Sublimation from snow surface in southern mountain taiga of eastern Siberia

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[1] Through intensive observations of snow sublimation and meteorological elements, cold season processes at the snow-atmosphere interface were clarified for both forested and open field conditions in the taiga region of eastern Siberia. Sublimation from snow surfaces differed with atmospheric stability. During early spring a significant difference in snow sublimation was observed between slopes and valley bottoms, despite variable vegetation cover. However, during the observation period, only one episode of significant snow sublimation was observed, which was caused by strong wind and ensuing light snowmelt. The sublimation for that week was ~50% of the sublimation observed in the study period (29 days). As atmospheric stability decreased, the effect of forest cover on snow sublimation was clear, with a significant difference between forested areas and open fields. Later in the spring, increased net all-wave radiation did not lead to an increase in sublimation but was consumed in meltwater production. The effect of forest cover on snow sublimation can be seen from the estimated bulk transfer coefficient for latent heat. The bulk transfer coefficient was larger for a larch forest than was that of open site. In the period 13 March–22 April (48 days), total snow sublimation was 15.7, 12.1, and 10.4 mm for open field, larch forest, and larch forest on a slope, respectively. This represents 14.3%, 13.0%, and 7.6% of the maximum water-equivalent snow cover, respectively. However, the study period was only a part of the entire snow season, so totals for the whole winter would be larger. *INDEX TERMS:* 1818 Hydrology: Evapotranspiration; 1863 Hydrology: Snow and ice (1827); 1878 Hydrology: Water/energy interactions; *KEYWORDS:* snow sublimation, taiga, Siberia, forest

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1. Introduction

[2] Sublimation from snow surfaces has been identified as an important hydrological process at high altitudes and in high-latitude regions, involving complex mass and energy exchanges. In the Colorado Frontal Ranges, measurements from snow evaporation pans indicated that net sublimation for the 5 month winter period from December to April was 135 mm [*Meiman and Grant*, 1974]. *Berg* [1986] estimated sublimation losses from snow cover to be 30–51% of precipitation over a 2 year period from 1973 to 1975. *Kattelmann and Elder* [1991] estimated sublimation from snow to be 18% of total precipitation over 2 years in the Sierra Nevada. In an extreme example, *Beaty* [1975] reported that sublimation was responsible for 80% of the ablation of fresh snow and for 60% of the ablation of older snow in the White Mountains of California.

[3] In the subarctic region, several studies have shown that sublimation from snow cover is a nonnegligible hydrological component that also affects river discharge and regional water resources. In western Canada, sublimation from snow during the winter season consumed from 15 to 40% of seasonal snowfall [*Woo et al.*, 2000] and 12–33% of annual snowfall [*Pomeroy and Li*, 1997]. *Suzuki et al.* [2002] estimated that sublimation from snow cover in eastern Siberia was significant at 25.6% of precipitation from October to April. However, these estimations were deduced from modeling and lack observational verification.

[4] On a global scale, subarctic ground surfaces are dominantly covered by subalpine and boreal forests. The impact of forests on snow cover has been extensively investigated via accumulation and melting processes. An increase of 30–45% in seasonal snow accumulation was measured after the removal of evergreen forest [*Pomeroy and Gray*, 1995; *Pomeroy et al.*, 1997]. *Pomeroy et al.* [1998] also found that snow water–equivalent (SWE) generally increases with evergreen canopy density in boreal forests. Simulation models for snowmelt under a forest canopy have been developed to examine the relationship between snowmelt and forest density [e.g., *Barry et al.*, 1990; *Yamazaki and Kondo*, 1992; *Wigmosta et al.*, 1994; *Hardy et al.*, 1997]. Canopy density is important in controlling snow ablation timing and rates because tree height and canopy properties control the transmission of solar

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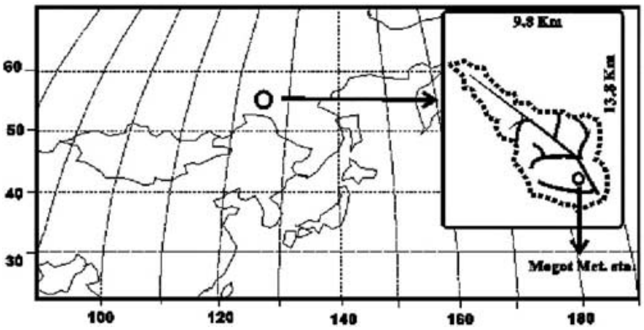

Figure 1. Location of the Mogot Basin.

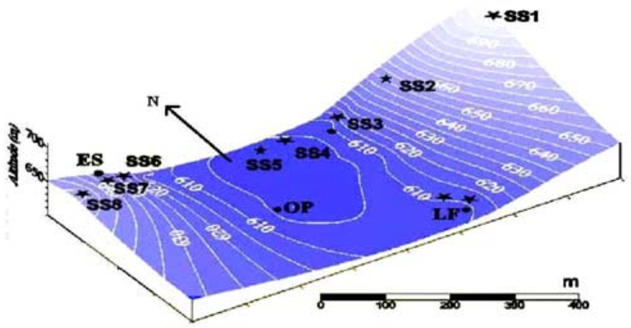

Figure 2. Location of observation points. LF, larch forest site; OP, open site; ES, larch forest site on east facing slope; SS1-8, snow survey point.

radiation [*Davis et al.*, 1997; *Ni et al.*, 1997; *Hardy et al.*, 1997].

[5] Previous studies of the effects of forest on snow sublimation have focused on the tops of forests, but few have addressed processes underneath the forest canopy. Much effort has been applied to model snow interception and sublimation by trees [*Hedestrom and Pomeroy*, 1998; *Pomeroy et al.*, 1998; *Darding and Pomeroy*, 1996; *Nakai et al.*, 1993]. However, understanding sublimation from snow-covered surfaces in forests and its role in the water cycle remains incomplete, although there has been recent progress in process analysis. In this study, differences in sublimation rates from snow surfaces in larch forest and open fields were examined, and the cause was investigated using data from intensive observations of snow sublimation and the heat budget made in eastern Siberia.

