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Sublimation from thin snow cover at the edge of the Eurasian cryosphere in Mongolia

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Abstract:

Sublimation from thin snow cover at the edge of the Eurasian cryosphere in Mongolia was calculated using the aerodynamic profile method and verified by eddy covariance observations using multiple-level meteorological data from three sites representing a variety of geographic and vegetative conditions in Mongolia. Data were collected in the winter and analysed from three sites. Intense sublimation events, defined by daily sublimation levels of more than 0·4 mm, were predominant in their effect on the temporal variability of sublimation. The dominant meteorological elements affecting sublimation were wind speed and air temperature, with the latter affecting sublimation indirectly through the vapour deficit. Seasonal and interannual variations in sublimation were investigated using long-interval estimations for 19 years at a mountainous-area meteorological station and for 24 years at a flat-plain meteorological station. The general seasonal pattern indicated higher rates of sublimation in both the beginning and ending of the snow-covered period, when the wind speed and vapour deficit were higher. Annual sublimation averaged 11·7 mm at the flat-plain meteorological station, or 20·3% of the annual snowfall, and 15·7 mm at the site in the mountains, or 21·6% of snowfall. The sum of snow sublimation and snowmelt evaporation represented 17 to 20% of annual evapotranspiration in a couple observation years.

KEY WORDS Mongolia; snow cover; sublimation

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INTRODUCTION

The climate of the Mongolia territory, locating in the periphery region of Eurasia cryosphere, has been demonstrated recently to have a strong relationship with the Indian monsoon and El Niño-southern oscillation (Kripalani and Kulkarni, 1999; Ye, 2001). Clear skies in winter due to high anticyclone dominance over Mongolia results in less snowfall. Snow contributes less than 20% to total annual precipitation, that is averaged over Mongolia as 160 mm (Batima and Dagvadorj, 2000). The average depth of snow varies from 0·5 to 25 cm, even if the duration of stable snow cover is 120–150 days in mountainous regions, 70–120 days in the eastern steppe and 30–60 days in the Gobi desert region (Batima *et al.*, 2005). On such thin snow cover, sublimation can therefore be expected to be proportionally high in relation to the annual snowfall and to be an important factor in the water cycle and climatic system. However, no reports have described the variation in sublimation in this region. Sublimation from snow cover in the target region must be addressed to improve understanding of the regional water cycle, with particular attention to situations where the snow cover is unusually thin. This is the central motivation of the present study.

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The magnitude of natural sublimation from the snow surface, by which snow changes directly to water vapour, has been discussed for many years. Kattelmann and Elder (1991) found that the annual proportion of sublimation to snowfall over two measurement periods was 18 and 33%, the latter occurring in a drought year with one-third of the normal annual snow cover and a shortened melt season. High sublimation rates were also observed in Canada (Hood *et al.*, 1999) and in the US (Marks *et al.*, 1992) with daily values of 2·35 and 2·17 mm, respectively. In western Canada, sublimation from snow during winter consumed 15 to 40% of seasonal snowfall (Woo *et al.*, 2000) and 12 to 33% of annual snowfall (Pomeroy and Li, 1997); Suzuki *et al.* (2002) estimated that sublimation from snow cover in eastern Siberia was significant, accounting for 25·6% of precipitation from October to April. Heat budget results for glacier surfaces also provide evidence of such deductions. Heat budget observations conducted on an ablation area of a glacier showed a low evaporation rate and even condensation (Zhang and Kang, 2000), but results near the equilibrium line of the glacier revealed sublimation of 0·37 mm day^{-1}, which contributed to 28% of annual glacier ablation (Zhang *et al.*, 1996).

These various results may be explained by environmental differences among study sites. Rising air temperature, by enhancing the vapour pressure deficit, support sublimation as long as snow is cold and snow-melt dose is not dominate. Due to atmospheric inversion caused by snow cover generally, increase rates of sublimation are achieved as wind speed increases the turbulent exchange, which provides energy available for the phase change (Zhang *et al.*, 2004b). Moisture flux over snow involves complex mass and energy exchanges and is not only determined by atmospheric conditions but also by surface conditions.

Sublimation as the moisture fluxes between the snow and the atmosphere are commonly calculated from measurements of the snow surface latent heat flux. The most common technique for measuring the sublimation by snow surface latent heat flux is the bulk aerodynamic or mean profile method (Moore, 1983). A more accurate method for calculating the snow surface latent heat flux is the aerodynamic profile method which requires the measurement of wind speed, temperature, and relative humidity at multiple heights above the snowpack (Hood *et al.*, 1999). The third, and most accurate, method for measuring latent heat fluxes is the eddy correlation method. However, this procedure requires a high frequency sonic anemometer that is too fragile to use for extended periods in an alpine environment (Hood *et al.*, 1999).

In the present study, tower observation data was used to examine sublimation from thin snow cover; sublimation results were calculated using the aerodynamic profile method and verified by eddy covariance observations. Observations, begun in 2002, were conducted at three sites representing a variety of geographic and vegetative cover conditions in Mongolia. Further, using data from meteorological stations, snow sublimation was estimated on a daily basis at two stations from 1980 to 2004. This data was used to assess whether snow sublimation rates differed between two land cover types, i.e. forest understory and grassland. Using the longer term data from the meteorological stations, seasonal and interannual variations in snow sublimation and their proportion to snowfall were also examined.

STUDY AREA AND METHODOLOGY

Site descriptions

Intensive observations were conducted at three sites. The flat plain (FP) site was located on sparse grassland at Nalaikh (47°42′N, 07°20′E; 1415 m asl), 40 km southeast of Ulan Bator (Figure 1). This site was on a fluvial plain in the large valley of the Tuul River; the topography at and around this site was very smooth. Annual precipitation from April 2003 to March 2006 averaged 167 mm, 79% of which fell during the warm season, from May to September. The annual mean air temperature was −3·8 °C, but air temperatures averaged 12·7 °C from May to September. The annual mean relative humidity was 61%. Specific humidity was 5·5 g kg^{-1} from May to September and 0·5 g kg^{-1} for the rest of the year. The prevailing wind was from the NW–ENE sector, with an annual mean wind speed of 2·6 m s^{-1}. The wind was stronger (mean wind speed of 3·2 m s^{-1}) during summer and weaker in winter.

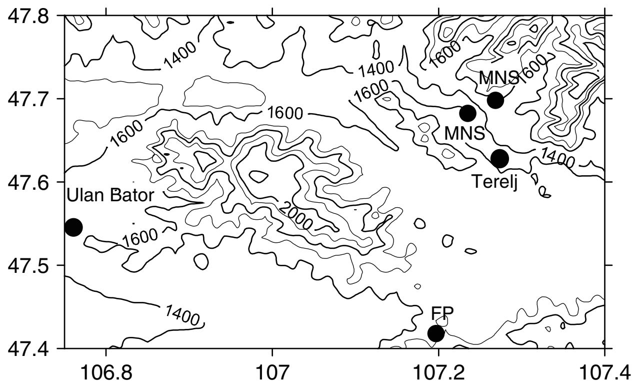

Figure 1. Locations of the study sites

The other two sites (MNS and MSS) were located in a small mountainous watershed of the Terelj River (47°68′N, 107°25′E; approximately 1640 m asl). From 1986 to 2000, the annual air temperature was −4·0 °C, and the annual precipitation was 354·9 mm, as recorded at a meteorological station near the study site. Site MNS was set on a north valley slope covered by forests dominated by *Larix* and *Pinus* species with mean height of 32 m. Site MSS was set on a southern slope, where there is grassland. Vegetation at sites FP and MSS is similar, with uniformly sparse grass with a coverage of 38–60% during the maximum growth period. Pasture areas had similar plant types and species. *Artemisia frigita* dominated, accounting for about 60% of the plant cover; other species included *Arenaria* sp. and *Leymus chinensis*. The maximum grass height in mid-July was less than 20 cm. In the snow-covered period, grass at the sites perished and was not apparent at the snow surface.

