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

Go Iwahana[1,*], Takashi Machimura[2], Yoshikazu Kobayashi[3], Alexander N. Fedorov[4], and Masami Fukuda[5]

[1]*Graduate School of Engineering, Hokkaido University, N13 W 8, Sapporo 060-8628, Japan*
[2]*Graduate School of Engineering, Osaka University, Yamadaoka 2-1, Suita, 565-087, Japan*
[3]*National Institute for Agro-Environmental Sciences, Tsukuba, 305-8604, Japan*
[4]*Permafrost Institute, Russian Academy of Science, 18, Yakutsk, 677018, Russia*
[5]*Institute of Low Temperature Science, Hokkaido University, N19 W8, Sapporo 060-0819, Japan*

[*]*Corresponding author: go@lowtem.hokudai.ac.jp*

1. INTRODUCTION

The average annual precipitation in Yakutsk, Central Yakutia, in 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 continuously underlying permafrost. The ice-rich layer of permafrost prevents deeper infiltration of the limited rainwater and the soil moisture in the active layer is effectively stored for vegetation activity. On the other hand, forest vegetation acts as a thermal insulation buffer against heat exchange between the ground and the atmosphere. Thus, the coexistence of Permafrost and forest vegetation sustains the Taiga ecosystem (Fukuda et al., 1997 and Fukuda et al., 1984). Therefore, increasing occurrence of forest disturbances (Kakisawa, 2003), such as forest fires and clear-cutting, cause considerably destruction of the permafrost-surface vegetation coexistence relationship by altering ground surface conditions.

There have been several studies of climatic, physical, and biotic factors controlling the active layer and upper permafrost thermal status. Numerous studies focused on the role of fire in changes of the active layer thermal regime, and a wide range of responses were found depending on the site features (e.g., Burn, 1998; Mackay, 1995; Swanson, 1996; Viereck, 1982). Yoshikawa (2002) concluded that the impact of wildfires on the permafrost and ground thermal regime is dependent on burn severity, i.e., the degree of surface organic mat loss by fire. On the other hand, Linell (1973) showed experimentally that clear-cutting was effective for permafrost degradation in discontinuous permafrost zones. Damage induced by forest disturbance also alters the hydrological regime of the active layer. In interior Alaska, Sharratt (1998) concluded that clearing forestland would enhance soil drying and Ping (1987) noted that removal of vegetation cover changed the drainage of the active layer from poorly to moderately well-drained.

Changes in permafrost condition related to surface disturbance can be treated as changes in the energy and water balance characteristics of the surface. In permafrost studies, ground heat flux plays an important role in determination of active layer and upper permafrost thermal conditions. Ground heat flux is often the smallest component in energy balance equations of the surface and has attracted little attention in micrometeorological studies. This paper focuses on clear-cutting as a surface disturbance and the energy partitioning of the ground heat flux in the active layer. Our objectives are to describe overall effects of the clear-cutting on the thermal and hydrological characteristics of the active layer and to find changes in the energy

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partitioning of the ground heat flux immediately after the disturbance.

2. SITE DESCRIPTIONS

The research area, Neleger, is located about 35 km northwest of Yakutsk (62°19’N, 129°31’E), Russia, 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 about 239 mm, based on data from 1951-2000. The landscape on this terrace was almost flat and consisted of pure and mixed stands of larch, birch, pine, swamp, and thermokarst features called “Alas.” The research area was 215-221 m above mean sea level.

Two research sites, an intact forest site and experimental cutover, were set in this area. Site F was a mature larch stand (*Larix gmelinii*; over 200 years old) with a mean tree height of 8 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). The mean canopy height was approximately 15 m, and the distance from the measurement point to the nearest Alas was about 100-150 m. 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 cutover site, which was deforested deliberately in November 2000. The deforested area was a rectangle of 70 m × 170 m located about 100 m from Site F. The surface vegetation was disturbed during logging; however, the surface organic mat (live-plant-layer) remained undisturbed.

3. MONITORING OF ACTIVE LAYER

Overall field study was performed mainly during the snow-free periods from 2000 to 2003 and precise monitoring in the active layer was conducted at sites F and C from 1 May to 4 October in 2001 and from 20 April to 9 October in 2002.

