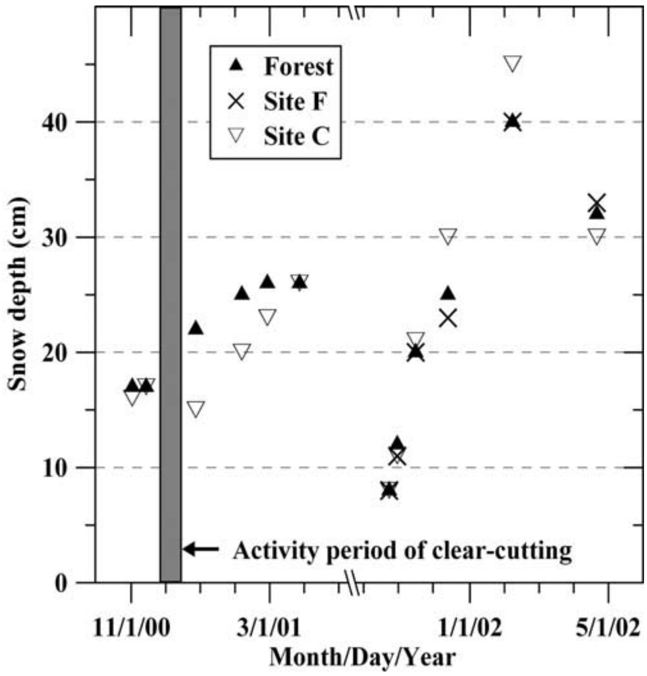

Figure 5. Seasonal changes in snow depths at forested and cutover sites based on average of 6 to 7 samples. "Forest" was a larch forest stand located 400 m south of site F.

pressed during the clearing operation (Figure 5). The density of the snow cover in the Yakutsk region during the winter varies from 0.13 to 0.17 Mg/m^3 [*Shver and Ijumenko*, 1982]. In central Yakutsk, with light winds, there is practically no difference in snow density between forests and open areas [*Pavlov*, 1984]. The thermal conductivity of snow depends on its density, but *Sturm et al.* [1997] and *Sturm and Johnson* [1992] experimentally showed smaller thermal conductivity in depth hoar, which is the metamorphosed structure in the lower layer of snow cover. The loss of vegetation, reduced thickness of snow and destruction of the hoar structure at site C means the loss of the thermal buffer layer. The protective effect of snow against ground cooling has been studied by a number of researchers [e.g., *Smith*, 1975; *Goodrich*, 1982; *Williams and Smith*, 1989]. Active layer development varied in 2001: The progress of the thaw front at site C was faster than that at site F, with average thawing rates during June and July of 0.010 and 0.015 m/day at sites F and C, respectively. Rapid progress of the thaw front ceased at the beginning of August at site F and reached the maximum thaw depth at the beginning of September, while at site C rapid thawing continued until the end of August when the maximum thaw depth was observed. Despite the colder soil temperature profile at the beginning of the observation, the maximum thaw depth determined from the temperature profile was significantly deeper (1.09 m) at site C than at site F (1.19 m) in 2001. The closure of the zero curtain [*Outcalt et al.*, 1990] at site C was very rapid, while that at site F seemed to have been completed much later.

[22] In 2002, the second year after experimental clear-cutting, the active layer development patterns was similar to 2001, but with a faster thaw down to a depth of 0.3 m at site F, which may have been due to soil moisture changes in the previous summer. At the beginning of May, differences in the soil temperature profiles were small between the two sites: At a depth of 1.6 m, for example, temperatures were −5.0° and −4.6°C on 1 May 2002 at sites F and C, respectively. The snow cover at site C was thicker than that at site F (Figure 5), reducing ground heat loss during winter at site C than at site F and producing a positive effect on the increase of thaw depth in the following year. The average thawing rates of frozen ground during June and July in 2002 were 0.011 and 0.013 m/day, and the maximum thaw depths