Permafrost underlies the study region and was found beneath both the northern forested slopes and the flat pasture plain. The permafrost was covered by a wet active layer on the gentle slopes and by a dry active layer on the plain; a dry active layer was also observed beneath the southern pasture slopes where permafrost was absent (Ishikawa *et al.*, 2004). Sharkhuu (2001) mapped the permafrost at regional and local scales and found the average thickness and temperature of continuous and discontinuous permafrost to be 50 to 100 m and −1 to −2 °C in the valleys and depressions, and 100 to 250 m and −1 to −3 °C in the mountains. The permafrost distribution also depends strongly on altitude.

Methodology

Automatic weather stations (AWS) were established at all three sites from November 2003 to March 2006. The AWS recorded air temperature, humidity, and wind speed at heights of 0·5, 1·0, 2·0, and 4·0 m. Short-wave and long-wave radiation were measured in both upward and downward directions at 2 m. In addition, sensors measuring air pressure and snow depth and an infrared radiative thermometer recording grass leaf temperature and net radiation were installed 1·5 m above the ground surface. Snow depth was measured by both sensor shown

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Table I. Instruments used in this study

Item	Unit	Instrument (model, manufacturer)	Record interval
Short-wave radiation	W m^{-2}	Radiometer (MS402, EKO, Japan)	10 min
Long-wave radiation	W m^{-2}	Infrared radiometer (MS202, EKO, Japan)	10 min
All-wave net radiation	W m^{-2}	Net radiometer (REBS Q7, REBS, Inc., USA)	10 min
Wind speed	m s^{-1}	Anemometer (AC750, Kaijo Corporation, Japan)	10 min
Wind direction	deg	Anemometer (VR036, Kaijo Corporation, Japan)	10 min
Snow depth	cm	Ultrasonic level-meter (SR50, Kaijo Corporation, Japan)	60 min
Precipitation	mm	Tipping bucket rain gauge (52 202, R. M. Young Co., USA)	10 min
Air temperature	°C	Humidity and temperature probe (HMD45D, Vaisala Oyj, Finland) with ventilation pipe (PVC-02, PREDE, Japan)	10 min
Relative humidity	%	Humidity and temperature probe (HMD45D, Vaisala Oyj, Finland) with ventilation pipe (PVC-02, PREDE, Japan)	10 min
Surface temperature	°C	Infrared radiation thermometer (CML303F, CLIM., Inc., Japan)	10 min
Heat flux in the soil	W m^{-2}	Heat flux plate (PHF01, REBS, Inc., USA)	10 min
Volumetric water content	m^3 m^{-3}	ECHO probe (EC-10, Decagon Devices, Inc. USA)	10 min
Soil temperature	°C	Pt-thermometer (TPT100S, CLIMATEC, Inc., Japan)	10 min
Air pressure	hPa	Analogue barometer (PTB101B, Vaisala Oyj, Finland)	10 min
Friction wind velocity	m s^{-1}	Ultrasonic Anemometer (Kaijo Corporation, Japan)	10 min
Latent and sensible heat	W m^{-2}	Ultrasonic Anemometer (Kaijo Corporation, Japan)	10 min

in Table I and daily manual observation of snow survey. From October 2005 to March 2006, wind profile, air temperature, humidity, and eddy covariance measurements were taken at MSS and MNS. Latent heat, sensible heat, and the friction wind (u^*) speed were measured at a height of 2·0 m. An open-path hygrometer/H_2O sensor (KH20, Campbell, USA) and a three-dimensional supersonic anemometer (81 000, YOUNG, USA) were used. Table I provides details of other instruments used in this study.

Snow sublimation (Es) was calculated for the three sites using the aerodynamic profile method, from the equation:

$$LEs = \rho L k^2 \frac{q_2 - q_1}{\phi_E \ \ln(z_2/z_1)} \frac{u_2 - u_1}{\phi_M \ \ln(z_2/z_1)} \qquad (1)$$

where k is the Kármán constant (0·4), L is the latent heat of vaporization of water, ρ is air density, u is wind speed, q is specific humidity, and z_2 and z_1 are the instrument heights, which were 2·0 and 0·5 m, respectively, during the observation period. Φ_E and Φ_M are stratification functions that depend on atmospheric stability through the Richardson number (Ri), determined by:

$$\mathrm{Ri} = \frac{g}{\overline{T}} \left(\frac{\delta T/\delta z}{(\delta U/\delta z)^2} \right) \qquad (2)$$

where g is the acceleration due to gravity, and $\overline{T}$ is the mean air temperature at two levels. ϕ_E and ϕ_M are calculated as described by Hood *et al.* (1999).

The principle difficulty with Equations (1) and (2) is that formulas were suitable as $\mathrm{Ri} < 0{\cdot}19$. The truth is, however, Ri was large when wind was rather weak. The calculation was filtered by criteria value of $\mathrm{Ri} = 0{\cdot}19$, the result of sublimation was set to be zero when Ri was larger than 0·19. This filtering may not cause significant bias due to vapour flux between atmosphere and snow surface must be very small in accordance to rather stable status of air mass. One more difficulty to determine snow sublimation on such thin snow cover (see later) is to assign the calculation to the reliability that snow exists, otherwise the computing results are from the vaporized amount from the soil surface but not from snow. Calculations were filtered by assessing snow surface albedo. The computation is filtered out when albedo is smaller than 0·46, which is the value of albedo that indicates significant snow melting.

Long-term sublimation estimation

For utilizability of routine data from meteorological station, the bulk aerodynamic was applied to estimating long-term snow sublimation (Suzuki *et al.*, 1999):

$$\mathrm{Es} \cong \rho \mathrm{Ce}[(1 - \mathrm{RH}/100) q_{\mathrm{SAT}}(T_{\mathrm{S}}) + \Delta(T_{\mathrm{S}} - T_{\mathrm{Z}})] U_{\mathrm{Z}} \qquad (3)$$

where Ce is the bulk coefficient for latent heat transfer to the snow surface. The subscripts ‘Z’ and ‘S’ mean ‘at reference height’ and ‘on the snow surface’, respectively. RH is the relative humidity (in %), $q_{\mathrm{SAT}}(T_{\mathrm{S}})$ is the saturation specific humidity at temperature of T_{S} (in °C) and $\Delta = dq_{\mathrm{S}}/dt$ at T_{Z} (in °C). Equation (3) is theoretically based on bulk aerodynamic formula described by Moore (1983). This method has the advantage of requiring meteorological measurements at only one height above the snow surface. However, a primary assumption of the bulk profile method, that snow surface temperature effectively tracks the air temperature, is often inaccurate below 0 °C and will result in the over-estimation of sublimation (Bernier and Edwards, 1989). Therefore, ‘estimation’ instead of ‘calculation’ is used.