2. Observations

2.1. Study Area

[6] The Mogot experimental watershed is located in the southern mountain region of eastern Siberia (55.5°N, 124.7°E) in the Amur region, Russia (Figure 1). *Ma et al.* [2000] demonstrated that southeastern Siberia is the main water catchment for several important rivers in Siberia, such as the Lena, which experiences heavy rainfall, especially in the summer. Furthermore, the boundary between the continuous permafrost zone and discontinuous permafrost regions is located here. Therefore the area is anticipated to be very sensitive to climate change.

[7] In the basin, altitudes range from ~580 to 1150 m above sea level. The land surface is predominantly covered by larch forest, but birch forest partly occupies ridge areas, and higher elevations are dominated by pine forest. The basin is ~12 km long and 2.5 km wide, with an area of ~30.8 km^2; valley slopes face predominantly to the northeast and southwest.

[8] Between 1975 and 1993, discontinuous field observations of meteorology and hydrology were conducted in the study area [*Sokolov*, 1997]. The basin is characterized by variable climate; weather conditions greatly depend on the Siberian anticyclone. Table 1 compares climatic measurements in the snow-covered (October–April) and snow-free (May–September) periods, recorded by the meteorological station at Mogot during 1980. The annual mean air temperature was −7.5°C. with a range of 64.8°C. No significant difference in wind speed was noted between snow-covered and snow-free periods. About 80% of precipitation occurred in the snow-free period; annual precipitation was 524.2 mm, but average atmospheric moisture was not excessive.

2.2. Site Descriptions and Observations

[9] From March to May 2002, intensive observations were made in the Mogot experimental watershed in eastern Siberia. Three sites with "typical" surface conditions were selected to clarify the influence of forest on snow ablation, the role of snow cover in the regional water cycle, and parameterization of snow sublimation calculations in this subarctic region. Forest covers site LF, site OP is grassland, and site ES contains larch forest on a gentle east facing slope with 12° angle (Figure 2). The plant area index (PAI) was estimated using fisheye photographs. Values for site LF and ES were 0.50 and 0.45, respectively. Forest stand density and mean tree height were 4128 trees ha^{-1} and 4.3 m for LF, and trees were thicker than 5 cm in diameter at breast height, respectively. The understory of forest is about 10-cm-thick true moss and lichen. Snowmelt began around 4 April 2002, and the snow finally disappeared on 7 May 2002.

[10] Observations included meteorological elements, snow surveys, and sublimation measurements from the snow surface using pan method. At every site, automatic meteorological observations were made at 10 min intervals. The terms included air temperature, relative humidity, and wind speed, were measured at 1.6 m height, and net all-wave radiation was measured at 1.3 m height; all measurements were taken under trees at forest sites. Surface temperature was measured by infrared thermometer.

Table 1. Comparison of Meteorological Observations Between Snow-Covered and Snow-Free Periods of 1980

	Air Temperature, °C					
	Average	Maximum	Minimum	Wind Speed, m s^{-1}	Relative Humidity, %	Precipitation, mm
Snow-covered period (October–April)	−20.0	−0.8	−39.1	1.5	70	104.8
Snow-free period (May–September)	10.0	25.7	−4.4	1.4	73	419.4
Annual	−7.5	25.7	−39.1	1.4	71	524.2

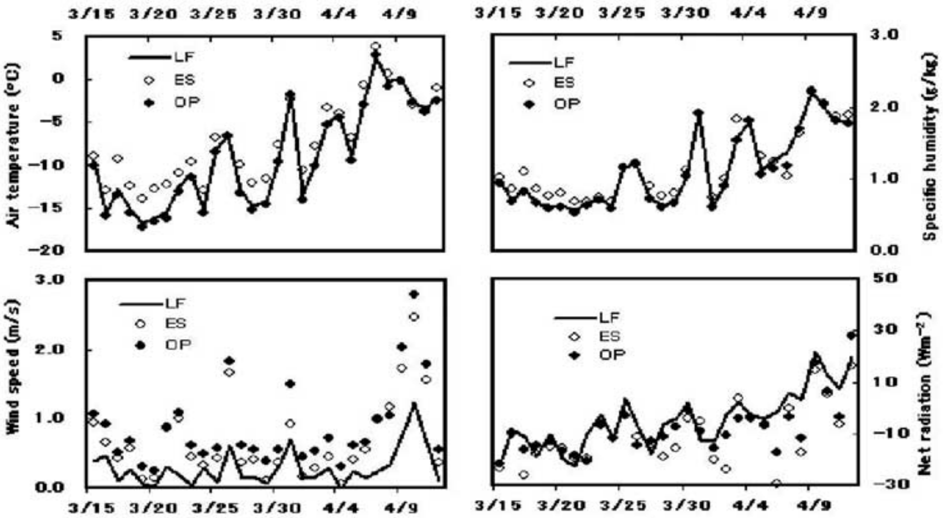

Figure 3. Variation of air temperature, humidity, wind speed, and net all-wave radiation for the forest (LF), east slope (ES), and open site (OP).

[11] The snow survey was carried out at line with eight stations (SS1-8 in Figure 2) at 1 week intervals. The observations included snow depth, snow temperature, and snow density profiles. Snow water equivalency was estimated from every profile. At the three selected sites, a snow survey was carried out everyday. To consider the spatial distribution of snow, snow depth was read at 16 points, some under the trees but some in the clearing.

[12] At every site, two transparent plastic pans were used for sublimation measurements. The size of the pan was 22 cm in diameter and 20 cm deep. A local snow sample of the same size and shape as the pan was set in the pan, which was placed 1 cm higher than the surrounding snow surface. The weight difference after a time interval, usually 3 hours, indicated the mass sublimated in the period. Sublimation in millimeters could then be calculated given the surface area of the pan.

[13] The snow sample in the pans was renewed when its temperature differed from the surrounding snow and was normally renewed within 24 hours, even when there was little temperature difference as compared with the surrounding snow.

