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Observation and estimation of evaporation from the ground surface of the cryosphere in eastern Asia

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Abstract:

The characteristics of evaporation from the ground surface of Asian cryosphere sites are presented, as estimated by the lysimeter method, a profile method, and a heat budget method. The observation sites were located on the eastern Tibetan Plateau, in the Qilian and Tianshan Mountains of China, and in eastern Siberia. The lysimeter method has been demonstrated to be a reliable observation technique for estimating daily evaporation from the land surface, given suitable experiment design and operation. Daily mean evaporation varied within the range of 0·3 to 3·5 mm on the permafrost surface, and regional differences in evaporation were strongly related to surface soil moisture. Locally, topography, by way of its influence on surface soil moisture, was found to control evaporation systematically. Seasonality of ground evaporation in permafrost regions is dominated by thaw–freeze cycles at the surface; evaporation from the melting permafrost surface is up to four to seven times greater than that from frozen ground. In forested terrain, the interception of precipitation can reduce daily evaporation by 60 to 70%. Sublimation from the snow surface was observed at some sites in the range of 0·2 to 1·0 mm daily; atmospheric conditions, such as wind speed and saturation deficit, were dominant factors in determining snow sublimation. Copyright © 2003 John Wiley & Sons, Ltd.

KEY WORDS Asia; cryosphere; evaporation; lysimeter; permafrost

INTRODUCTION

The cryosphere, which may serve as a sensitive indicator of global climate change, consists of ice in all its forms: permafrost, snow cover, and glaciers. In fact, cryospheric observations alone provide nearly unequivocal evidence of 20th century global warming. There is an increasing awareness that the cryosphere plays a significant role in climate change, and is also a valuable water resource in its own right (Kulkla and Kukla, 1974; Williams, 1975; Hahn and Shukla, 1976; Barnett *et al*., 1988; Lawford, 1993). The cryosphere, a heat sink for the atmosphere, is an important element in the control of the water cycle and the heat budget in the atmosphere–hydrosphere system. To understand better the water/energy cycle involving the atmosphere and cryosphere, evaporation from the ground surface of the cryosphere needs to be accurately measured, because it determines the vapour and latent heat flux between the atmosphere and cryosphere; the latent heat of evaporation significantly influences the heat available for ice and snow melt.

One characteristic of the cryosphere is seasonal variation in hydrologic processes. In summer, snow and ice melt at the surface and the surface layer of permafrost thaws to form an active layer. Evaporation on the surface can then be quite significant. In winter, evaporative flux takes the form of sublimation from the frozen surface. These seasonal variations make it clear that different formulas must be used for periods when the ground surface is frozen and melting to estimate total annual evaporation.

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Evaporation also plays a significant role in other types of cryospheric surface, characterized by the freezing and melting of ice. On the surface of a continental-type glacier, evaporation can account for as much as about 20% of the total loss due to ablation (Zhang *et al.*, 1998). The annual evaporation from permafrost surfaces of the eastern Tibetan Plateau has been estimated at 310 mm, corresponding to a heat transfer of 775 MJ m^{-2} (Zhang *et al.*, 1994). Since 1986, evaporation from the ground surface of permafrost, glacier, and snow cover has been estimated based on lysimeter observation, a profile method, and the heat budget method. The results presented and analysed in this paper provide a clearer understanding of the water cycle in this region than has previously been achieved.

STUDY SITES AND METHOD

Study sites

Cryosphere areas are widely distributed in Asia; the most northerly region is in Siberia, south of the Arctic Ocean. This region is characterized by a long period of snow cover, continuous permafrost, and little precipitation. Another principal cryospheric region is the Tibetan Plateau, which is characterized by its high elevation, on average 4000 m. The climate of the Tibetan Plateau is strongly affected by the Asian monsoon; more than 80% of its precipitation falls during the summer. Between the two main cryosphere regions described above, there are many others, located in arid and semi-arid areas; this type of cryosphere is characterized by alpine glaciers and surrounding alpine permafrost. In lower alpine regions, no permafrost develops, but seasonal thaw–frost cycles are widely evident. This paper presents the results of evaporation observations at four sites that represent each type of cryosphere region described above: the eastern Tibetan Plateau, the Qilian Mountains, the Tianshan Mountains, and Tiksi, in tundra area of eastern Siberia (Figure 1).

Table I shows selected hydro-climatic and ground surface characteristics for each region studied. The observation site on the eastern Tibetan Plateau is located at 33°02′N, 92°02′E, near the Dongkemadi River, in the Tanggula Mountains. The area of the watershed is about 21·5 km^2, of which 16·4 km^2 is covered by a duplex valley glacier. An automatic weather station was installed in 1989, and from data for 1989–92 the annual air temperature was found to be $-5{\cdot}8\,^{\circ}C$, the mean wind speed 3·4 m s^{-1}, and the annual precipitation about 560 mm, of which more than 68% occurs in summer. An observation site for evaporation was set up on a smooth ground surface of permafrost. The soil type at the surface is silt containing organic matter, with a porosity of 64%. The ground surface is sparsely covered by a single type of grass, which grows no higher than about 5 cm, with a normalized differential vegetation index (NDVI) ranges from 0·31 to 0·44, as estimated from Landsat TM data (Yabuki *et al.*, 1998). The ground starts to thaw in about mid April and thaws to a