Data from the Ulan Bator meteorological station (UB, 47°55′N, 106°52′E, 1300 m asl) and the Terelj meteorological station (47°67′N, 107°35′E, 1530 m asl), observed at 3-h intervals, were used for long-term estimations of snow sublimation. Wind measurements were made at the standard height used in Mongolia of 10 m. Wind speeds at the height of the observation instrument in this study

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(2 m) were estimated from the standard height measurements and a logarithmic wind profile. As precipitation gauges tend to underestimate true precipitation amounts, the precipitation data from the earlier stations were corrected using the procedure of Zhang *et al.* (2004a), which implies a bias of 17%. The bulk coefficient Ce was fixed to be 0·0020, which was deduced from the observed result of sublimation to the value of $\rho(q_S - q_Z)U_Z$ (not shown). The value is consistent with the results gained in Japan and Siberia (Suzuki *et al.*, 1999; Zhang *et al.*, 2004b).

RESULTS AND ANALYSIS

Meteorological and radiation conditions during the snow-covered period

Figure 2 shows the daily mean values of air temperature, humidity, wind speed, and net all-wave radiation as measured for the snow-covered period at the three study sites. From November to March in 2003 to 2006, for snow-covered periods, the relative humidity averaged 64, 59, and 63%, wind speed at 2 m was 2·5, 1·4, and 0·4 m s^{-1}, and the air temperature was $-18·5$, $-12·9$, and $-15·0\,°C$ at sites FP, MSS, and MNS, respectively.

Significant differences in air temperature among the three sites were found in mid-winter, from the beginning of December to mid-February. The mean air temperature for 1 December 2003 to 15 February 2004 at site FP was lower than that at site MSS by 8·2 °C and lower than that at site MNS by 7·5 °C. This suggests that a strong inversion layer may have been present above the study region, even though the difference in altitude among sites MSS, MNS, and FP was only 250 m. A similar strong inversion layer has been reported during the snow-covered period in eastern Siberia (Zhang *et al.*, 2004b). While it would have been desirable to assess atmospheric stability by measuring the lapse rate of air temperature, no radiosonde observations were made during this study. To clarify atmospheric stability during the study period, the Richardson number was calculated using Equation (2) and the atmospheric stability was evaluated following the classification method of Hood *et al.* (1999). Figure 3 presents the frequency of stable,

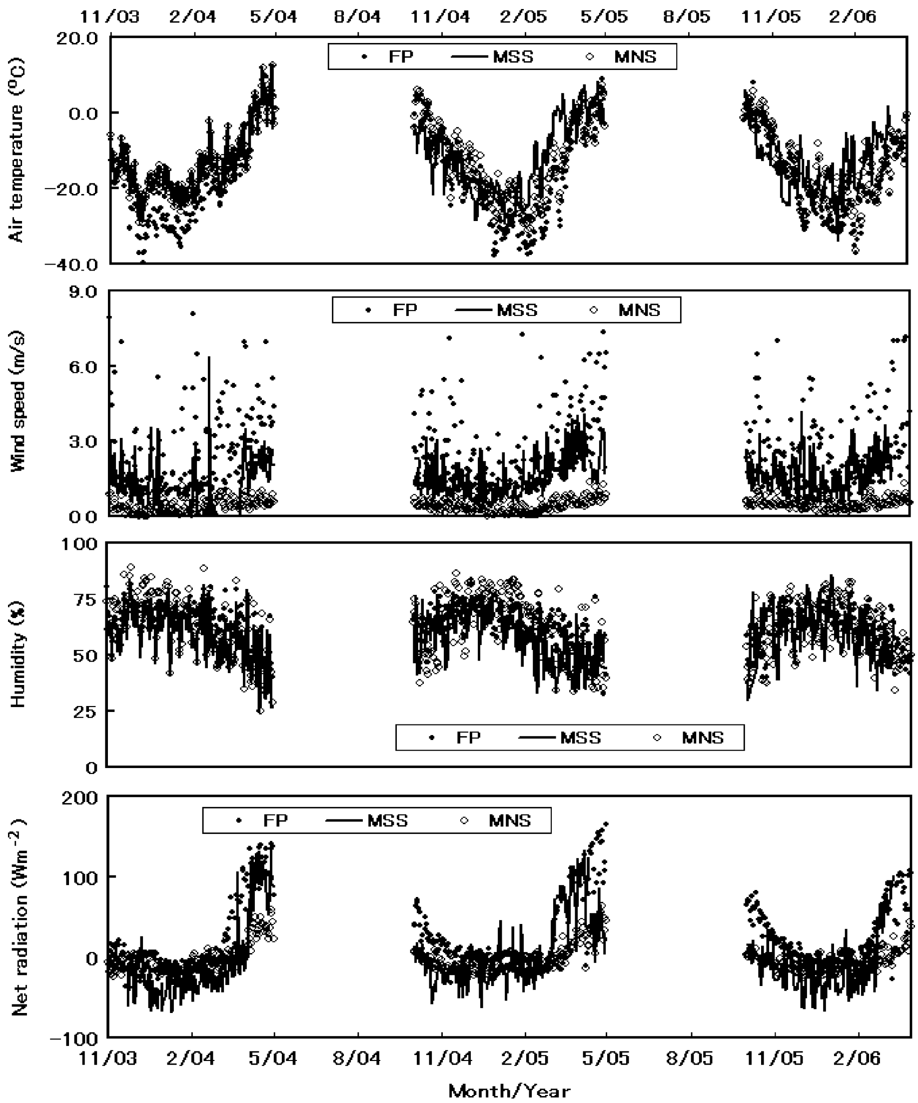

Figure 2. Variation in daily values of climatic elements at the three sites during the snow-covered period from November 2003 to March 2006