3. Results and Analysis

3.1. Meteorological and Radiation Conditions

[14] The daily mean values of air temperature, specific humidity, wind speed, and net all-wave radiation as measured at the three study sites are shown in Figure 3. The most consistent observation between sites was the humidity. In the period 15 March–12 April, specific humidity averaged 1.1, 1.2, and 1.1 g kg^{-1} for sites LF, ES, and OP, respectively.

[15] The net all-wave radiation flux at site LF, located in larch forest, showed larger values than those of the other two sites. This was due to the effect of forest cover on downward long-wave radiation, which has been demonstrated to increase with forest density [*Suzuki et al.*, 1999]. The average net all-wave radiation for sites LF, ES, and OP were −4.5, −10.5, and −8.1 Wm^{-2}, respectively. Significantly, the daily mean net all-wave radiation flux was negative before 6 April and became positive after that date.

[16] Significant differences in air temperature among the three sites can be seen before 6 April. The mean air temperature for 15 March–6 April at site ES was higher than at site LF by 2.0°C and higher than at site OP by 2.4°C. This suggests that an inversion layer might exist above the valley floor, even though the difference in altitude between sites OP, LF, and ES was only 30 m. In the Mogot basin, an inversion layer has been reported, especially in during periods of snow cover, and the annual mean vertical lapse rate can attain 2.5°–3.0°C per each 100 m [*Sokolov*, 1997]. The stability of atmosphere should be evaluated by the lapse rate of the air temperature. Unfortunately, a sonde observation was not carried out in this work.

[17] Surface temperature was measured at all sites in this work. The difference in measured air temperature and surface temperature ($T_{air} - T_{sur}$) might be an indirect evidence of temporal variation of atmospheric stability over study sites. Examples of $T_{air} - T_{sur}$ both on 21 March and 9 April are show in Figure 4 on a 10 min basis. In an earlier observation period (21 March) the air temperature was significantly higher than the surface temperature for the whole day at all sites. The value of $T_{air} - T_{sur}$ on 9 April was clearly smaller than that on 21 March and even became negative in the daytime.

[18] In a snow-covered region the parameters determining the difference in wind speed between forested and open areas should include atmospheric stability as well as forest density. The wind speed observed at site LF was generally smaller than at the other two sites, but a more significant difference was found after 6 April, which coincided with air temperature variation. Average wind speeds at sites LF, ES, and OP from 15 March to 6 April were 0.2, 0.5, and 0.7 m s^{-1}, respectively, but were 0.6, 1.4, and 1.5 m s^{-1}, respectively, for the period from 7 to 12 April.

3.2. Snow Cover

[19] Snow cover, as a landscape element, is characterized by its seasonality. Its spatial and temporal variability are

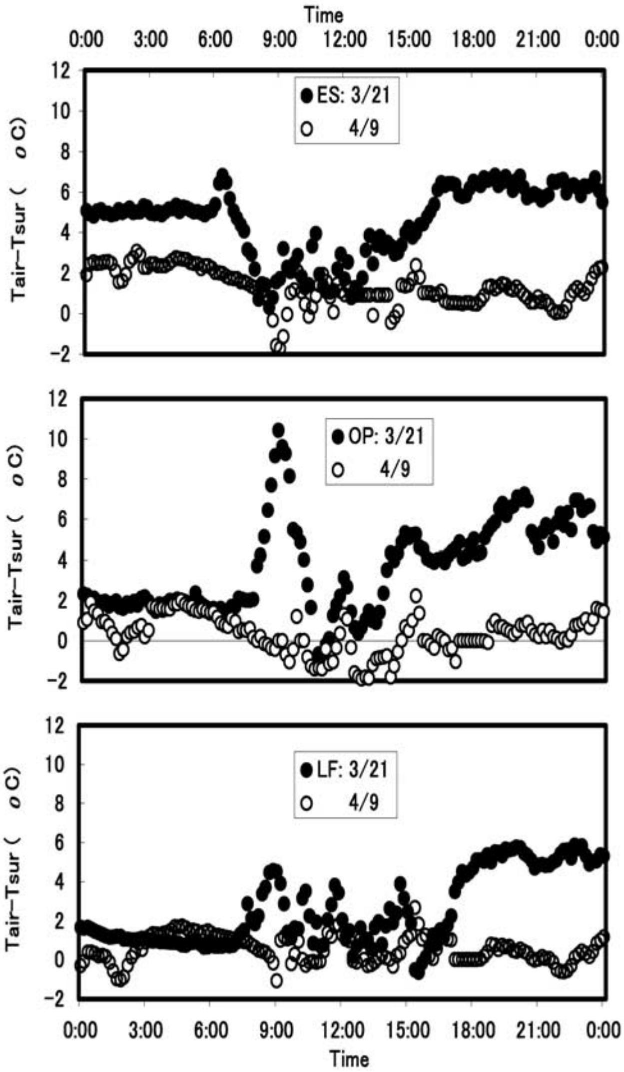

Figure 4. Diurnal variation of difference between air temperature and surface temperature ($T_{\mathrm{air}} - T_{\mathrm{sur}}$) on both 21 March and 9 April at all study sites.

high, even over a small spatial scale such as the repeated profiles sampled at nearby locations within each of our study sites. Figure 5 shows the snow depth and SWE observed at all sites for the study period (8 March 8–1 May 2002). The heterogeneity of snow distribution, both in depth and SWE, are clear between 21 March and 12 April. At site ES (larch forest on the slope), snowfall on 5 April caused a significant increase in snow depth and SWE, leading to a maximum of 70 cm and 146.0 mm, respectively. However, a light decrease can be seen at the LF site, which was caused by melting.

[20] Snowmelt began around 4 April, and the snow finally disappeared on 7 May. Before snowmelt started, both snow depth and SWE at sites LF and OP were similar. On 4 April the SWE was 94 mm for site LF and snow depth was 44 cm, while SWE was 102 mm for site OP, with a snow depth of 52 cm. Significant differences were seen in snow depth and SWE between sites LF and OP once melting commenced. This shows the effect of forest on the snow melting process, which has been reported extensively [*Price and Dunne*, 1976; *Ohta et al.*, 1993; *Hardy et al.*, 1997; *Suzuki et al.*, 1999; *Woo and Giesbrecht*, 2000].