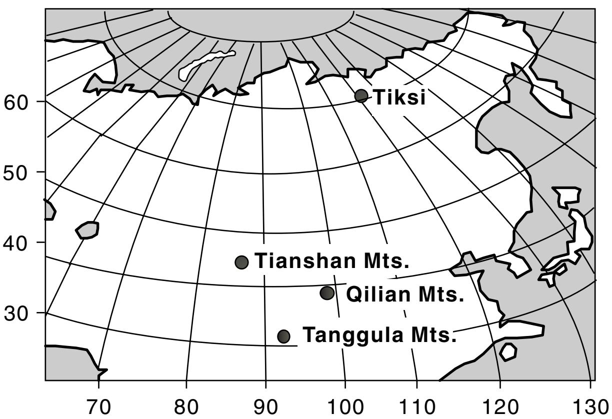

Figure 1. Location of observation sites

Table I. Hydro-climatic condition of the study regions

	Eastern Tibetan Plateau	Qilian Mts	Tianshan Mts	Tiksi, eastern Siberia
Location	33°02′N, 92°02′E	39°14′N, 99°45′E	43°06′N, 86°49′E	71°35′N, 128°46′E
Altitude (m a.s.l.)	5070	2100	3549	110
Annual mean air temperature (°C)	−6·8	2·1	−5·3	−13·5
Annual mean relative humidity (%)	71	63	58	75
Annual mean wind speed (m s^{-1})	3·4	1·0	2·5	5·0
Annual precipitation (mm)	560	390	431	345
Ground surface	Permafrost, snow cover and glacier	Vegetated soil (seasonal frost)	Permafrost, snow cover and glacier	Permafrost, snow cover
Surface conduct evaporation observation	Permafrost, snow cover	Soil, soil in forest	Permafrost, snow cover	Permafrost
Soil type	Silt with organic matter	Organic soil	Silt with organic matter	Silt with organic matter
Vegetation	Sparse grass	Grass in forest and grass	Dense grass	Tundra
Maximum soil thaw depth (cm)	125	250 (frost table)	220	80
Observation method for evaporation	Lysimeter, profile	Lysimeter	Lysimeter, profile, heat budget	Lysimeter

maximum depth of about 125 cm. More than 80% of precipitation occurs between May and September, some of it as snow; thus, temporary snow cover has been observed in May and June. Evaporation of the permafrost and sublimation over snow cover were observed in the summers of 1989, 1992, and 1993.

The observation site in the Qilian Mountains is located on the margin of the Tibetan Plateau (39°14′N, 99°45′E). In 1997, the annual air temperature was 2·1 °C, and annual precipitation, annual mean wind speed, and mean relative humidity were 390 mm, 1·0 m s^{-1}, and 63% respectively. No permafrost develops in the ground, but the surface layer is affected by a controlling seasonal thaw–freeze cycle. The surface starts to freeze from late October; frozen soil may extend to maximum depth of about 250 cm in January. The surface layer is organic soil, derived from very dense grass, or pine-dominated forest. Evaporation was observed in the summer of 1997, both in the grassland and in the forest.

The watershed studied in the Tianshan Mountains is located in the source region of the Urumuqi River. The catchment area of the watershed is 28·9 km^2, including seven glaciers, which cover 5·74 km^2. The altitude of the basin ranges from 3403 to 4479 m a.s.l., with a surface comprised of glaciers, bare rocks, and moraines above 3600 m and alpine tundra below. The soil of the alpine tundra is dominated by silt with organic matter. Permafrost develops in the vicinity of the glaciers, with the maximum depth of the thaw table reaching 220 cm. According to observations conducted at the Daxigou Meteorological station (3539 m a.s.l., 30 m distant from study site) from 1959 to 1996, the mean annual air temperature was about −5·1 °C. The mean annual precipitation was 436·4 mm. Most of the precipitation was concentrated in the summer, from June to August, accounting for 66% of the annual total. Precipitation in winter falls as snow, with a total depth of less than 30 cm.

The observation site in eastern Siberia is located 7 km southeast of Tiksi (71°35N, 128°46E), in permafrost terrain. The thawing table is less than 80 cm deep, even in mid-summer. The ground surface is covered by typical tundra plants, including mosses, lichens, and sphagnum. The vegetation cover develops in a spatially heterogeneous fashion, with bare bedrock normally exposed at the tops of hills. The leaf area index has been reported to vary between 3·4 and 3·6. The soil in the area is thin and relatively undeveloped, consisting of poorly decomposed organic material overlying gravel or regolith. A multi-layered system has developed, consisting of 0 to 0·2 m of accumulated organic material on 0·05 to 0·30 m of partially decomposed organic matter, which in turn lies over mineral silt above the permafrost (Watanabe *et al.*, 2000). Bulk density, hydraulic conductivity, and thermal conductivity range from 0·25 to 1·0 g cm^{-3}, 4 to 140 × 10^{-4} cm s^{-1} and 0·70 to 1·21 Wm K^{-1} respectively. Snow covers the whole region from November to May, and small areas of snow pack persist to mid July. In 1999, the annual mean air temperature, relative humidity, and wind speed were −13·0 °C, 75%, and 5·0 m s^{-1} respectively. Annual precipitation was 345 mm. Precipitation in summer (from June to August) averaged about 170 mm, but showed high year-to-year variability, ranging from 110 mm in 1987 to 290 mm in 1996.