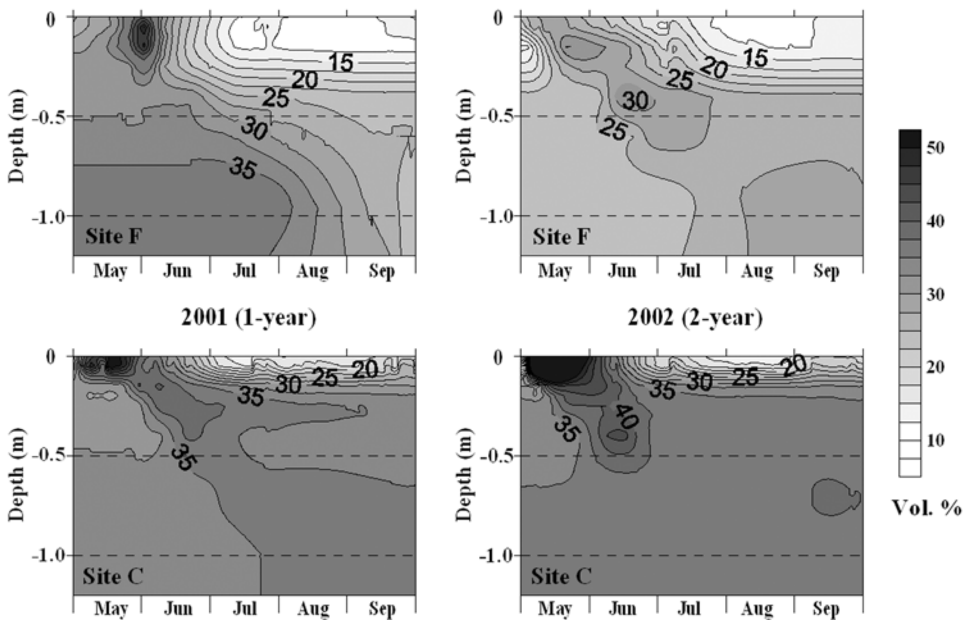

Figure 6. Seasonal and successive changes in the total soil water (liquid water and ice) content profiles at sites F and C in 2001 and 2002 based on daily mean measurements of volumetric water content (%). Contour lines are drawn with increments of 2.5%. See color version of this figure in the HTML.

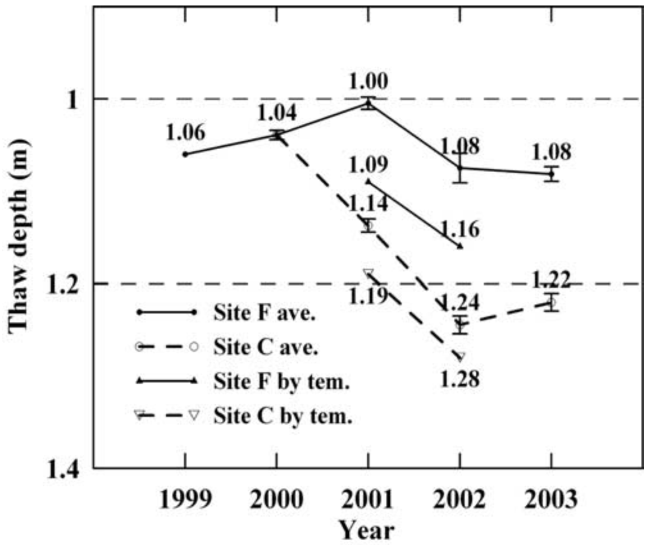

Figure 7. Successive changes in the average maximum thaw depths determined by pounding a steel rod at sites F and C. The maximum thaw depths determined from the temperature profile are also shown for the summers of 2001 and 2002. Error bars denote the standard errors.

were 1.16 and 1.28 m at sites F and C, respectively. We expected further increases in the difference in maximum thaw depth between the two sites, but this was only 0.02 m.