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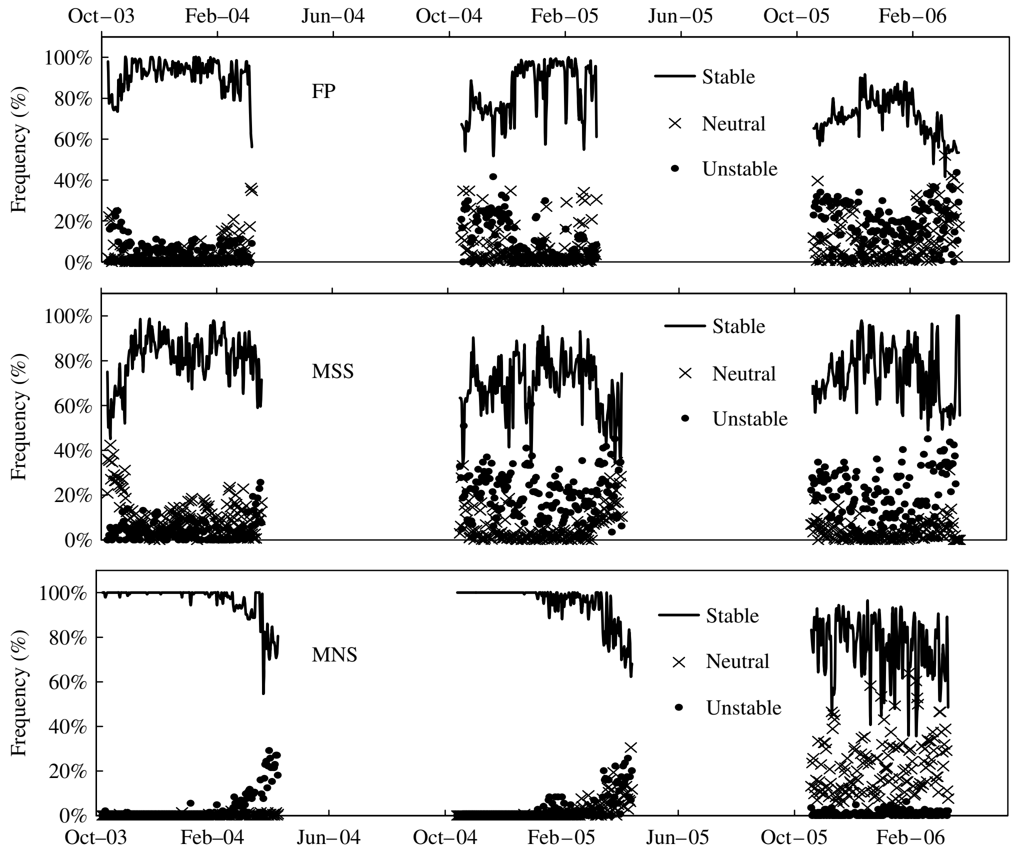

Figure 3. The frequency of atmospheric stability types over the study sites in the study period

neutral, and unstable atmospheric conditions on a daily basis, accounting for all data for each site during the study period. Stable and neutral conditions predominated over the study sites during the study period.

The net all-wave radiation flux at site MNS, located in a forest, showed similar values to net all-wave radiation at the other two sites before the beginning of April. However, after that time, results differed significantly because of the effect of forest cover on downward long-wave radiation, which has been demonstrated to increase with forest density (Suzuki *et al.*, 1999). In the period of snow cover, the average net all-wave radiation for sites FP, MSS, and MNS was $-1{\cdot}1$, $-4{\cdot}6$, and $0{\cdot}1$ W m^{-2}, respectively. Negative net radiation during the snow-covered period was also observed in eastern Siberia (Zhang *et al.*, 2004b). As noted by Zhang *et al.* (2004b), 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. Because of the atmospheric inversion and likely stable status, as shown in Figure 3, sensible heat flows toward the surface and serves as an efficient heat source for snow sublimation.

Snow depth

Snow cover, as a landscape element, is characterized by its seasonality. Spatial and temporal variability are high, even at small spatial scales, as shown by the repeated profiles sampled close to each of the study sites. Figure 4 illustrates the snow depth observed during the study period (November 2003 to March 2006) at sites FP, MSS, and MNS, the former two of which

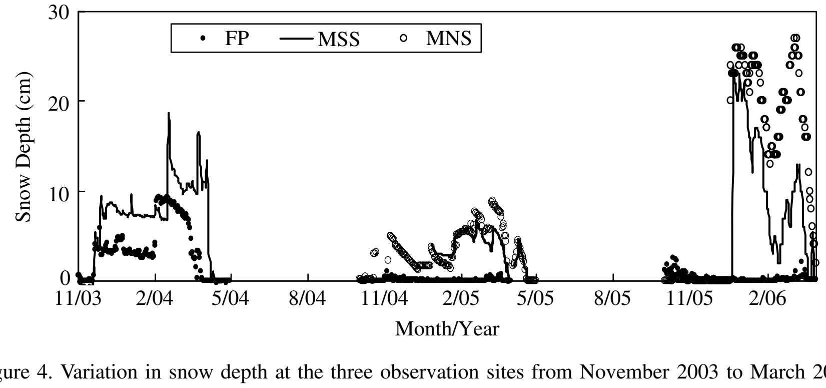

Figure 4. Variation in snow depth at the three observation sites from November 2003 to March 2006

were located on grassland, and the other a forest site. Comparing with other areas, snow cover at the study sites was thin, with significant spatial-temporal variability. During the winter of 2003–2004, the maximum snow depth was 9·3 cm at FP and 18·6 cm at MSS. However, the snow depth was smaller at FP during the winters of 2004–2005 and 2005–2006, with a maximum snow depth of 5·4 cm. With such thin snow cover, the snow mass may be consumed up due to wind drift or sublimation even if in mid-winter. From checking the albedo data, however, the days without snow cover have been taken off from the calculation period to meet the applicability of Equation (1).

Performance of flux measurements by eddy covariance and aerodynamic profile calculations

As latent heat fluxes were only measured using the eddy covariance system from 4 October 2005 to 21 March 2006 at MSS, calculations were also made using the aerodynamic profile formula presented in Equation (1) to present the full seasonal variation in latent heat flux. The values estimated by eddy covariance and the aerodynamic profile formulas are compared in Figure 5 for the calibration period; the regression slope is 1·0017 with an R^2 value of 0·7940. The latent heat flux calculated using the aerodynamic profile method was about 5% higher on average than the eddy covariance measurements. The agreement between the hourly values from the two series was good, giving credibility to the values from subsequent calculations.

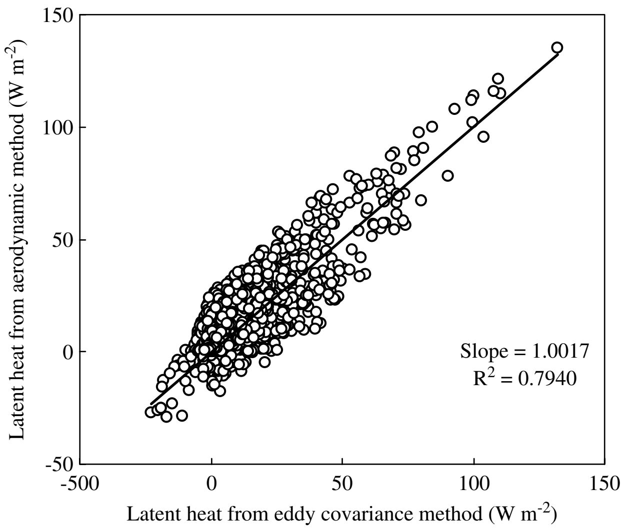

Figure 5. Relationship between the heat fluxes calculated using the aerodynamic profile method and the eddy covariance measurements from 4 October 2005 to 21 March 2006 at site MSS; each circle indicates an hourly mean

Snow sublimation

Snow sublimation was calculated from AWS data using Equations (1) and (2). Daily sublimation from the snow surface at the three study sites is shown in Figure 6 for the period November 2003 to March 2006. Results shown in Figure 6 ranged with 0 to 1·2 mm on a daily basis, which imply vapour flux was dominated by positive-sublimation but not condensation at the study site. Having checked