Methods

Lysimeter method. The lysimeters used in this study were installed by setting a cylindrical container into the soil, at the level of the natural surface, enclosing a block of natural soil of the same size and shape as the drum. The weight changes were measured every day, and thus the evaporation could be evaluated by using the following equation (Allen, 1990):

$$E_l = \Delta W/S + \mathrm{Pr} \tag{1}$$

where E_l is evaporation measured by the lysimeter method, ΔW is the weight difference measured by the lysimeter, S is the surface area of the soil within it, and Pr is precipitation. The soil in the drum was changed every 3 to 5 days, depending on climatic conditions. It should be emphasized that the soil enclosed in the drum was carefully selected to ensure that its layer-by-layer constitution was as close as possible to that of the nearby, undisturbed soil, and placement relative to the natural surface was maintained. Daily potential evaporation E_t can be measured by the lysimeter method by keeping the soil in the drum saturated.

Profile method. The vapour flux between the ground surface and the atmosphere E_p is computed by the following formula:

$$E_p = \rho C_E U(q_a - q) \tag{2}$$

where E_p is evaporation calculated by profile method, ρ is the air density, U is the wind speed, q_a and q are specific humidity at height z and at the ground surface respectively, and C_E is the bulk vapour transfer coefficient, which can be calculated as (Brutsaert, 1984):

$$C_E = k^2 \left[\ln\left(\frac{z}{z_0}\right) - \Psi_w\right]^{-1} \left[\ln\left(\frac{z}{z_0}\right) - \Psi_m\right]^{-1} \tag{3}$$

where k is the Karmann constant (0·4), z and z_0 are height and roughness, and Ψ_w and Ψ_m are stratification functions, which depend on atmospheric stability.

Heat budget method. Considering the heat budget at the ground surface, the latent heat flux can be calculated by (Brutsaert, 1984):

$$\lambda E_h = \frac{Q_N - Q_G}{1 + \beta} \tag{4}$$

where E_h is evaporation calculated by the heat budget method, λ is the specific latent heat of evaporation, Q_N is net radiation, Q_G is heat flux in the soil, and β is the Bowen ratio.

Inter-comparison of the results by the different methods. The lysimeters used in this part of the study were simplified. Versions of those used by Howell *et al.* (1991), Prueger *et al.* (1997), Young *et al.* (1996) etc. were 150–200 cm in diameter and 500 cm in length. The size of lysimeter used in this work was reduced in size to 33 cm in length and 30·5 cm in diameter. Such small lysimeters have long been used; their reliability was first evaluated by Boast and Robertson (1982). Subsequently, the refinements and modifications have been discussed by Shawcroft and Gardner (1983), Martin *et al.* (1994), Lascano and Van Bavel (1986), Allen (1990), and Daamen *et al.* (1993). It has been demonstrated that the results from such lysimeters relate closely to those obtained by the water budget method (Allen, 1990), by the usual, larger lysimeters, and by the infrared temperature method (Matthias *et al.*, 1986). Since the early 1980s, many studies have used the lysimeter method and evaluated its accuracy, both for bare soil and for canopy-covered soils. Shawcroft and Gardner (1983) considered the factors that affect the accuracy of lysimeters. They proposed that it is important to establish a correlation between evaporation and the moisture content of the surface soil layer, and then use that empirical relationship to estimate evaporation. Allen (1990) asserted that lysimeters should be used only during periods of no precipitation. Daamen *et al.* (1993) experimented with lysimeters of different diameters, and concluded that the inside diameter did not affect the accuracy of evaporation measurements.

The experimental results obtained at Tianshan, China, are presented in Table II. In rainy periods, the mean value of E_l was appreciably different from E_p and E_h. E_p and E_h averages were very similar; they differed by only 0·1 mm, and the relation between them seems to be constant, as is indicated by the fact that both were larger than E_m by a similar ratio. E_l was 11% smaller than E_m in rainy periods. During periods without precipitation, however, E_l differed from E_m by only 3%. At these times, the relation between E_p and E_h was no longer stable; E_h was greater than E_m, but E_p was smaller than E_m.

To review the accuracy of lysimeters, the observed and calculated results obtained in the summer of 1986 by the lysimeter (E_l), profile (E_p), and heat budget (E_h) methods at the Tianshan, China, site are compared in Figure 2. In an 80 day period (from June 13 to August 31), the average differences between the lysimeter and profile results and the lysimeter and heat budget results were 0·12 mm day^{-1} and $-0·13$ mm day^{-1} respectively. The error in the lysimeter method compared with the profile and heat budget methods was thus 4% and 5% respectively; overall, E_l was larger than E_p but less than E_h.