[23] Marked differences were observed in moisture in the first thawing season after clear-cutting, assuming no migration in the frozen layer during winter and the profiles of soil water contents just before freezing were maintained until the thawing season in the next year (Figure 6). There was substantial water loss from the soil at site F, whereas soil water was retained at site C during the thawing season in 2001 and the difference in soil water content continued in 2002. At site F, a sharp decrease in the soil water in the upper 0.5 m layer of the soil occurred in June and the beginning of July and the soil water in the deeper layer was gradually lost in 2001. In the early period of the thawing season in 2002, soil water content increased owing to the input water from melting snow but was lost quickly until the middle of July. The surface organic mat was desiccated during most of the thaw season in 2001 and 2002, but soil water deeper than 0.2 m remained almost constant at site C in 2001 and 2002, except in early spring when there was infiltration of meltwater. Striking gradients of soil water content were found in the upper 0.2 m after July at site C and the surface organic mat was very dry, as observed at site F.

5.3. Spatial Distribution of the Active Layer Thickness

[24] The average values of thaw depth were somewhat less, by 0.05 to 0.10 m, than those determined by point measurement of soil temperature profiles. This shows the difficulty of determination of representative thaw depth at a site from a point measurement of soil temperature profile. Nevertheless, the differences in thaw depths between sites were consistently similar in both methods, so we concluded that relative differences in soil thermal regime could be obtained from the temperature measurements. Although measurements of the active layer thickness were different in each year (from the end of August to the middle of September), we treat the statistics as their maximum values, because changes in thaw depth during this period were very small (less than 0.01 m). The difference in average maximum thaw depth in the first thawing season after experimental cutover (2001) between sites F and C was significant (0.14 m), in the second thawing season (2002) there was little difference in these values (0.02 m), and in the third thawing season the difference decreased to 0.08 m (Figure 7). Marked changes in the average active layer thickness were observed only in 2001 and the permafrost degradation affected by the cutover seemed to cease in 2003.

5.4. Impacts on Energy and Water Balance Characteristics of the Active Layer

[25] We computed monthly energy balance components as averages of daily values in the active layer at sites F and C in 2001 and 2002 (Figure 8). Seasonal changes in G values were similar at both sites and years, except for their magnitude. The values of G were large in the first 3 months, in August the G values decreased markedly, and in September, G values became negative, indicating net loss of heat from the ground to the air. Total G value from May to August at site C was 33% larger than that at site F in 2001. At site F, the G in September had a very small negative value, while that at site C had a much larger negative value, indicating faster cooling in the fall. The absolute values of $Q_S(=Q_A + Q_P)$ consistently accounted for small portions of the energy balance components in the active layer at both sites and years, except in September at site F. The Q_A values, the energy consumed for temperature changes in the

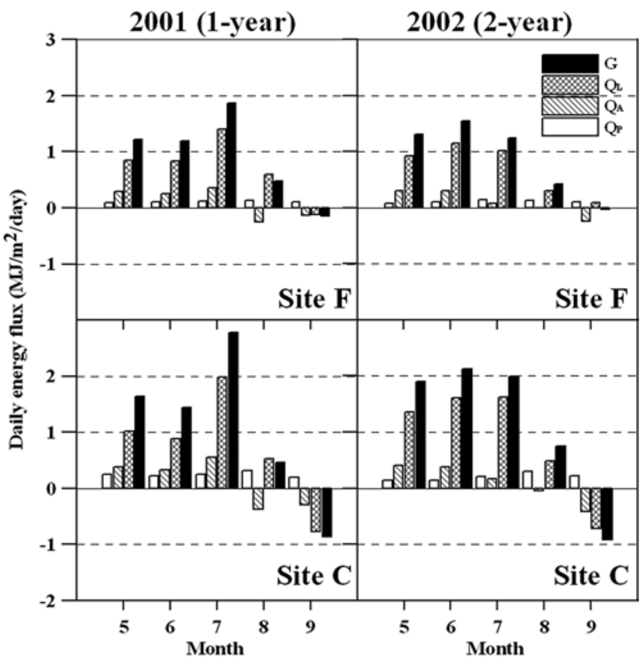

Figure 8. Monthly energy balance components in the active layer at sites F and C in 2001 and 2002. G is ground heat flux, Q_L is the storage of latent heat, Q_A is storage of sensible heat over the upper 1.5 m, and Q_P is the conductive heat transfer into the permafrost layer deeper than 1.5 m.