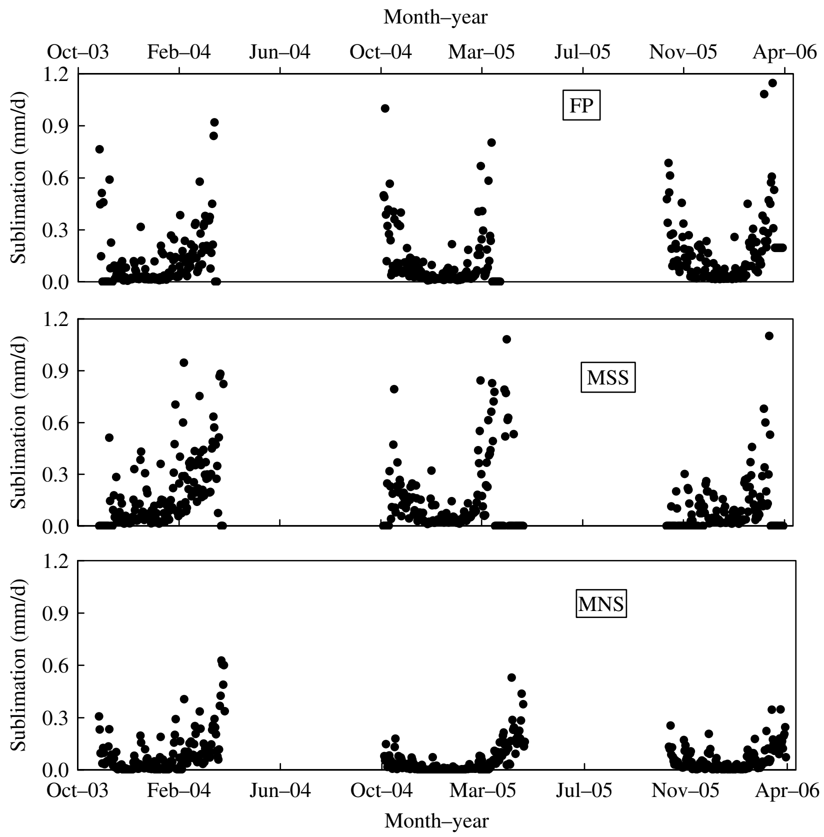

Figure 6. Variation in daily sublimation at the three sites from November 2003 to March 2006

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diurnal behaviour of observed latent heat flux, it was found to be negative at night-time in mid-winter, which indicated condensation processes. However, the negative amplitude was rather low on a daily basis.

The three time series show similar seasonal variations. Generally, the daily sublimation for the three sites varied synchronously. At all study sites, sublimation values were higher at the beginning and end of the snow-covered period. Nearly all peak values occurred in early November or after early February, but sublimation was rather low in the middle of winter. Similar seasonal variations have been described for Tibet and Siberia (Zhang *et al.*, 2003, 2004b). Sublimation from February to April accounted for about 63% of the total sublimation in the entire observation period at site FP and 75% at sites MSS and MNS. Minor variations at all sites occurred in mid-winter, when atmospheric inversions occurred (Figure 2). Zhang *et al.* (2004b) showed that stable or neutral atmospheric conditions suppress vapour flux from the snow surface.

For the entire study period, sublimation averaged 0·08 mm day^{-1} at sites MNS and FP and 0·16 mm day^{-1} at MSS. Table II presents total sublimation for early winter (November), mid-winter (December and January), and late winter (after February) and also difference in air temperature between 4 and 1 m and wind speed at 2 m among the three sites for different periods. The differences of sublimation between sites FP and MSS mainly tally with strength of atmosphere inversion, the coupling of a lower sublimation to a larger difference of air temperature between 4 and 1 m. The difference in sublimation between MSS and MNS, the former with values two-fold higher than those of the latter, coincided with significant increasing atmospheric stability near the ground which implied lower wind speed and larger difference in air temperature between 4 and 1 m.

Hood *et al.* (1999) demonstrated that sublimation is a somewhat 'episodic phenomenon', implying that sublimation losses generally proceed at low levels but are often punctuated by short 'sublimation events' during which sublimation may reach levels above 0·4 mm day^{-1} for 24 h. To examine the impact of such sublimation events, daily sublimation was classified as intense, medium, or weak in the ranges of $\geq$0·4, between 0·1 and 0·4, and <0·1 mm day^{-1}, respectively. Table III presents occurrences of events based on this classification, total sublimation in each category, and the proportion of the entire period such events occupied. To better illustrate the meteorological forces driving the sublimation process, the average values of air temperature, vapour deficit, and wind speed for every situation are also presented.

Table II. Comparison of snow sublimation (Es, in mm day^{-1}), difference of air temperature between 4 and 1 m ($AT_{4m} - AT_{1m}$, in ° C) and wind speed at 2 m (WS_{2m}, in m s^{-1}) among the three sites for different periods

	FP			MSS			MNS		
	Es (mm day^{-1})	$AT_{4m}-AT_{1m}$	WS_{2m}	Es (mm day^{-1})	$AT_{4m}-AT_{1m}$	WS_{2m}	Es (mm day^{-1})	$AT_{4m}-AT_{1m}$	WS_{2m}
Before 31 November	0·13	0·58	2·4	0·12	0·20	1·6	0·07	0·60	0·4
1 December–31 January	0·02	0·88	1·6	0·07	0·25	1·1	0·02	0·68	0·3
After 1 February	0·24	0·49	3·1	0·28	0·14	1·9	0·09	0·54	0·5
Mean	0·13	0·65	2·4	0·16	0·20	1·5	0·06	0·61	0·4

Table III. Statistical analysis of calculated daily snow sublimation (*Es*) at the three study sites

Sites	FP			MSS			MNS		
Classification	Intense	Medium	Weak	Intense	Medium	Weak	Intense	Medium	Weak
Range of daily Es (mm day^{-1})	Es $\geq$ 0·4	0·1 $\leq$ Es < 0·4	Es < 0·1	Es $\geq$ 0·4	0·1 $\leq$ Es < 0·4	Es < 0·1	Es $\geq$ 0·4	0·1 $\leq$ Es < 0·4	Es < 0·1
Frequency (days)	15	80	163	58	138	223	22	106	427
$\sum$Es (mm)	5·8	10·5	3·2	29·3	32·1	6·8	10·3	23·8	12·5
Proportion of $\sum$Es to that in entire period (%)	30	54	16	43	47	11	22	51	27
Mean air temperature (°C)	−9·8	−17·2	−25·0	−5·9	−12·7	−18·0	−1·8	−14·7	−18·0
Mean vapour deficit (hPa)	0·8	0·4	0·2	0·8	0·6	0·4	2·9	1·2	0·4
Mean wind speed (m s^{-1})	2·5	2·4	1·5	2·5	2·3	1·0	0·9	0·6	0·5

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During the three observed winters, sublimation totalled 19·5, 68·2, and 46·8 mm at sites FP, MSS, and MNS, respectively. Intense sublimation events covered 15, 58, and 22 days at sites FP, MSS, and MNS, i.e. 6, 14, and 4% of the entire period, respectively. However, the total sublimation occurring during these 'intense' periods at sites FP, MSS, and MNS represented 30, 43, and 22% of the overall total, respectively. Weak sublimation occurred at all sites 53 to 77% of the time, with a quantitative contribution of 11 to 27%. Medium sublimation occurred for 19 to 33% of the time and accounted for around 50% of the total.