Table II. The statistical results of evaporation observation using lysimeter (E_l), profile (E_p) and heat budget method (E_h) at the site in Tianshan, China

		Lysimeter E_l	Profile method E_p	Heat budget method E_h	Average E_m
Rainy days	Mean value (mm day^{-1})	2·6	3·0	2·9	2·8
	$(E_m - E)/E_m$ (%)	11	−6	−4	
Days without rain	Mean value (mm day^{-1})	3·3	3·1	3·4	3·3
	$(E_m - E)/E_m$ (%)	3	7	−6	

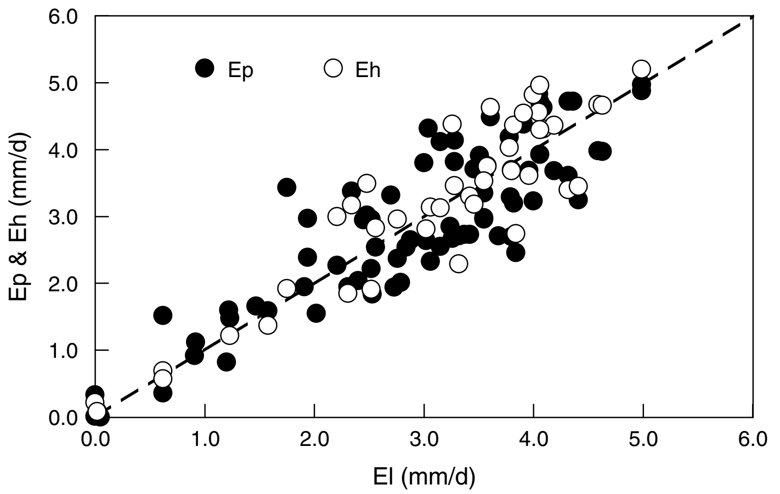

Figure 2. Comparison of daily evaporation observed by the profile method (E_p) or heat budget method (E_h) versus lysimeter results at the Tianshan site

Overall, the lysimeter results are reasonable and provide a practical basis for estimating point evaporation under less than ideal field conditions, as shown in the following section.

RESULTS AND ANALYSIS

Observed daily evaporation on permafrost surface

Table III shows the observed daily evaporation from the ground surface at the study sites on the eastern Tibetan Plateau, in the Tianshan Mountains, at Tiksi in eastern Siberia, and in the Qilian Mountains. The latter site is on seasonally frozen ground, whereas the others are on permafrost. For inter-comparison, E_l was taken as the definitive estimate of daily evaporation, except on rainy days, when E_p or E_h were used. Mean values are shown for some factors that may affect evaporation, as these factors were not observed at the same time.

In the eastern Tibetan Plateau region, daily mean evaporation ranged from 0·8 to 3·5 mm day^{-1}. In May, as the thawed depth increased, evaporation decreased. The daily mean values in May were 1·7 mm day^{-1} in both 1989 and 1992, and 0·8 mm day^{-1} in 1993. It is notable that daily mean evaporation in June was very variable from year to year: 3·5 mm day^{-1} in 1989 and 2·1 mm day^{-1} in 1993. The former value accorded well with the average potential evaporation value of 4·9 mm day^{-1} in response to the average wind speed of 3·9 m s^{-1}, even though both the air temperature and saturation deficit were lower.

Table III. Observation results of daily evaporation E, potential evaporation E_t, surface soil moisture Θ, air temperature T_a, wind speed U, and saturation deficiency De on soil surface of selected sites

Observation site	Vegetation	Observation period (y/m)	E (mm day^{-1})	E_t (mm day^{-1})	E/E_t	Θ (%)	T_a (°C)	U (m s^{-1})	De (hPa)
Eastern Tibetan Plateau	Sparse grass	1989/5	1·7				−1·4	3·1	1·7
		1989/6	3·5	4·9	0·71		−0·9	3·9	1·2
	Sparse grass	1992/5	1·7	4·8	0·35				
		1992/10	1·4	3·9	0·36				
	Sparse grass	1993/5	0·8	2·5	0·32	43	0·1	2·3	3·1
		1993/6	2·1	3·3	0·64	40	2·3	1·4	2·1
		1993/7	2·7	3·4	0·79	46	5·1	1·2	2·1
		1993/8	2·5	3·6	0·69	46	4·9	1·2	1·5
		1993/9	1·8	3·0	0·60	47	2·0	1·3	1·4
Qilian Mts	Dense grass	1997/8	2·2	4·6	0·48		12·9	0·9	3·0
		1997/9	1·2	4·6	0·26		8·4	1·1	3·4
		1997/10	0·4	3·3	0·12		2·7	1·3	2·7
	Grass in forest	1997/8	0·7						
		1997/9	0·5						
		1997/10	0·3						
Tianshan Mts	Dense grass	1986/6	2·7	3·3	0·82	47	2·9	2·4	2·3
		1986/7	3·2	3·4	0·94	54	6·8	2·3	3·4
		1986/8	2·6	3·7	0·70	47	5·4	2·4	3·6
Tiksi, eastern Siberia	Tundra	1999/7	2·1	4·3	0·49	35	10·0	3·6	3·3