Table 2. Total Sensible and Latent Heat Storage Changes in the Soil, Net Radiation, and Ground Heat Flux Over 4 Months From May to August in 2001 and 2002 at Sites F and C[a]

Year	Site	Q_S	Q_L	G	R_n	G/R_n
2001 (1-year)	F	29	104	133	1383	0.10
	C	51	126	177	1125	0.16
2002 (2-year)	F	35	104	139	1332	0.10
	C	52	156	208	1076	0.19

[a]Q_S, sensible heat storage; Q_L, latent heat storage; R_n, net radiation, G, ground heat flux. Units for Q_S, Q_L, G, and R_n are MJ/m^2.

active layer, were positive from May to July as monthly averages. From August, Q_A values were negative indicating a decrease in the average temperature of the active layer. Meanwhile, the changes in Q_P value, the conductive heat flux at the bottom of the energy balance compartment, were consistently downward indicating heat storage in the permafrost throughout the observation period at both sites and years. On the other hand, Q_L comprised a large portion of G (78% and 71% at sites F and C, respectively) as total amounts from May to August in 2001 and similar significance of Q_L was found in 2002 (Table 2). Q_L appeared to be the most important component controlling energy balance characteristics of the active layer in this region.

[26] Water balance in the active layer was analyzed from June to September and showed different behaviors of the seasonal changes in ΔS and E at sites F and C in the first thawing season after clear-cutting. ΔS changed markedly in June and July at site F and large E from the forest was estimated and E decreased gradually toward the defoliation period (Table 3). In contrast, soil moisture was retained during the thawing season in 2001 at site C with smaller values of E with less seasonal variation, compared to site F. Total storage changes of soil water at sites F and C over the four months from June to September were -143 and -2 mm, respectively, with a precipitation of 69 mm. Consequently, much less of the total E at site C was estimated (71 mm): The difference in E values between sites F and C was 141 mm over the same period. This is comparable to the fact that there was large difference in spatial mean ice contents in the active layer during freezing season before the second thawing season (Figure 9). The difference in water equivalents of ice contents down to 1.4 m in Figure 9 between two sites was 133 mm.

Table 3. Monthly Averages (mm/day) of Daily Precipitation (P), Water Storage Change in the Soil (ΔS), and Evapotranspiration (E) From June to September in 2001 and 2002 at Sites F and C

Year	Site	Period	P	ΔS	E
2001 (1-year)	F	June	0.7	−1.9	2.6
		July	0.6	−1.4	2.0
		Aug	0.4	−0.8	1.2
		Sept	0.6	−0.5	1.1
	C	June	0.7	0.1	0.6
		July	0.6	−0.3	0.9
		Aug	0.4	0.2	0.2
		Sept	0.6	−0.1	0.6
2002 (2-year)	F	June	0.4	0.1	0.4
		July	0.9	−0.5	1.4
		Aug	0.5	0.0	0.5
		Sept	0.5	0.0	0.5
	C	June	0.4	−1.1	1.5
		July	0.9	0.0	0.9
		Aug	0.5	0.3	0.2
		Sept	0.5	0.0	0.5