Soil temperatures in these periods 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. For the temperature measurements, we used calibrated handmade temperature probes with thermistors (104ET, Ishizuka Denshi, Tokyo, Japan) at site F and calibrated temperature probes (107 probe, Campbell Scientific, Inc.) at site C. Soil moisture was measured with CS615 probes (Campbell Scientific, Inc.) by the TDR method 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 in the same soil profiles with temperature measurement.

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. Calibrations for TDR probes were conducted gravimetrically using 3-10 samples at each depth for each site.

The soil temperature measurements were conducted every 10 sec and stored every 10 minutes as averages. Soil moisture measurements were conducted every hour and recorded. All data were logged using a CR10X datalogger (Campbell Scientific, Inc.).

Soil samples were taken from layers every 0.05-0.1 m down to the maximum thaw depth in the excavated pits at each site during the period 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 were determined from the pits surveyed at each site.

The spatial distribution of the thaw depths was measured before the clearing treatment on a 150 m × 50 m grid set (856 points) at site C at the end of August 2000. The measurements were conducted on a 50 m × 50 m grid set (152 points) at site F and on a 150 m × 50 m grid set (295 points) at site C at the end of August 2001. Similarly, spatial measurements of thaw depths were conducted at site F (35 points) and site C (239 points) at the beginning of September 2002, and also at the end of August in 2003 at site F (121 points) and at site C (250 points). 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. However, as 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.

Net radiation was measured at each site using a 4-component net radiometer (CNR1, Kipp and Zonen, Saskatoon, SK, Canada). Calibrations were performed after observation and found to differ from each other by less than 1%. Radiometers were mounted at heights of 21m at site F and 9 m at sites C and B. 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 seconds and stored as 10-min averages using data loggers (CR10X, Campbell Scientific Inc., Edmonton, AB, Canada) *via* multiplexers. 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 in 2000 and at site C in 2001 and 2002.

4. ANALYSIS

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 for 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 adequately 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 analysis

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 that 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 considered 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 follows:

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

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

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

The soil compartment was divided into 5-7 layers associated with the depth at which temperature and water content probes were installed at each site. Q_A was calculated as in (3) where $\Delta T_j/\Delta t$ 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. Q_P was calculated by (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. k_f was estimated using the experimental equation of Kersten for frozen silty-clay (Kersten, 1949).

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. The properties of soil components and value of L_f used in this analysis are given in Table 1.

Table 1. Specific heat and density of the active layer materials at 0 °C used in the energy budget analysis

	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
		Values for T = 0 °C
Latent heat of fusion	0.333 (MJ/kg)	

4.2 Water balance analysis

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:

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

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 rainfall, 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).

Saturated moisture condition near the permafrost table, which was confirmed by boring surveys, supported the assumption of no infiltration into the lower layers of the compartment.

5. RESULTS AND DICUSSION

5.1 Climate condition and soil physical properties

Daily mean solar radiation (S_i) reached its peak value of about 30 MJ/m^2/day at the end of June and the values approached zero at the beginning of October in each year. The peaks of daily mean air temperature (T_a) were delayed by about half a month from S_i peaks in each year and daily mean relative humidity (RH) tended to increase from spring to fall. Meteorological measurements were compared as averages from June to July (S_i, T_a, and RH) and as total rainfall together with 50-year average (Table 2). Precipitation levels during the observations in each year were consistently smaller than the long-term average value. Especially, thawing seasons in 2001 and 2002 were extremely dry. Relatively high RH values in 2001 and 2002 also indicated dry conditions. Average air temperatures were higher in 2001 and 2002 and considerably lower in 2000. Based on this comparison, we treat year 2000 as relatively wet and cold conditions, and years 2001 and 2002 as hot and dry conditions. There were no marked differences between forest and disturbed sites in daily mean values through the thawing season.

Table 2. Weather conditions (S_i; solar radiation, T_a; mean air temperature, RH; mean relative humidity) at study sites. 50-year averages were are values in Yakutsk

Year	S_i (MJ/m^2/day)	T_a (°C)	RH (%)	Rain (mm)
	Jun - Jul			May - Sep
2000	20.5	15.7	68.2	128.3
2001	22.0	17.7	59.0	90.0
2002	20.0	17.2	58.5	79.0
50-yr ave.		17.1		159.2

Table 3. Surface organic mat thicknesses and particle size distributions of mineral soil in the active layer.