3.3. Daily Sublimation Result by the Pan Method

[21] Two pans were used at every observation site. Table 2 presents statistical values of the results and variance between two pans at every site, respectively. The mean values were quite close, but a big difference can be found between their maxima. The variances, which were larger than 0.92 at all sites, imply the reasonability of the results.

[22] The results of daily sublimation from the snow surface at our three study sites are shown in Figure 6 for 15 March–12 April. Before 6 April, sublimation was higher at the bottom of the valley than on the east slope, combined with modest differences in wind speed between forest and open field and a more stable atmosphere. From 15 March to 5 April the total snow sublimation was 6.2, 6.8, and 4.8 mm at sites LF, OP, and ES, respectively.

[23] After 6 April the effect of forest cover on snow sublimation became notable. This was coincident with significant differences in wind speed between forested areas and open fields and a decrease of stability of atmosphere near the ground. From 15 March to 12 April 2002 the total snow sublimation was 9.3, 12.4, and 9.0 mm at sites LF, OP, and ES, respectively. The total sublimation was larger for the open field site than for forest sites by 33–39%.

3.4. Estimation of Bulk Transfer Coefficient

[24] The evaporation rate E (g cm^{-2}) can be calculated by the bulk formula

$$E = \rho\, Ce\, (q_S - q_Z) U_Z, \quad (1)$$

where ρ is air density (kg m^{-3}), Ce is the bulk coefficient for latent heat transfer to the snow surface, and q and U are specific humidity (kg kg^{-1}) and wind speed (m s^{-1}), respectively. The subscripts Z and S mean "at reference height" and "on the snow surface," respectively. When automatic meteorological observation data are used in the calculation, equation (1) can also be expressed [*Suzuki et al.*, 1999] as

$$E \cong \rho\, Ce\, [(1 - \mathrm{rh}/100) q_{\mathrm{SAT}}(T_S) + \Delta(T_S - T_Z)] U_Z, \quad (2)$$

where rh is the relative humidity (%), $q_{\mathrm{SAT}}(T_S)$ is the saturation-specific humidity at temperature of T_S (°C), and $\Delta = dq_S/dt$ at T_Z (°C).

[25] Using our observations of snow sublimation and meteorological elements, the bulk coefficient Ce was determined from correlation of daily sublimation (E) to the value of $\rho(q_S - q_Z)U_Z$ (Figure 7). Daily sublimation (E) varied significantly in a linear relationship with $\rho(q_S - q_Z)U_Z$.

[26] There was a large difference in Ce between LF and the other two sites (0.0053 versus 0.0020 and 0.0019 for sites LF, ES, and OP, respectively). *Suzuki et al.* [1999] reported results of Ce = 0.0031 for similar larch forest with PAI of 0.48 and Ce = 0.0023 for an open site in northern Japan.

3.5. Calculation of Sublimation

[27] Using the estimated Ce, snow sublimation can be calculated from meteorological data. Our statistical analysis

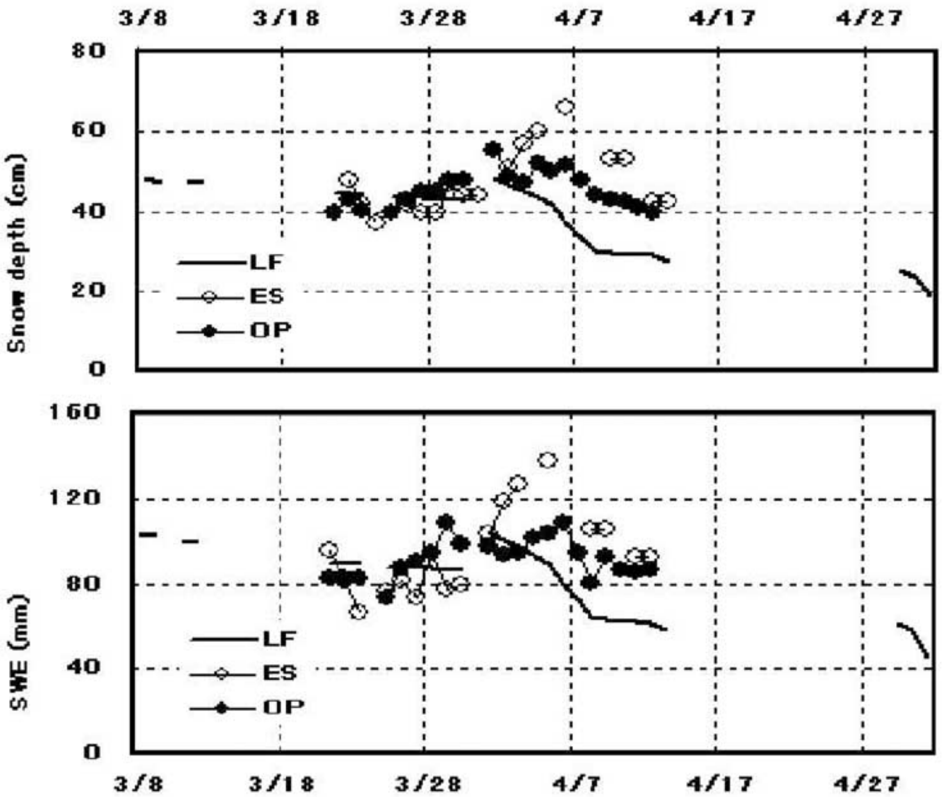

Figure 5. Snow depth and snow water equivalence at our study sites from 8 March to 1 May 2002.

of the observed and calculated daily snow sublimation for the calibration period (15 March–12 April 2002) is summarized in Table 3. The mean calculated sublimation appears to be somewhat smaller than that observed (MOS). The mean absolute error ranged from 0.01 to 0.03 mm d^{-1}, 2–7% to MOS. Thus the error in the calculation results using estimated *Ce* is no more than 7%. The standard error of the calculation (SE) ranged from 0.02 to 0.05 mm d^{-1}, and the variances (R^2) were 0.88–0.97.