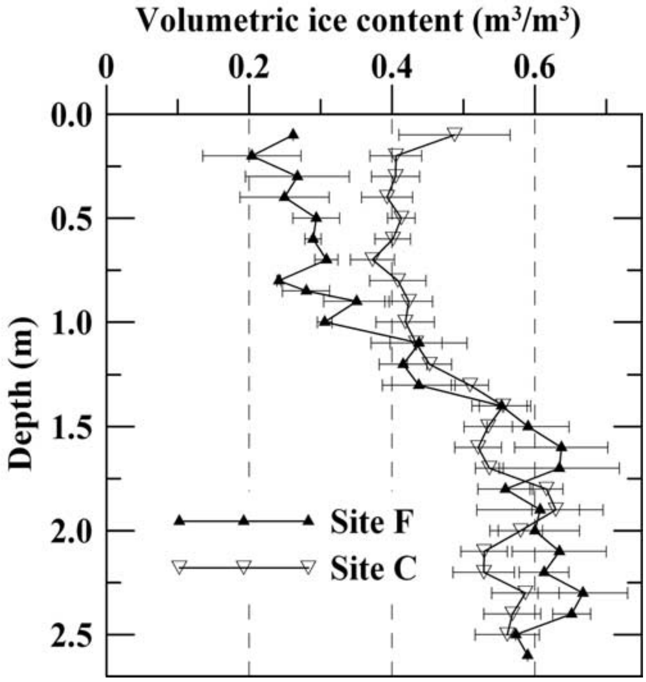

Figure 9. Volumetric ice content profiles of frozen soils in spring 2002 at sites F and C, based on averages from 6 and 11 boring points for sites F and C, respectively. Error bars denote the standard errors.

[27] In the second thawing season after clear-cutting (2002), estimated E values at site F were considerably decreased despite the very dry and hot weather. Owing to the very dry conditions in the soil (Figure 6), rainwater was evapotranspired rapidly at site F in June, August, and September. As the thaw depth proceeded from 0.3 to 1.1 m in June and July, it is possible that trees at site F could use the soil water in this zone. Large soil water loss was observed at site C in June 2002. The snow water could readily infiltrate into the mineral soil layer at site F because the upper mineral layer was kept dry from the fall of the previous year, while the upper layer at site C was kept wet. Hence the soil water supplied by melted snow remained in the upper layer at site C until June (Figure 6) and evaporation from the surface was enhanced by the hot and dry weather conditions. From July, the surface organic mat dried significantly and evaporation from the mineral soil layer was inhibited. There was little change in soil water storage at site C from July to September indicating that most rainwater evaporated immediately after each rain event. The controlling effect of evaporation by surface organic layer has also been observed by *Riseborough and Burn* [1988]. No marked differences in the total changes in soil water storage or evapotranspiration from each site were found in 2002 (11 mm).

6. Discussion

6.1. Energy Balance in the Active Layer in Other Permafrost Regions

[28] In the arctic tundra of Alaska, where the active layer thickness is about 0.3 m, the ratio of Q_L to G was small

(15–20%) during the thawing season and the majority of G was conducted into the permafrost [*Yoshimoto et al.*, 1996]. In contrast, *Rouse* [1984] reported that Q_L comprised 87–90% of G during the thawing season in the Canadian arctic tundra and forest and *Boike et al.* [1998] also found a large portion of Q_L in G (70–100%) during the thawing season in the tundra of central Siberia; both reported active layer thicknesses exceeded 1 m, as in our study sites. The different characteristics of energy balance at the site reported by *Yoshimoto et al.* [1996] can be explained by the small amount of G during summer: Heat loss from the active layer started before G was consumed for the phase change of ice into liquid water.

[29] A 33% larger G was calculated at site C over the 4 months from May to August than at site F, while net radiation over surface vegetation was 19% smaller at site C in the first thawing season after clear-cutting (2001). In the next thawing season (2002), the difference in total amount of G increased significantly to 50% although net radiation over surface vegetation was still 19% smaller at site C than at site F. G comprised 16 and 19% of net radiation during the same period at site C in 2001 and 2002, respectively, smaller than the value of 35% obtained by *Haag and Bliss* [1974] in a Canadian forest. *Yoshikawa et al.* [2002] reported that the impact of wild fires on changes in active layer thickness depends largely on the degree of the surface organic mat loss by fire. The surface organic mat remained after disturbance at our sites preventing large increases in the ratio of G to net radiation, but the increased energy partitioning in G must take into account all components in the energy balance of the surface vegetation layer.