Site	Live - plant - layer (m)	Grain size		
		Sand (%)	Silt (%)	Clay (%)
F	0.1 - 0.2	21.0	59.8	19.2
C	0.1 - 0.2	39.9	47.0	13.1

The thickness of the surface organic mat and the average grain size characteristics in the active layer mineral soil are shown in Table 3. Although some spatial variations in the grain size distribution were found, the active layer mineral soils were classified as silty-clay soil at sites F and C. Little differences in the profiles of the bulk densities and organic contents were observed between sites F and C.

It was assumed that there were spatially little differences in the soil properties between sites F and C, because the two sites 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 similar at both sites. The moisture conditions in the soil will be discussed in the next section.

5.2 Comparisons of seasonal variations in soil temperature and moisture

Fig. 1 shows seasonal changes in the soil temperature profiles based on daily mean values. At site C, active layer thawing commenced at the beginning of May in 2001 when the snow had melted completely. Although there was a lack of temperature data at the beginning of May in 2001 at site F, we can infer that thawing at site F commenced on the same day as at site C as we observed snow disappearance at the temperature measurement points on the same day. The difference in soil temperature profiles at the beginning of the observation in 2001 was significantly large between sites F and C despite the close proximity of the measurement points in the same forest until the forest at site C was clear-cut. The temperatures at a depth of 1.6 m, for example, 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, as the surface conditions were obviously similar and spatial variability in a forest is unlikely to result in marked differences in soil temperature profile. This difference is probably due to intensive cooling at site C during the previous winter after experimental clear-cutting. The forest at site C was cleared in the previous early winter of the observation (during November 2000); the surface vegetation was completely removed and the accumulated snow was compressed during the clearing operation. The loss of vegetation and reduced thickness of snow at site C means loss of the thermal buffer layer, which prevents large heat loss from the ground to the air.

The progress of the thaw front at site C was faster than that at site F. The average thawing rates of frozen ground during June and July in 2001 at sites F and C were 0.010 and 0.015 m/day, 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 much colder soil temperature profile at the beginning of the observation, the maximum thaw depth determined from the temperature profile was significantly deeper at site C than at site F in 2001 (1.09 m and 1.19 m at sites F and C, respectively). As the thickness of the surface organic mat remained the same as before the disturbance at site C, the difference in maximum thaw depth is the net influence of forest vegetation loss on the active layer.

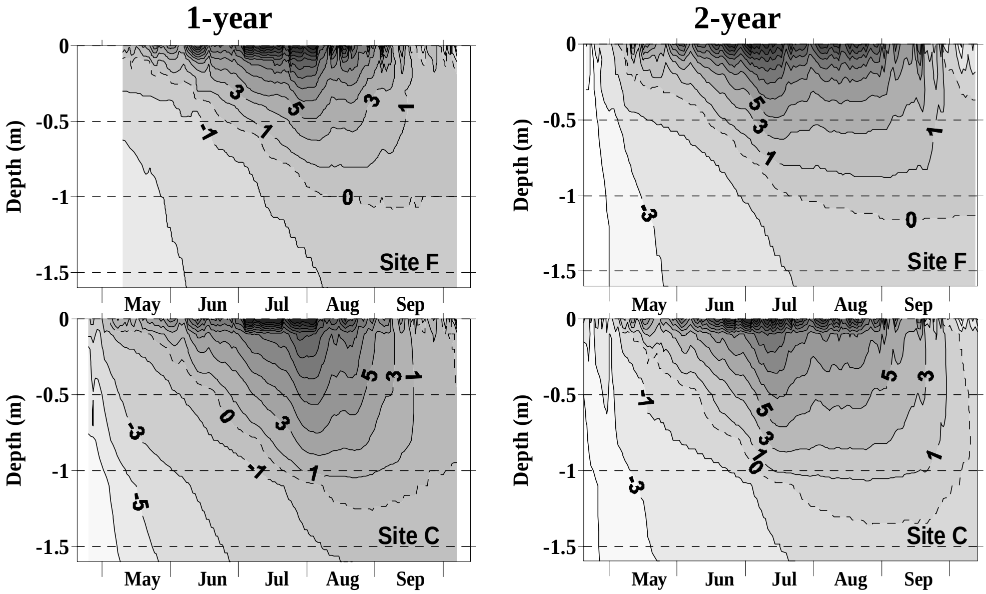

Fig. 1. Seasonal changes in the soil temperature profiles at sites F and C in 2001 and 2002 based on daily mean measurements. Temperatures are in degrees Celsius. Solid isotherms are drawn with increments of 2°C. The dashed line indicates the 0°C isotherm.