Potential evaporation (E_t) in summer (June to August) was surprisingly similar at the Tianshan and eastern Tibetan Plateau sites, but the E/E_t ratio showed a clear difference between the two regions; the average daily value was 0·64 to 0·79 on the eastern Tibetan Plateau, but 0·70 to 0·94 in the Tianshan Mountains, resulting from the lower daily evaporation at the eastern Tibetan Plateau site. This difference was probably due to the effect of surface soil moisture on evaporation. Such an effect can be seen more clearly in results obtained at the Tiksi site, in Eastern Siberia. The average E_t value was 4·3 mm day^{-1}, as calculated from air temperature, wind speed, and saturation deficit values of 10·0 °C, 3·6 m s^{-1}, and 3·3 hPa respectively, but the E/E_t ratio was 0·49, in response to the low surface soil moisture (35%).

Daily evaporation from seasonally frozen ground was normally lower than that on the permafrost surface, perhaps because of lower wind speeds. The mean wind speed in August was 0·9 m s^{-1} at the Qilian Mountains site, but 2·4 m s^{-1} in the Tianshan Mountains and 1·2 m s^{-1} on the eastern Tibetan Plateau. In addition, the presence of forest restricts the evaporation process owing to interception of precipitation, as is shown by the results from the sites in the Qilian Mountains (see Table III), where the difference in the monthly average daily evaporation between open grassland and the ground surface in a forest is evident. The former is three times higher in August and two times higher in September. Evaporation is similar in October because few precipitation events occur, which is controlled by the seasonality of the monsoon climate.

Evaporation is known to be affected by available energy, near-surface air turbulence, soil water, and vegetation (Xu and Singh, 1998). These effects can be partially separated by expressing evaporation as a fraction of its potential value; the ratio of actual evaporation to potential evaporation E/E_t, or evaporation efficiency, is a measure of the extent to which the land surface exerts control over the evaporation process. To investigate the dependence of evaporation efficiency on the availability of stored water, the variation of

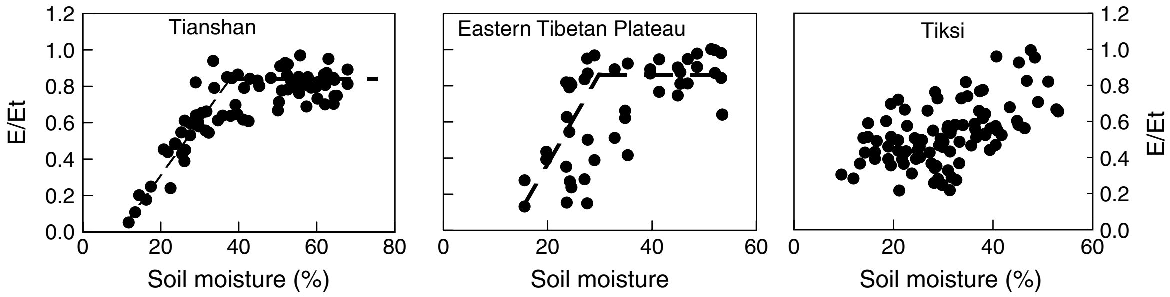

Figure 3. The variation of evaporation efficiency versus surface soil moisture at sites in the Tianshan Mountains, on the eastern Tibetan Plateau, and at Tiksi, eastern Siberia

E/E_t with surface soil moisture is shown in Figure 3. Kondo *et al.* (1990) noted a similar critical value of E/E_t, and reported that evaporation efficiency for a loam soil became 1·0 when soil moisture reached 28%.

In Figure 3, the different shapes of the relation between E/E_t and the soil moisture content at different sites are clear. At the Tianshan site, E/E_t increases linearly with surface moisture when water content is less than 40%, but shows no tendency to increase with moisture content beyond that level. This result stresses the importance of the critical value of approximately 40% soil moisture. When soil moisture is less than 40%, the evaporation process is restrained by the deficiency of available water. At the eastern Tibetan Plateau site, the level of surface soil moisture seldom rose above about 30%, and E/E_t decreases as soil moisture drops below that value; but at the Tiksi site, the plot of E/E_t shows little definite relation, beyond a tendency to increase with surface soil moisture, in the observed soil moisture range (less than 60%). This phenomenon can be explained by the characteristic structure of the surface soil layer at the study site in Tiksi. In the tundra region of Tiksi, the surface soil layer consists dominantly of organic material, which came from the moss. Higher percentages of organic material lead to greater porosity; the bulk density of the soil in the 0–5 cm layer depth has been reported to be 0·21 g cm^{-3} at the study site (Watanabe *et al.*, 2000). This implies that the soil seldom becomes saturated and this restrains evaporation.

Observed daily sublimation on snow surface

Observed daily sublimation values on the snow surface, at both the Tianshan Mountains and eastern Tibetan Plateau sites, are shown in Table IV. It must be emphasized that all the results shown were obtained on days without snowfall; inaccurate snowfall measurement meant that snow sublimation in bad weather was estimated with little success.