Clearly, the difference in total sublimation among the three observation sites was dominated by the occurrences of intense sublimation, which were driven by meteorological conditions. Air temperature was higher than the overall averages, at 11·9, 10·0, and 14·9 °C at sites FP, MSS, and MNS, respectively. In contrast, the mean air temperature for weak sublimation days was lower than the overall averages, at 3·3, 2·1, and 1·3 °C at sites FP, MSS, and MNS, respectively.

The vapour flux between the atmosphere and ground surface, known as evaporation or sublimation, is determined by the vapour deficit and wind speed. The correlation between intense sublimation events and air temperature, presented in Table III, may be related to the plot of vapour deficit versus air temperature shown in Figure 7. The tendency of vapour deficit to increase with air temperature reveals an indirect correlation between enhanced snow sublimation and higher temperature.

The effect of wind speed on sublimation can be seen in the results from site FP, located in the sparse vegetation region. Here, the wind speed during intense sublimation events averaged 2·5 m s^{-1}, which was higher than the value of 2·3 m s^{-1} for the overall period. The low wind speed at site MNS, with a mean value of 0·4 m s^{-1}, was likely the primary reason why total sublimation was only half that at site MSS.

PROPORTION OF SUBLIMATION TO SNOWFALL

The earlier analysis indicates that snow sublimation varies with changes in climate variables, including air temperature, wind speed. Variations in sublimation also produce changes in the role that snow plays in the water cycle on a local scale. The relation of snow sublimation to the regional water cycle over longer time scales should thus be clarified. Most previous studies of snow sublimation have been carried out over a short time scale. The longest study period was reported by Suzuki *et al.* (2002), who simulated snow sublimation from 1989–1995.

Many models of snow behaviour, for numerous regions, have been published, all requiring meteorological data, including radiation, as input. However, the data that was possible to collect for this study region were not sufficient to run any of these models. Therefore, snow sublimation was estimated using the bulk formula

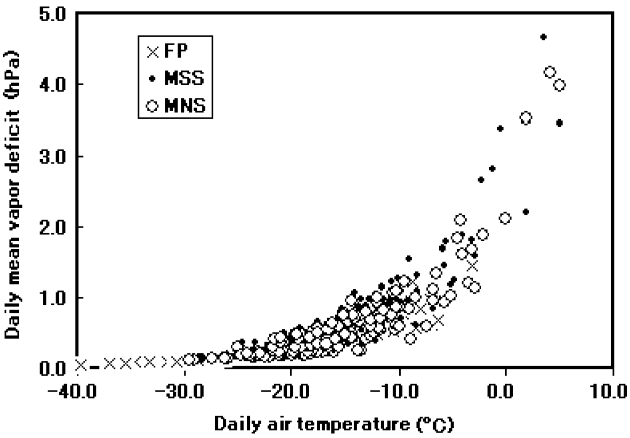

Figure 7. Variation in the daily mean vapour deficit versus the daily mean air temperature at the study sites

presented in Equation (3), and using the data from the meteorological stations, snow sublimation was estimated on a daily basis at both Ulan Bator for 1980–2004 and Terelj for 1986–2004. Figure 8 compares observed Es at site MSS and estimated Es at Terelj station. At both sites, grasslands extend for 1·2 km. The correlation coefficient between the observed and estimated values was 0·9662 with a significance level of 95% and standard error of 0·038 mm.

Seasonal variation in sublimation at the flat plain and mountainous sites

The mean seasonal variation in the estimated result is shown in the upper panel of Figure 9. The mean curves of wind speed and vapour deficit for the same period are also displayed in the lower panel to show the impact of atmospheric driving forces on the seasonality of sublimation. Similar seasonal patterns can be seen in both the flat plain and mountainous sites; the amount of sublimation was higher at both the beginning and the end of the snow-covered period. Such seasonality can be explained by the seasonality of wind speed and vapour deficit, the key parameters controlling sublimation.

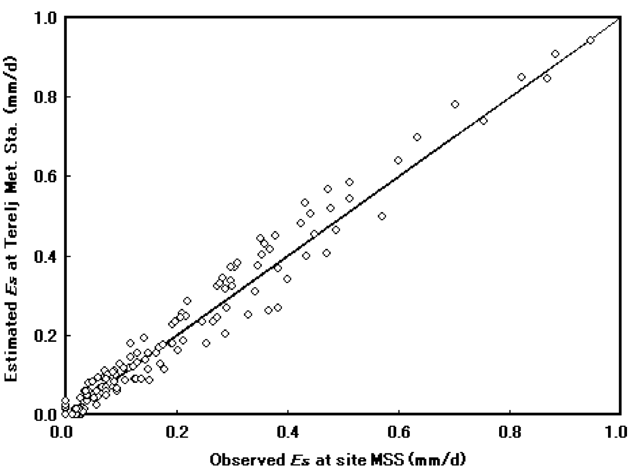

Figure 8. Comparison of observed snow sublimation (Es) at MSS versus estimated Es at Terelj station for the calibration period

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Generally, sublimation was higher in October and November, when wind speed and vapour deficit were higher, followed by declining sublimation through mid-winter. Near the end of the snow-covered period, sublimation reached its highest level when the vapour deficit and wind speed reached their peaks. The seasonal patterns shown in Figure 9 differ from those obtained for a seasonal snow pack in a moist region (Hood *et al.*, 1999; Leydecker and Melack, 1999), where the sublimation peak in mid-winter was dominated by stronger winds and a larger vapour deficit, followed by declining sublimation and increasing condensation near the end of snow melt. At the observation site, the predominant status of the atmospheric layer was stable or neutral, implying rather weak winds even in mid-winter and suggesting that the seasonality of sublimation from the snow surface relates closely to vapour deficit. As discussed earlier, variation in vapour deficit was significantly correlated with air temperature. If no significant changes in atmospheric vapour content occurred, the increasing saturated vapour pressure associated with higher air temperature would lead to an intensive increase in vapour deficit, resulting in higher sublimation.

It is worth to mention that the mean seasonal curves shown in Figure 9 do not correspond so well to the seasonal amplitude of wind speed and vapour deficient. The estimated sublimation at Terelj station were averagely higher than that at Ulan Bator station, while both wind speed and vapour deficient were lower than those at Ulan Bator station. This could be elucidated by the impact of atmospheric stability to snow sublimation as discussed earlier. Ulan Bator station located on the flat plain, and at the bottom of an inversion in the snow-covered period. Sublimation, therefore, was lower than that at the Terelj station, located in a mountain region.

Interannual changes in snow sublimation

Figure 10 shows the interannual changes in snowfall (Ps), Es, and the proportion of sublimation to snowfall (Es/Ps) at Ulan Bator (flat plain) for 1980–2004 and Terelj (mountainous) for 1986–2004. Snow-cover conditions varied synchronously but to different degrees at the two sites. The annual number of snow-cover days ranged from 20 to 148 at Ulan Bator and 76 to 177 at Terelj. The annual maximum snow depth ranged from 4 to 14 cm at Ulan Bator and 7 to 30 cm at Terelj. The snowfall averaged 67·6 mm, ranging from 24·4 to 139·4 mm, at Ulan Bator and 84·6 mm, ranging from 41·0 to 155·5 mm, at Terelj; the maximum snowfall was recorded in 1991 at both stations.