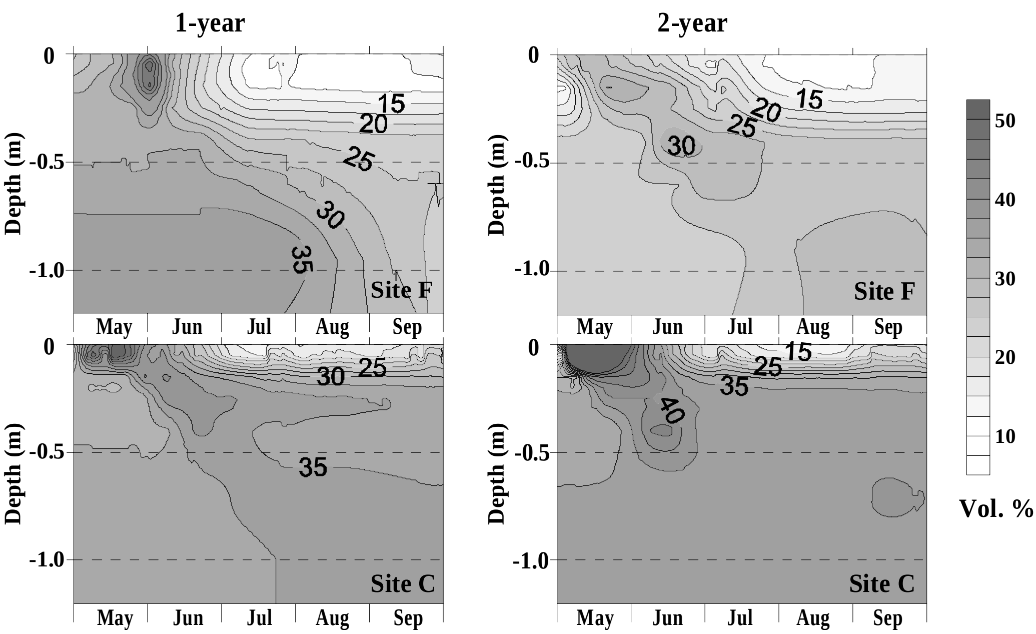

Fig. 2. Seasonal and successive changes in the soil water content profiles at sites F and C based on daily mean measurements. Water contents are shown as total water storage, including liquid water and ice, and in volumetric water content (%). Contour lines are drawn with increments of 0.25%.

In the thawing season of the second year after experimental clear-cutting (2002), the development patterns of the active layer showed almost the same tendencies at each site as in 2001. However, a relatively faster thaw down to a depth of 0.3 m was observed at site F, which may have been due to soil moisture changes in the previous summer. The changes in soil moisture profiles will be discussed below. At the beginning of May when thawing of the active layer commenced, the differences in the soil temperature profiles were small between the two sites. 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 were 1.16 and 1.28 m at sites F and C, respectively. The thaw depths at both sites increased from the previous year reflecting the changes in the physical soil conditions. We expected further increases in the difference in maximum thaw depth between the two sites, but the increase in the difference was only 0.02 m.

With regard to the soil moisture conditions, marked differences were observed in the first thawing season after clear-cutting (Fig. 2). Fig. 2 was drawn 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. Presuming similar soil moisture profiles at sites F and C, Fig. 2 displays the substantial water loss from the soil at site F and the retention of soil water at site C during the thawing season in 2001. The large difference in soil water content between sites F and C continued in 2002.

5.3 Spatial distribution of the active layer thicknesses

Although the measurements of the active layer thicknesses were performed during a period of few days in each year and the dates for each measurement period were different in each year (from the end of August to the middle of September), we treat the statistics from the measurements as values for the day when the thaw depths reached the maximum values. This is because changes in thaw depth during this period were very small (less than 0.01 m).

A comparison of the successive changes of the maximum thaw depths at sites F and C is shown in Fig. 3. The increase in 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 increase in the difference in these values (0.02 m), and in the third thawing season the difference decreased to 0.08 m. 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 the 2-3 years of this experiment.

The average values of thaw depths were somewhat different from those determined by point measurement of soil temperature profile. 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. Thus, we will focus on the differences in the characteristics of energy and water balance in the active layer between sites in later sections.