[28] The results of estimated sublimation for sites OP, LF, and ES are shown in Figure 8. The calculation was also extended to the period without sublimation observations. Generally, the daily sublimation for the three sites varied synchronously. Minor variations occurred under inversion conditions, but snow sublimation revealed a significant peak from 5 to 12 April, the period when snow started to melt at all three study sites. At this time, surface snow was observed to be wet, caused by incipient melting that did not percolate as meltwater but instead adhered to snow grains, increasing the evaporation rate. Sublimation from 5 to 12 April was about 53% of the total sublimation in the entire observation period (15 March–12 April). After 12 April, sublimation decreased at all sites because most of the incoming heat was apparently consumed in meltwater production. The latent heat flux at snow surfaces becomes low once intensive melting begins [*Ohta et al.*, 1993; *Suzuki et al.*, 1999].

[29] In the period 13 March–22 April (48 days), snow sublimation totaled 15.7, 12.1, and 10.4 mm for sites OP, LF, and ES, respectively. This amounts to 16.4%, 14.8%, and 8.1% of the maximum SWE for each site, respectively. The calculations cover the period from early March almost to the time that the snow finally disappeared but do not extend to any prior snow cover period (early October to 12 March). *Suzuki et al.* [2002] estimated that total sublimation from snow cover between October and April in eastern Siberia was 25.6% of all precipitation.

4. Discussion and Concluding Remarks

[30] Through intensive observations of snow sublimation and meteorological elements the cold season process at the

Table 2. Statistical Values of Sublimation Observation Between Two Pans at All Sites

	Site OP		Site LF		Ste ES	
	Pan OP1	Pan OP2	Pan LF1	Pan LF2	Pan ES1	Pan ES2
Maximum, mm d^{-1}	1.27	1.43	1.10	1.29	1.19	1.05
Minimum,[a] mm d^{-1}	0.00	0.01	0.00	−0.01	0.00	0.00
Mean, mm d^{-1}	0.42	0.43	0.32	0.32	0.32	0.30
Variance	0.97	0.97	0.92	0.92	0.93	0.93

[a]Negative value implies condensation.

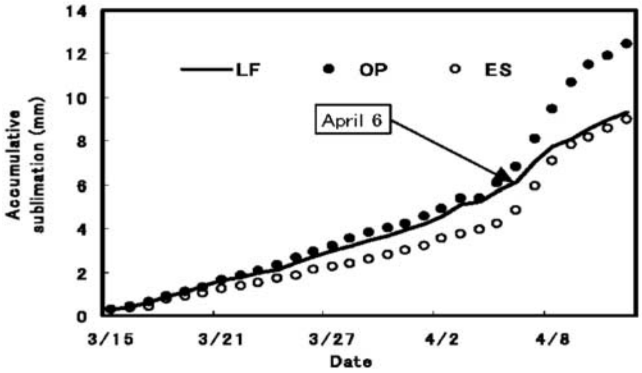

Figure 6. Accumulative pan-measured daily snow sublimation on LF, OP, and ES sites for the period from 15 March to 12 April 2002.

Table 3. Statistical Analysis of Observed (MOS) and Calculated (MCS) Daily Snow Sublimation at Our Three Study Sites in the Calibration Period (15 March–12 April 2002)[a]

Site	OP	LF	ES
MOS, mm d^{-1}	0.44	0.42	0.30
MCS, mm d^{-1}	0.43	0.38	0.28
MAE, mm d^{-1}	−0.01	−0.03	−0.02
SE, mm d^{-1}	0.02	0.05	0.02
R^2	0.97	0.94	0.88

[a]MAE, mean absolute error; SE, standard error of the calculation; R^2, variances.

snow-atmosphere interface was clarified for both forested and open areas of the taiga region of eastern Siberia in spring 2002. Sublimation from snow surfaces varied with atmospheric and ground surface conditions and plays a significant role in hydrological processes.

4.1. Temporal Variation of Snow Sublimation

[31] The temporal variation should be elucidated by investigation of the heat budget, which requires that heat energy be consumed in sublimation. When net all-wave radiation is negative, the energy for vaporization should come from the delivery of sensible heat to the surface, which is predominantly determined by atmospheric stability.

[32] An example of energy budget analysis was conducted for site OP, as shown in Figure 9. The energy budget at the snow surface can be expressed as (the flux is positive when come to snow surface)

$$Qn + Qh + Qe + Qm + Qg = 0, \tag{3}$$

where Qn is net all-wave radiant heat (W m^{-2}), Qe is latent heat (W m^{-2}), Qm is melting heat (W m^{-2}), which is estimated from snow survey data, Qg is conductive heat at the snow surface (W m^{-2}), which accommodates any small residual energy, and Qh is sensible heat (W m^{-2}), which is calculated as

$$Qh = C_P \rho\, C_H\, (T_S - T_Z) U_Z, \tag{4}$$

where C_P is the specific heat of air (J kg^{-1}), C_H is the bulk coefficient for sensible heat transfer to the snow surface, and other variables are the same as in equation (1). C_H is assumed to be the same as Ce [*Suzuki et al.*, 1999]. In the

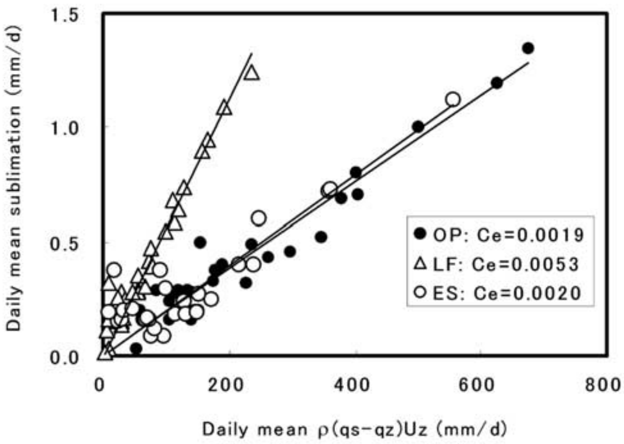

Figure 7. Comparison between evaporation and bulk formulae at the LF, ES, and OP sites.