The results from the Tianshan region show that daily mean sublimation from the snow surface increased from February to April, but then decreased in May. Sublimation values were higher at the eastern Tibetan

Table IV. Observation results of daily sublimation on snow surface E on days without snowfall, and the air temperature T_a, wind speed U, and saturation deficiency De at selected sites

Observation site	Observation period (y/m)	E^a (mm day^{-1})	T_a (°C)	U (m s^{-1})	De (hPa)
Eastern Tibetan Plateau	1989/5	0·7	−5·7	3·2	1·4
	1989/6	1·0	−1·8	3·7	1·1
Tianshan Mts	1991/2	0·2	−14·6	2·5	0·9
	1991/3	0·5	−10·5	2·3	1·3
	1991/4	0·9	−6·6	3·0	1·5
	1991/5	0·5	0·1	2·6	2·6

[a] For no snowfall days only; the sublimation for snowfall days was observed to be zero.

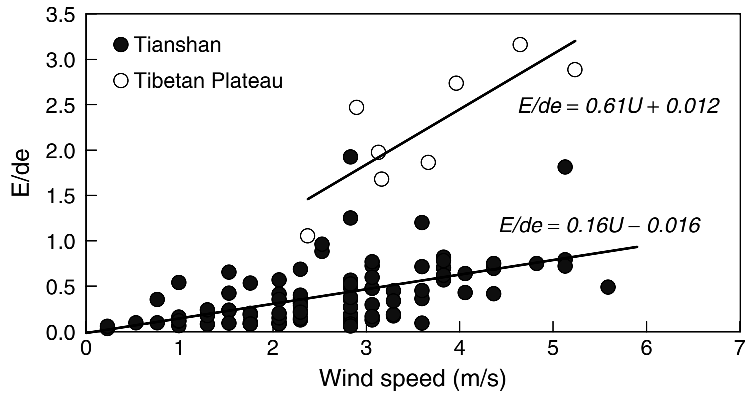

Figure 4. The ratio of daily sublimation to saturation deficiency *E*/De versus daily mean wind speed in the Tianshan region and on the eastern Tibetan Plateau

Plateau site than at Tianshan. Gaps in the overlap of observed time series mean that the results at the two sites are difficult to compare directly, although it can be seen that sublimation increased with wind speed. Kojima (1979) demonstrated that snow sublimation is determined by wind speed and atmospheric saturation deficit using the following equation:

$$E' = (aU + b)(e_{sa} - e_a) \tag{5}$$

where E' (mm day^{-1}) is the sublimation, U (m s^{-1}) is wind speed, e_{sa} (hpa) and e_a (hpa) are the saturated vapour pressure and the vapour pressure at a given height above the ground surface respectively, and a and b are constants. The variation of the ratio of daily sublimation to daily mean saturation deficit ($e_{sa} - e_a$) versus wind speed, obtained on no-snowfall days at the Tianshan Mountain and eastern Tibetan Plateau sites, is shown in Figure 4. The ratio on the eastern Tibetan Plateau, where greater sublimation was observed, is clearly larger than in the Tianshan Mountains, and is more sensitive to wind speed fluctuation.

Effects of topography on evaporation patterns

Topography affects evaporation indirectly by affecting meteorological variables or soil moisture, which in turn affect evaporation. In the summer of 1993, evaporation was observed at the top, middle, and bottom points on the profile of a 100 m hill at the eastern Tibetan Plateau site. The results showed that the spatial distribution across such a low hill profile was affected by variation in soil moisture. When precipitation produced similar soil moisture content across the surface soil of the hill, the stronger wind and higher vapour transmissivity at the top of the hill generally resulted in more evaporation. During rain-free periods, however, less evaporation occurred at the top, because of lower water retention (due to downslope flow) and consequent lower water content in the soil. The difference in evaporation among the three points was small; the mean difference was within 0·3 mm in June, when the ground surface tended to have a lower water content (on average, 38%). From July to September, the difference in evaporation between these points increased, reaching 1·0 mm or more, with increasing water content in the surface soil.

A network of 19 observation sites for soil moisture and evaporation was established in a small watershed near Tiksi, in eastern Siberia, in July 1999, with the aim of investigating the influence of topography on soil evaporation. The observation sites were set up 50 m apart, along a line that transected a river. Evaporation was observed by the lysimeter method, and surface soil water content was measured by time domain reflectometry (TDR).