6.2. Self-Retention Mechanism of Active Layer Thickness

[30] Forest disturbance strongly altered soil thermal and hydrological conditions, but the differences in temperature and water content profiles were limited to the first thaw season after disturbance. The stabilization of the increase in the thaw depth difference between the forest and disturbed sites was unexpectedly rapid and it seemed that the active layer thickness at the disturbed site will approach that in the forest in this study, as surface vegetation recovers. Changes in the cumulative energy balance components in the active layer from 2001 to 2002 (Table 2) indicated that there was a marked increase in the difference in Q_L between sites F and C, although the differences in Q_S(=Q_A + Q_P) were almost the same in both years. The large difference in Q_L between sites F and C in 2002 indicates a much greater consumption of energy for thawing of ground ice at site C as a consequence of water storage in the previous summer.

[31] The primary influence of forest removal was an increase in water storage and the magnitude of the change in soil water content obviously depends on the initial condition before the disturbance and also on the precipitation condition before and after the disturbance. Our study showed the response of soil moisture for clear-cutting during two consecutive dry years after relatively humid period, leading to a marked retention of soil water at the disturbed site. The sudden increase in water storage ability can act as a buffer for permafrost degradation: the excess ice content in the active layer requires much more latent energy of fusion at disturbed sites delaying the advance of the thaw depth. This is one of the reasons for the rapid stabilization of the increase in maximum thaw depth at disturbed sites, even though the increased ground heat flux was calculated in the next thawing season after the disturbance. Vegetation recovery within a few years after forest disturbance also has negative effects on the degradation of permafrost [*Iwahana et al.*, 2003]. Eventually, there is strong tolerance to rapid deep thawing of the permafrost, which leads to thermokarst processes, as a self-retention mechanism of the active layer thickness immediately after forest disturbance in this ecosystem.

7. Summary and Conclusions

[32] Thermal and hydrological conditions in the active layer were investigated at a mature larch forest and a cutover site during the thawing season in 2001 and 2002. Overall impacts of forest clearing on the active layer were as follows: (1) There was a significant increase in soil temperature and the maximum thaw depth at the 1-year cutover site due to a larger ground heat flux. (2) There was a marked increase in soil water storage caused by reduced evapotranspiration at the 1-year cutover site.

[33] Other responses to clear-cutting may be expected if the site was located on a slope, or if the site was underlain by thin permafrost, for example, in the discontinuous permafrost zone. Lateral groundwater flow in an active layer or the entire degradation of permafrost causes thoroughly alteration of the hydrological regime in the active layer. This paper displayed the influence of clear-cutting on the active layer conditions in the case of a flat continuous permafrost zone.

[34] Marked changes in the active layer conditions were limited to the first thawing season after clear-cutting. The difference in the maximum thaw depth did not expand significantly in the second thawing season, despite the increased ground heat flux at the cutover site. Thermal and hydrological analyses of the active layer revealed that the storage of latent heat was a predominant component in the energy balance in the active layer. Thus soil moisture condition, especially spring ice content in the active layer, plays an important role in controlling the energy balance of the active layer. Further increases in the maximum thaw depth at the cutover site were inhibited by the thermal inertial effect of the larger amount of ice in the second spring after disturbance. Active layer thickness at disturbed forest sites seems to be maintained from the second year after the disturbance because of the thermal inertial effect mentioned above and the effect of recovering vegetation preventing energy exchange between the ground and the atmosphere. In conclusion, there is a strong effect preventing rapid deep thawing of the permafrost, a self-retention mechanism of the active layer thickness after forest disturbance in this continuous permafrost zone.

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