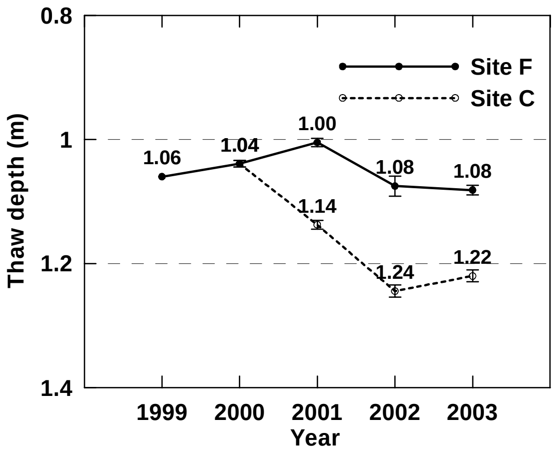

Fig. 3. Successive changes in the maximum thaw depths at sites F and C. Error bars denote the standard errors

5.4 Energy balance characteristics of the active layer

The absolute values of Q_S (=Q_A+Q_P) consistently accounted for relatively small portions of the energy balance components in the active layer during the observation period at both sites and years except in September at site F. The Q_A values, the energy consumed for temperature changes in the active layer, were positive from May to July as monthly averages and the average temperature in the active layer increased at both sites. 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 4). Q_L appeared to be the most important component controlling energy balance characteristics of the active layer in this region.

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, 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. 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. The active layer thicknesses at the study sites of Rouse (1984) and Boike *et al.* (1998) exceeded 1 m as in our study sites. The different characteristics of energy balance at the site reported by Yoshimoto *et al.* (1998) can be explained by the small amount of G during summer. It is likely that heat loss from the active layer started before G was consumed for the phase change of ice into liquid water.

The temporal trends of cumulative energy balance components in the active layer at sites F and C in 2001 and 2002 are plotted in Fig. 4. Comparison of the results from sites F and C indicated that the temporal trend of the cumulative values of Q_P at site C changed with the amounts being twice those at site F in both years. The cumulative Q_A values at site C were also consistently larger by about 1.5-fold than that at site F from May to September in both years. With regard to the cumulative values of Q_L, the magnitudes were similar until the end of July, then the difference between sites increased to 22 MJ/m^2 as a four-month total (Table 4) in 2001.

Table 4. Total sensible and latent heat storage changes in the soil (Q_S and Q_L), net radiation (Rn), and ground heat flux (G) over four months from May to August in 2001 and 2002 at sites F and C. Values are in MJ/m^2

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

(MJ/m^2)

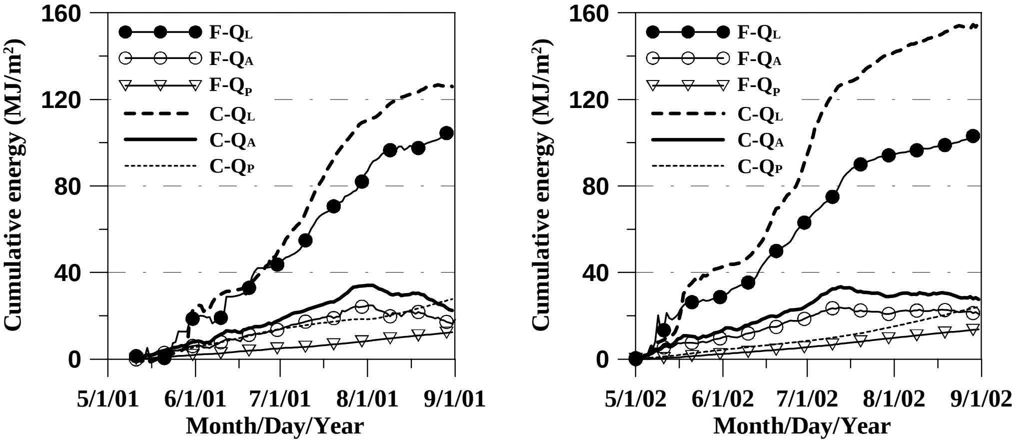

Fig. 4. Temporal trends of cumulative energy balance components in the active layer at sites F and C in 2001 (left) and 2002 (right). G is ground heat flux, Q_L is the storage of latent heat, Q_A is storage of sensible heat over the upper 1.5 m, and Q_P is the conductive heat transfer into the permafrost layer deeper than 1.5 m.