The results did not show a direct relation for evaporation with any topographic index. However, a good logarithmic regression fit was deduced between the ratio of daily evaporation to potential evaporation (E/E_t) and the elevation difference (DZ) between the observation point and the nearby stream. Variation of the

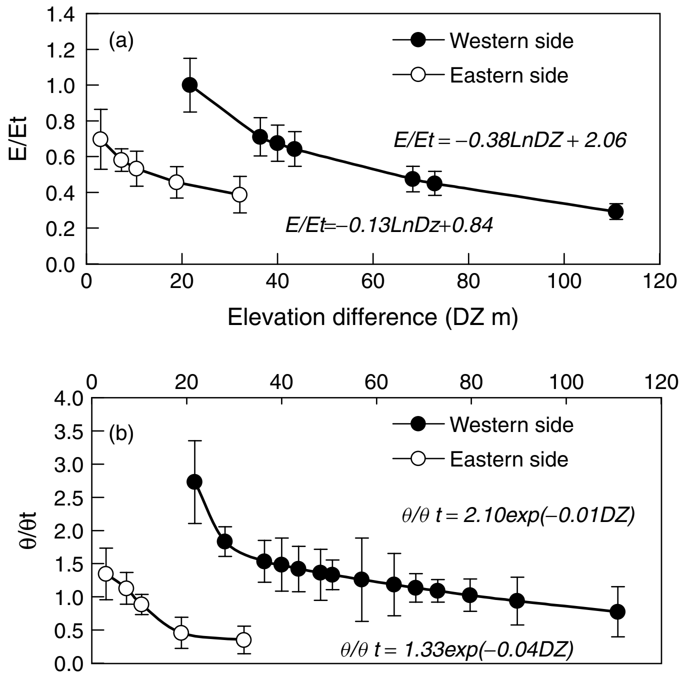

Figure 5. Variation of ratio of daily evaporation to potential evaporation E/E_t versus elevation difference (a) and ratio of soil moisture to saturate moisture θ/θ_t versus elevation difference at the Tiksi network

ratio (E/E_t) versus elevation difference, as predicted from the regression equation, is shown in Figure 5a, along each side of the river. It is clear that evaporation was more sensitive to DZ at the bottom part of the slope. The spatial distribution functions on either side of the river had different coefficient values, which may have been caused by the presence of a small area of snow pack on the western side of the river. The difference in evaporation between the two sides of the river reduced as DZ increased. When DZ equalled 20 m, the mean daily evaporation values differed by 1·8 mm day^{-1}; as DZ increased to 30 m, the difference became 1·2 mm day^{-1}. Atmospheric conditions can be expected to vary little over such a small spatial range (within 1 km). The variation in evaporation, therefore, is likely related to the spatial distribution of surface soil moisture, as shown in Figure 5b. The ratio of the surface soil moisture at points on the network to their saturated moisture θ/θ_t decreases as the elevation difference increases, according to the regression equation shown in Figure 5b.

Seasonality of evaporation on permafrost surface

On the cryospheric ground surface, the thaw–freeze cycle and its influence on moisture availability in shallow soil is the predominant component affecting hydrologic processes on land. For more than half the year, the ground surface is covered by snow and the ground is frozen to a depth of several hundred metres. Vapour is exchanged at low levels between the ground and atmosphere by sublimation. Beginning in May, the ground starts to melt from the surface downward as the air temperature rises; as soil moisture becomes more abundant near the ground surface, evaporation occurs until September, when the ground starts to freeze again, from the surface.

To investigate the seasonality of evaporation on the ground surface of permafrost regions, daily evaporation was evaluated at the study sites in the Tianshan Mountains and on the eastern Tibetan Plateau. Owing to restricted availability of data, calculations were performed for the period from 1 October 1985 to 30 September 1986 for the site in the Tianshan Mountains, and for the period from 1 October 1992 to 30 September 1993

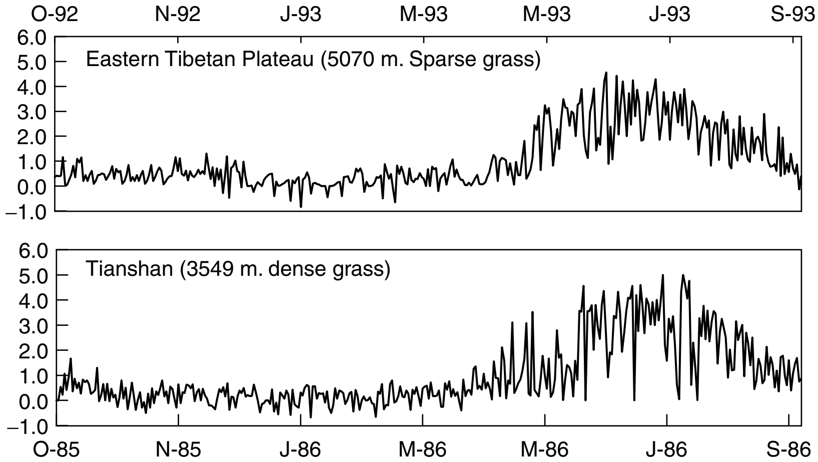

Figure 6. Calculated result of daily evaporation at the study sites at Tianshan and on the eastern Tibetan Plateau

for the site on the eastern Tibetan Plateau. This computation employed the profile method (as shown in Equations (2) and (3), which has been widely used for computing heat budgets on the Tibetan Plateau (Koike, 2000). Unfortunately, tower observations were carried out only from June to August 1986, for heat budget studies at the site in Tianshan, China, one of the study regions described here. To present a seasonal curve for the full year, Equation (5) was used to calculate the sublimation from the frozen ground surface, using meteorological data as necessary. The regression equation results are shown in Figure 4.