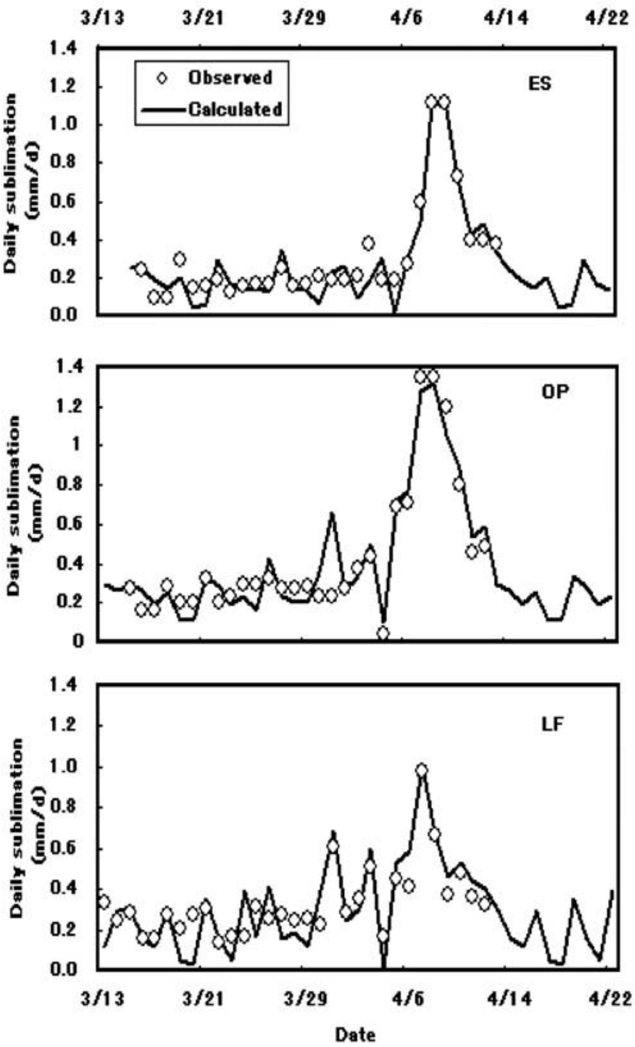

Figure 8. Calculation of daily sublimation for the three sites using the derived bulk transfer coefficient Ce.

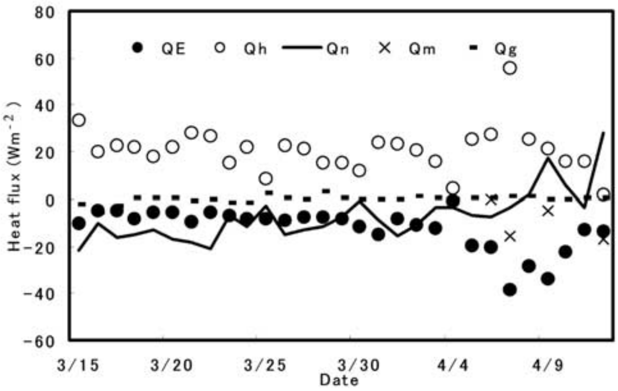

Figure 9. Variation in heat budget components at the open site from 15 March to 12 April 2002. *Qn*, net radiation; *Qh*, sensible heat; *Qe*, latent heat; *Qm*, heat for snow melting; *Qg*, conductive heat.

case when *Qn* is dominantly negative, *Qh* balances the energy loss from radiative transfer and vaporization processes. As *Qn* increases, both *Qn* and *Qh* can affect the variation of *Qe*. Therefore sensible heat *Qh* is an efficient heat source for snow sublimation, especially under an atmospheric inversion.

[33] The sensible heat flux, however, is determined by both wind speed and air temperature profiles, which relate to stability of the atmosphere. Unfortunately, we do not have direct evidence to evaluate stability of the atmosphere over study sites. However, the difference of surface air between at the slope and at bottom of the valley should partly, at least, indicate stability of the atmosphere. Therefore the following index is defined as

$$\omega = T_{\mathrm{ZES}} - (T_{\mathrm{ZLF}} + T_{\mathrm{ZOP}})/2, \tag{5}$$

where ω is the atmospheric inversion index (°C), T_{ZES} is the air temperature at site ES (°C), T_{ZLF} is the air temperature at site LF (°C), and T_{ZOP} is the air temperature at site OP (°C). The value of $T_{\mathrm{ZLF}} - T_{\mathrm{ZOP}}$ implies the mean temperature at bottom.

[34] Owing to the sensible heat flux, which, deduced to be an efficient heat source for snow sublimation, is determined by both wind speed and air temperature profiles, variation of the ratio of sensible heat flux to wind speed (Qh/U_Z), the value of sensible heat normalized by wind speed, may be reflected in atmospheric stability. The daily (Qh/U_Z) versus the atmospheric inversion index (ω) is shown in Figure 10. The increase of Qh/U_Z with ω reveals that the sensible heat becomes dominant for the heat budget as the inversion layer becomes stronger (large ω). In fact, the sensible heat flux in periods of inversion development (15 March–5 April) was averaged to be 20.1, 19.8, and 17.5 W m^{-2} for sites OP, LF, and ES, respectively. ES was located on a slope, but the other two sites were at the bottom of the valley. Observed sublimation in an earlier observation period was most closely correlated with atmospheric condition but not vegetation cover. Thus it appears that strong turbulent mixing is necessary to enhance the movement of water vapor along prevailing temperature gradients during negative net radiation. *King et al.* [1996] and *Hood et al.* [1999] have documented similar snow sublimation events on Antarctic snow cover and a continental season snow cover. In a broader sense, temporal variations of snow sublimation are likely driven by specific synoptic weather patterns, which affect the atmospheric stability of the study region. This hypothesis is supported by *Cline* [1997], who demonstrated a link between trends in overall snowpack energy balance and prevailing synoptic weather patterns in Colorado, United States.