5.5 Water balance characteristics of the active layer

Water balance was analyzed from June to September in 2001 and 2002. Components of water balance in the active layer for this period are shown in Table 5 as monthly averages of daily totals. Completely different behaviors of the seasonal changes in ΔS and E were observed at sites F and C in the first thawing season after clear-cutting. ΔS changed markedly in June and July and large E from the forest was estimated in this period at site F. E decreased gradually toward the defoliation period at site F. In contrast, soil moisture was retained during the thawing season in 2001 and consistently smaller values of E were estimated at site C and there was much less seasonal variation in E as compared with site F. The strong restriction of evapotranspiration from the surface and forest vegetation at site C, which was found from the latent flux measurement by the eddy correlation method (Unpublished data; *Kobayashi et al.*), was confirmed from the observation of storage changes of the soil water in 2001. Total storage changes of soil water at sites F and C over the four months from June to September were -143.2 and -2.3 mm, respectively (Table 6). Consequently, much less of the total E at site C was estimated (Table 6): the difference in E values between sites F and C was 140.9 mm over the same period.

In the second thawing season after clear-cutting, estimated E values at site F were considerably decreased despite the very dry and hot weather conditions. Due to the very dry conditions in the soil layer where vegetation can take up the soil water (Fig. 2), rainwater was consumed immediately by evaporation from vegetation and the forest floor or transpiration by forest vegetation at site F in June, August, and September.

A relatively large E value was still found at site F in July. As the thaw depth proceeded from 0.3 to 0.7 m and from 0.7 to 1.1 m in June and July, respectively, it is possible that trees at site F could use the soil water efficiently from the layer where sufficient soil water existed in July. 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 (Fig. 2) and evaporation from the surface was enhanced by the hot and dry weather conditions. From July,

Table 5. Monthly averages 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. Values are in mm/day

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

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 the surface organic layer was also observed previously 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 (Table 6).

Table 6. Total precipitation (P), water storage change (ΔS), and evapotranspiration (E) over four months from June to September in 2001 and 2002 at sites F and C. Values are in mm.

Year	Site	P	ΔS	E
2001 (1-year)	F	68.8	-143.2	212.0
	C	68.8	-2.3	71.1
2002 (2-year)	F	73.3	-12.5	85.8
	C	73.3	-23.9	97.2

5.6 Self-retention mechanism of active layer thickness

The disturbance of the forest strongly altered soil thermal and hydrological conditions. However, the dynamic changes in the differences in temperature and water content profiles were limited only to the first thaw season after the forest 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 disturbed site will approach that in the forest in this study, as surface vegetation recovers. Inspection of the changes in temporal trends of cumulative energy balance components in the active layer from 2001 to 2002 (Fig. 4) 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: *i.e.*, there was large difference in ice content in the active layer during freezing season before the second thawing season.

The primary influence of forest vegetation loss was an increase in water storage ability at disturbed sites. Loss of vegetation will alter the water budget in the active layer, resulting in a significant increase in ice content at disturbed sites as compared with the forest at the beginning of the next thawing season. 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 and thus the retention of soil water at the disturbed site was marked in this study. However, it is likely that net increase of soil moisture will occur after most forest disturbance in most cases in this continuous permafrost region and the sudden increase in water storage ability can acts as a buffer for permafrost degradation. The excess ice content in the active layer requires much more latent energy of fusion at disturbed sites. That is, the excess ice content has a thermal inertial effect that delays 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 estimated 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, as the recovering vegetation makes the site's surface high reflector of solar radiation and it makes the surface ground heat flux smaller (*Iwahana et al.,* 2003). Eventually, there is strong tolerance to rapid deep thawing of the permafrost as a self-retention mechanism of the active layer thickness after forest disturbance in this ecosystem.

6. SUMMARY AND CONCLUSIONS

Thermal and hydrological conditions in the active layer were investigated simultaneously at a mature larch forest and 1-year experimental cutover site during the thawing season in 2001 and the investigation was continued in the thawing season in 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 1-year cutover site due to larger ground heat flux.
2) There was a marked increase in the difference in soil water storage caused by the strong restraint of evapotranspiration at the 1-year cutover site.

Marked changes in the active layer conditions were limited only 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, the 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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