The complete yearly seasonal variation in daily evaporation/sublimation from permafrost ground surfaces at both the eastern Tibetan Plateau and Tianshan Mountains sites is shown in Figure 6. The general pattern of seasonal variation in evaporation at the two sites was similar, although differences can be seen between the two areas in climatic conditions and soil moisture (Table I). Evaporation occurred predominantly during the summer at both study sites. From October to April, the ground surface was covered by snow (even when uncovered, it was frozen), and sublimation was low or sometimes negative because of condensation. When the ground surface was frozen (from October to April), condensation totaled 13·7 mm at the Tianshan Mountains site, and total sublimation was 63·8 mm. Net sublimation at the site was 50·1 mm and averaged 0·2 mm day^{-1}. Condensation during the same season totalled 5·8 mm at the eastern Tibetan Plateau site, and total sublimation was 77·0 mm. This yields a net sublimation of 71·2 mm, averaging 0·3 mm day^{-1}.

After May, the surface of the permafrost starts to melt and vapour is exchanged by evaporation. Melting water in the soil cannot penetrate very deeply, owing to the underlying impermeable layer. In addition, annual precipitation is concentrated in this period. Daily evaporation increased sharply, reaching a maximum of greater than 5 mm day^{-1}. Daily evaporation from the thawed surface averaged 2·1 mm day^{-1} at both the Tianshan Mountains and eastern Tibetan Plateau sites. These values are 6·4 times the daily evaporation from frozen ground at the Tianshan Mountains site and 4·5 times that at the eastern Tibetan Plateau site. Total annual evaporation was estimated to be approximately 370 mm at the Tianshan Mountains site, based on observations from October 1985 to September 1986, and 388 mm at the eastern Tibetan Plateau site, based on observations from October 1992 to September 1993.

SUMMARY AND DISCUSSION

Evaporation from the ground surface of the Asian cryosphere was observed and estimated using the lysimeter method, the profile method, and the heat budget method at study sites in the eastern Tibetan Plateau region, the Qilian Mountains, the Tianshan Mountains of China, and in Tiksi, eastern Siberia. The simplified lysimeter method provided a reliable measurement of daily evaporation from soil on days with no precipitation. The shortcomings of the lysimeter method on days with precipitation can be avoided by using other methods or by measuring the correlation between evaporation and soil moisture content to estimate evaporation.

Daily mean evaporation from the permafrost surface varied within 0·3 to 3·5 mm. Regional differences was related to surface soil moisture, but variation in evaporation efficiency versus surface soil moisture showed a different tendency. Across small spatial scales, topography altered evaporation by affecting surface soil moisture. Evaporation efficiency decreased logarithmically as elevation increased. The seasonality of ground evaporation in permafrost regions was dominated by the thaw–freeze cycle at the surface. Evaporation from a melting permafrost surface was four to seven times that from the same surface when it was frozen. Interception of precipitation in forests reduced daily evaporation by 60 to 70%. Daily sublimation from snow surfaces was observed to range from 0·2 to 1·0 mm at selected sites. The atmospheric conditions of wind speed and saturation deficit were the dominant factors in the control of snow sublimation.

One of the hydrologic consequences of evaporation is the role that it plays in the water cycle and the water budget. Differences between precipitation and evaporation alter other water-cycle components, such as runoff and soil water content. Using observed daily evaporation data, monthly evaporation was estimated at the study sites, and its relation to monthly precipitation was summarized by calculating the ratio of monthly evaporation to precipitation (E_{month}/Pr; Table V). On a permafrost surface, the ratio of E_{month}/Pr could exceed 100% at the beginning and end of the melting season (May through September). The former result can be explained by the effect of melting water on the evaporation process, but no process observed to date can explain the latter result. On the surface of snow, the ratio of E_{month}/Pr was quite large when precipitation was low, but decreased as precipitation increased.

Table V. Monthly precipitation (*P*r) and evaporation/sublimation (E_{month}) at the study sites

Observation site	Period (y/m)	Surface condition	Pr (mm)	E_{month} (mm)	E_{month}/Pr (%)
Eastern Tibetan Plateau	1989/5	Sparse grass/snow cover	41·7	45·9	110
	1989/6		101·8	91·0	89
	1992/5	Sparse grass		52·7	
	1992/10			42·0	
	1993/5	Sparse grass	16·7	24·8	149
	1993/6		96·7	63·0	65
	1993/7		144·9	83·7	58
	1993/8		133·2	77·5	58
	1993/9		52·7	54·0	102
Qilian Mts	1997/8	Dense grass	98·9	68·2	69
	1997/9		67·7	36·0	53
	1997/10		17·7	12·4	70
	1997/8	Grassland in forest	98·9	21·7	22
	1997/9		67·7	15·0	22
	1997/10		17·7	9·3	53
Tianshan Mts	1986/6	Dense grass	65·5	81·0	124
	1986/7		155·25	99·2	64
	1986/8		92·75	80·6	87
	1991/2	Snow cover	5·5	5·4	99
	1991/3		6·2	4·9	79
	1991/4		42·4	24·5	46
	1991/5		29·7	12·3	33
Tiksi, eastern Siberia	1999/7	Tundra	27·6	65·1	236

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