[35] However, one major peak of snow sublimation was observed under neutral atmospheric conditions in this work. The sublimation from 5 to 12 April was about 50% of the total sublimation in the observation period (15 March–12 April). This rapid sublimation was coincident with strong winds and modest positive net all-wave radiation, which led to incipient melting of the snow surface. Under these neutral atmospheric conditions the effect of forest cover on snow sublimation was clear, with significant differences of sublimation between forested and open areas. Later in the spring thaw, increased net incoming radiation energy did not lead to an increase in sublimation but was consumed in meltwater production.

4.2. Spatial Variation of Snow Sublimation

[36] In an observation period of 29 days an examination of daily sublimation reveals two spatial patterns in the record. First, during early spring a significant difference in the snow sublimation rate was observed between sites on the slopes and at the bottom of the valley, despite variations in vegetation cover. This implies that the forest cover did not affect vaporization processes during atmospheric inversions. The second pattern evident in daily sublimation is that, at times, there appeared to be a larger sublimation observed at open site under neutral atmospheric conditions. Under these neutral atmospheric conditions the effect of forest cover on snow sublimation was clear, with significant differences of sublimation between forested and open areas.

[37] Forest should affect sublimation via the heat budget and climatic elements such as wind speed and humidity. From the observations shown above, such effects did not apply under stable atmospheric conditions, associated with negative net all-wave radiation flux. As the atmospheric stability decreased and net all-wave radiation became pos-

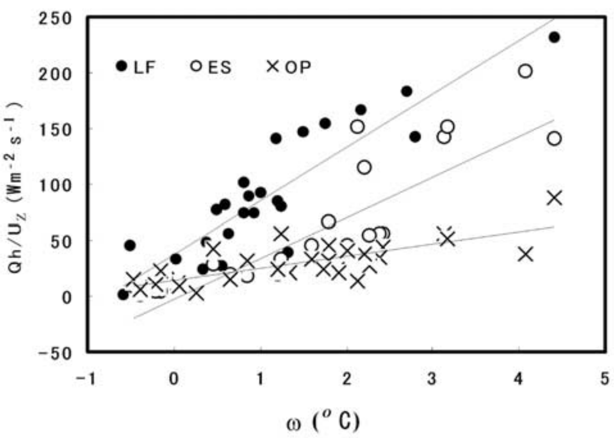

Figure 10. Variation of Qh/U_Z versus the atmospheric inversion index (ω).

Table 4. Interannual Changes of Maximum of Snow Water Equivalence (SWE), Precipitation, and Sublimation in the Period of 26 September to Next 29 April (120 Days) From 1978 to 1982 in Study Basin

	Maximum of SWE, mm	Precipitation *Pr*, mm	Sublimation *E*, mm	*E*/*Pr*
Period				
26 Sept. 1978–29 April 1979	85.3	115.8	20.2	0.17
26 Sept. 1979–29 April 1980	113.2	131.9	29.6	0.22
26 Sept. 1980–29 April 1981	48.3	67.0	16.3	0.24
26 Sept. 1981–29 April 1982	43.6	69.8	36.3	0.52
26 Sept. 1982–29 April 1983	120.5	139.6	18.7	0.13
Mean		104.8	24.2	0.26

itive, the effect of forest cover on snow sublimation became significant. At the same time, the difference in wind speed between forested and open areas became clear (Figure 3). Therefore wind speed still has a significant role in variations in sublimation.

[38] The effect of forest cover on snow sublimation can be seen in the parameterized bulk transfer coefficient for latent heat, which indicates actual evaporation efficiency. The bulk transfer coefficient for snow sublimation was found to be 0.0053, 0.0020, and 0.0019 for the larch forest, larch forest on slope, and open sites, respectively.

[39] The temporal variation of sublimation in the record, which shows that 50% of the total sublimation in the observation period was observed under neutral atmospheric conditions, implies the significance of vegetation affecting snow ablation processes in the study region. *Koivusalo and Kokkonen* [2002] documented a similar variation in southern Finland by significant difference of latent heat flux between forested and clearing sites.

4.3. Annual Changes of Snow Sublimation and Its Hydrological Consequence

[40] For the period 13 March–30 April (48 days), snow sublimation totaled 15.7, 12.1, and 10.4 mm for open grassland, larch forest, and larch forest on eastern slope, respectively, which represents 14.3%, 13.0%, and 7.6% of the maximum water-equivalent snow cover, respectively. However, the study period was only a part of the entire snow season, so totals for the whole winter would be larger.

[41] Interannual changes in snow sublimation originate from changes in climate variables, including air temperature, wind speed, and precipitation. Sublimation changes also reflect the role snow plays in the local-scale water cycle. A better understanding of the interannual changes in the snow sublimation coupled to changes in precipitation will improve the application of the results of this work. To reveal interannual changes of snow sublimation, the snow model of SNTHERM [*Jordan*, 1991] bas been applied in the study basin using data from Mogot meteorological station in the period of 26 September to next 29 April (120 days) from 1978 to 1982.

[42] SNTHERM is a process-driven, one-dimensional energy and mass balance model that calculates melt and other snowpack fluxes using mathematical equations based on known physical processes. SNTHERM is essentially an accounting program that distributes energy, mass, and momentum to and through the snowpack and underlying soil as a function of meteorological driving variables. SNTHERM utilizes a variety of microprocesses that all affect the energy balance of the snowpack. SNTHERM calculates the surface energy flux of the snowpack for each time step. Energy, mass, and momentum are distributed through the snowpack based on net fluxes of those same variables. SNTHERM grows and melts "layers or nodes" in the snowpack, grows snow grains, increases or decreases density as a function of both compaction and metamorphism, and decreases or increases the depth of the snowpack.