Lower variability was observed in interannual sublimation than in snowfall. Annual sublimation averaged 11·7 mm at Ulan Bator, ranging from 4·9 to 19·3 mm, and averaged 15·7 mm, ranging from 9·8 to 22·8 mm, at Terelj. Generally, no correlation was found between sublimation and snowfall. The ratio Es/Ps was variable, as determined by the variability in annual snowfall, and ranged from 8·5 to 52·1%. The value of Es/Ps averaged 20·3% at Ulan Bator and 29% at Terelj. The proportion of seasonal snow cover was reported to be 20% by Marks *et al.* (1992), 18% by Kattelmann and Elder (1991), and 15% by Hood *et al.* (1999).

DISCUSSION AND CONCLUSION

Variability of sublimation versus climate

Leydecker and Melack (1999) discussed the sensitivity of sublimation over a short time scale using snow surface roughness, instrument height, and wind speed, and demonstrated that wind speed was the critical variable for

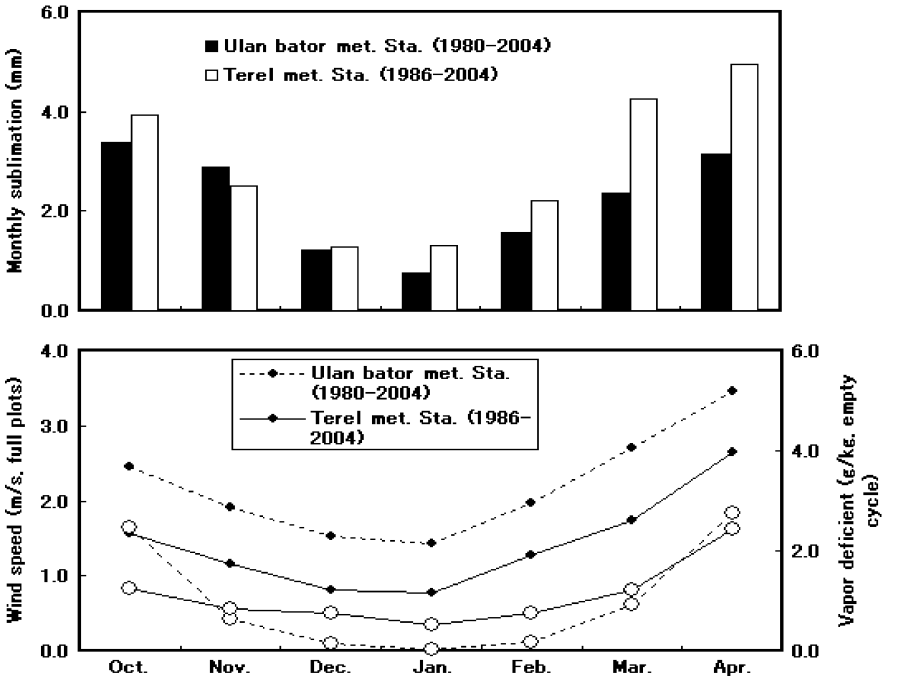

Figure 9. Monthly snow sublimation from 1980 to 2004 at Ulan Bator meteorological station and from 1986 to 2004 at Terelj meteorological station

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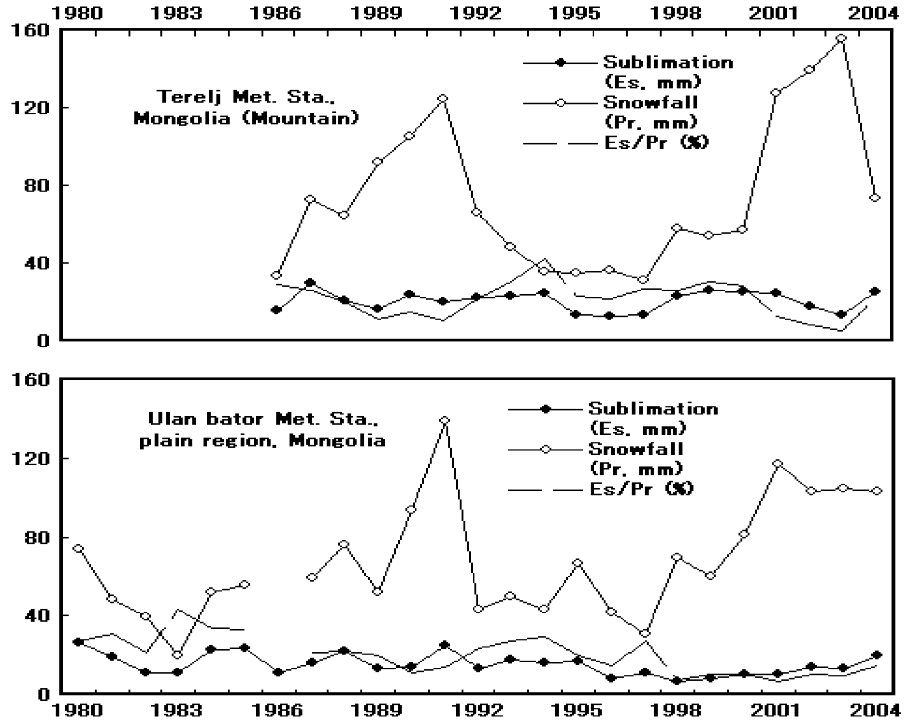

Figure 10. Interannual variation in snowfall (Ps), sublimation (Es), and the ratio of Es/Ps from 1980 to 2004 at Ulan Bator meteorological station and from 1986 to 2004 at Terelj meteorological station

determining sublimation, and that doubling wind speed can triple the sublimation. To examine the atmospheric driving force on sublimation on an interannual basis, we compared the significant correlation coefficients between annual sublimation (Es) and the mean values of air temperature (T), vapour deficit (Dv), and wind speed (U) during the snow-covered period at the Ulan Bator and Terelj meteorological stations (Table IV). The correlation coefficient between Es and U was similar to that between Es and T. The higher correlation coefficient between Es and Dv, i.e. 0·586 at Ulan Bator and 0·582 at Terelj, included the contribution from the correlation between T and Dv, where the coefficients were 0·334 and 0·356, respectively.

Table IV also suggests the complexity of the sublimation response to atmospheric forcing. Zhang *et al.* (2004b) demonstrated that when the wind speed is less than 2·0 m s^{-1}, the saturation deficiency is predominant in determining sublimation, and sublimation increases significantly when the wind speed is above 2·0 m s^{-1}. Combining the results of this work and those of Zhang *et al.* (2004b), the variation in vapour transfer coefficient versus wind speed can be deduced (Figure 11, left). This result shows that the vapour transfer coefficient for the snow surface is larger under weak wind and decreases sharply as wind speed increases. However, the vapour transfer coefficient is nearly constant when monthly mean wind speed is above 2·0 m s^{-1}. The impact of the coefficient on the sublimation rate is not so clear. Daily sublimation shows a tendency toward correlation with the vapour transfer coefficient (Figure 11, right).

Table IV. Significant correlation coefficients between annual sublimation (Es) and the mean values for air temperature (T), vapor deficit (Dv), and wind speed (U) during the snow-covered period from 1980 to 2004 at Ulan Bator and from 1986 to 2004 at the Terelj meteorological stations

	Es	T	U	Dv
Ulan Bator				
Es		0·364	0.375	0·586
T			−0.066	0·334
U				−0·043
Terelj				
Es		0·373	0.348	0·582
T			−0047	0·356
U				−0·065

It is somewhat difficult to make comparisons between sublimation magnitudes recorded in this study to those in previous studies because of the short periods of record seen in many of the previous studies. However, the values of 20·3–21·6% of total snowfall lost to sublimation compares favourably with the percentage of the snow cover estimated lost to sublimation of 25·6% in eastern Siberia by Suzuki *et al.* (2002), of 12–33% of annual snowfall (Pomeroy and Li, 1997).

The consequence of snow sublimation to hydrological cycle

The long-term estimations for both the flat plain and mountainous sites indicate that snow sublimation is important to the hydrology of the study region. If a large amount of snow accumulates, annual sublimation of less than 30 mm may not be significant; in the study region, however, the mean value of 20·3–21·6% for snow sublimation to snowfall (Es/Ps) implies the importance of

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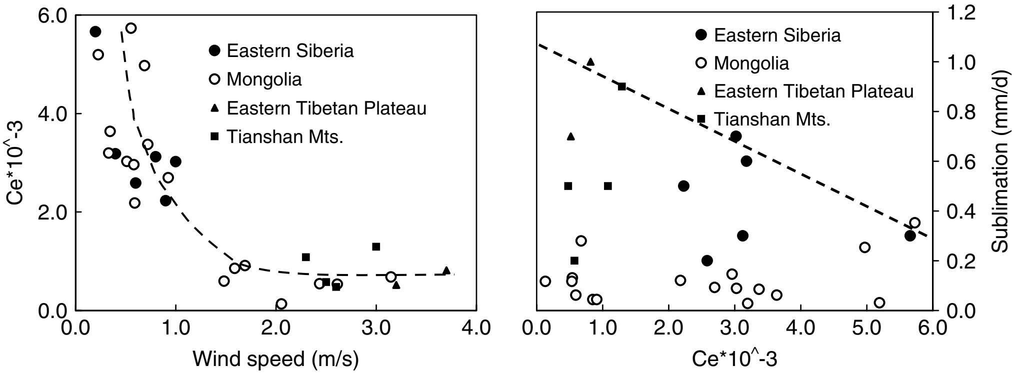

Figure 11. Vapor transfer coefficient versus wind speed (left) and daily sublimation versus the vapour transfer coefficient (right), partly quoted from Zhang *et al.* (2004b)

snow sublimation to the water cycle, which reveals that almost all snowfall in the study will evaporate back to the atmosphere finally, because snowfall is a proportion of 20% of the annual precipitation over Mongolia generally. Another important hydrological component in regions of thin snow cover is the effect of melt water. Zhang *et al.* (2004c) demonstrated that in the period from 10 to 15 days after snow melting, soil evaporation matched evapotranspiration exactly. Evaporation occurred from snowmelt penetrating the soil but was rarely related to precipitation. Table V shows the observation results of snow sublimation and soil evaporation of snowmelt water at site FP for two hydrological years, i.e. July 2002 to June 2003 and July 2003 to June 2004, and also the proportion to the annual amount. The contributions of sublimation and snowmelt water evaporation to annual evapotranspiration were 10 to 13% and 7 to 9%, respectively.

If loss of snow cover is counted by sublimation pulsing snow-melt water evaporation (snow sublimation/evaporation), which may be characterized by extreme situation winter water cycle by arid climate and fewer snowfall. Globally comparison may deduce that snow sublimation/evaporation plays a major role in snow ablation. Gray and Prowse (1993) reported that 40% of annual snowfall in Canada is lost by sublimation/evaporation. Beaty (1975) found 50–80% of the total springtime snow pack was estimated to sublimation and/or evaporation in White Mountains of Sierra Nevada, USA, owing to arid climatic conditions at the border to the Mojave Desert.

Table V. Annual sublimation, snowmelt water evaporation, and proportions to annual evapotranspiration for two hydrological years (from Zhang *et al.*, 2004c)

	July 2002–June 2003	July 2003–June 2004
Precipitation (Pr, mm)	122·0	150·5
Evapotranspiration (Et, mm)	123·4	178·2
Snow sublimation (Es, mm)	16·0	18·5
Soil evaporation of snow—melt water (Em, mm)	11·4	12·7
Es/Et (%)	13	10
Em/Et (%)	9	7

Vuille (1996) measured that sublimation could amount to 30–90% of the seasonal snow in South American Andes. By contrast, in the Alps under moist climatic condition, snow sublimation is less important on a seasonal basis or for water balance estimations (Lang, 1981; Tarboton *et al.*, 2000). According to Braun and Lang (1986) snow sublimation/evaporation is significant only over short periods in a moist environment.

Possible errors in calculating results

Compared to evapotranspiration, the range of variation in the amount of sublimation is small and therefore accuracy in calculation is especially important. Two main sources of error in the calculation by the aerodynamic profile method may be anticipated. The first of these is sublimation from blowing snow. Hood *et al.* (1999) estimated that sublimation from blowing snow may account for 2 to 20% of the annual snowfall, based on a mean wind speed of 7·3 m s^{-1} at their study site. For the study region in this article, the wind is much weaker, with a mean wind speed of less than 2·3 m s^{-1}. The error caused by blowing snow in this study is therefore probably lower. The second potential source of error is the accuracy of the aerodynamic profile, in which steady-state flow and homogeneous surface conditions are assumed. It is impossible to guarantee the steady-state requirement, but most authorities feel that deviating from it introduces little error even on the most complex surface (Moore and Owens, 1984). Brutsaert (1992) recommended using average data over a reasonable time interval as a good compromise between seasonally constant atmospheric conditions and adequate sampling. The data used in this study were 10-min average values for the entire period. Overall, the results from the meteorological station observations verify our estimations, and increase confidence in the validity of the results.

CONCLUSION

From continuous observations at three sites in Mongolia with thin snow cover, snow sublimation values were calculated using the aerodynamic profile method and the

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results were verified with observations taken from area meteorological stations during the calibration period. The findings can be summarized as follows.

The results indicate that sublimation variability was dominated by intense sublimation events characterized by daily sublimation exceeding 0·4 mm. The dominant meteorological elements affecting sublimation were wind speed and air temperature, the latter affecting sublimation indirectly through vapour deficit.

Both the observed and estimated results show that seasonal variation in sublimation was characterized by low values in mid-winter, relating to wind speed and air temperature through vapour deficit. Evidently, higher wind speed appears at the beginning and end of the snow-covered period.

Annual sublimation averaged 11·7 mm at the flat-plain meteorological station, or 20·3% of the annual snowfall, and 15·7 mm at the site in the mountains, or 21·6% of snowfall, which reveal that almost all snowfall in the study is evaporated back in to the atmosphere finally.

The sum of snow sublimation and snowmelt evaporation represented 17 to 20% of annual evapotranspiration in a couple of observation years.

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