[43] Table 4 shows the interannual changes of the annual maximum of SWE, precipitation, and sublimation in the period of 26 September to next 29 April (120 days) from 1978 to 1982 in the study basin, which were calculated using SNTHERM. The largest snow sublimation was found

Table 5. Observations of Surface Snow Sublimation (*E*), Air Temperature (*Ta*), Wind Speed (*U*), and Saturation Deficiency (*De*) at Selected Sites

Observation Site						
Region	Surface Condition	Observation Period	*E*, mm d^{-1}	*Ta*, °C	*U*, m s^{-1}	*De*, hPa
Eastern Siberia	open field	March 2002	0.3	−12.7	0.8	1.4
		April 2002	0.7	−4.4	1.0	2.7
	larch forest on slope	March 2002	0.2	−10.1	0.6	1.5
		April 2002	0.5	−3.0	0.9	2.9
	larch forest	March 2002	0.3	−12.4	0.2	1.4
		April 2002	0.6	−4.1	0.4	2.7
Mongolia[a]	open field	Feb. 2002	0.1	−18.3	0.6	0.3
Eastern Tibetan Plateau[b]	open field	May 1989	0.7	−5.7	3.2	1.4
		June 1989	1.0	−1.8	3.7	1.1
Tianshan Mountains[a]	open field	Feb. 1991	0.2	−14.6	2.5	0.9
		March 1991	0.5	−10.5	2.3	1.3
		April 1991	0.9	−6.6	3.0	1.5
		May 1991	0.5	0.1	2.6	2.6

[a]From Y. Zhang et al. (Energy budget for active layer frosting of grassland underlain by warm permafrost in Mongolia, submitted to *Hydrological Processes*, 2004).
[b]From *Zhang et al.* [2003].

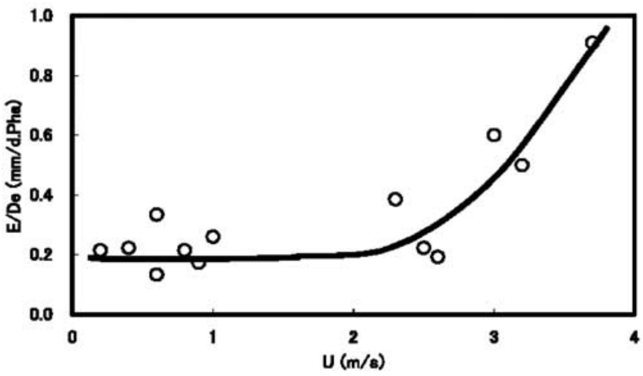

Figure 11. The correlation of *E*/*De* versus *U* (from Table 5).

in the winter of 1981/82, with a value of 36.3 mm coupled to precipitation of 69.8 mm and a maximum of SWE of 43.6 mm; the proportion of snow sublimation (*E*) to precipitation (*Pr*) was highest as well, with value of 0.52. Generally, no correlation of sublimation to precipitation is found. That proportion (*E*/*P*r) was variable from 1978 to 1983, with a range of 0.13 to 0.52. Consequently, the total water-equivalent in the snow cover at the site was reduced on average by 25% through sublimation, which will be reflected in runoff process in spring as well. That proportion (*E*/*P*r) on seasonal snow cover has been reported to be 20% by *Marks et al.* [1992], 18% by *Kattelmann and Elder* [1991], and 15% by *Hood et al.* [1999].

4.4. Sublimation From Eurasia Cryospheric Ground Surface

[44] In Table 5, the observed result of daily snow surface sublimation at the sites in this study has been compared to the result gained in the Eurasian cryosphere, including sites in the Tianshan Mountains, eastern Tibetan Plateau, Mongolia. Positive values are found at all sites; it may be deduced that the water flux between the atmosphere and snow surface in the Eurasia cryosphere is dominated by sublimation but not condensation, similar to the result achieved in New Zealand [*Moore and Owens*, 1984]. Sublimation values were higher in the eastern Tibetan Plateau than in other regions.

[45] Differences in the observed time series mean that the results at different sites are difficult to compare directly, but the spatial distribution could be investigated by relating monthly mean sublimation to wind speed (*U*) and saturation deficiency (*De*) as shown in Figure 11, in which the variation of *E*/*De* versus *U* is plotted. The purpose of the normalization of *E*/*De* in Figure 11 is to investigate the effect of wind speed to sublimation. On a monthly basis, snow sublimation correlates well to wind speed (*U*) and saturation deficiency (*De*). When wind speed is <2 m s^{-1}, saturation deficiency is predominant to determine sublimation, and sublimation increases significantly when wind speeds are <2 m s^{-1}. Actually, once wind speed is >2 m s^{-1}, sublimation from blowing snow must be taken into account [*Sugiura et al.*, 2001]. More than half of wind-transported snow can sublimate before reaching the field edge [*Pomeroy and Gray*, 1995]. The estimates indicate that sublimation from blowing snow can account for 2–20% of the annual snowfall amount [*Hood et al.*, 1999].

[46] In northern Eurasia, snow accumulation and ablation have a significant influence on hydrological processes. For the winter precipitation of 50–150 mm, snow sublimation is nonnegligible to a better understanding of hydrology in northern Eurasia. However, because of the difficulties associated with measurement in winter, detail observations of snow sublimation and subsequent estimation of water fluxes are limited in this area. The main barrier remaining to accurately estimating sublimation in previous, even in this study, is the lack of observed evidence to dress the factors that dominate variation of sublimation. The establishment of a comprehensive, season-long data set for sublimation in the spatial scale of the whole northern Eurasia cryosphere, including Siberia, Tibetan Plateau, and Mongolia, is an important step toward both understanding the role of sublimation in the subarctic hydrological cycle and estimating its importance in affecting the discharge to the Arctic Ocean. Further works looking at sublimation in lower snow years will reveal the extent to which sublimation rates vary on an annual basis and in a thin snow cover region like Mongolia and will reveal the extent to which sublimation varies on a continental scale.

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