Environmental and biophysical controls on the evapotranspiration over the highest alpine steppe

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SUMMARY
Characterizing the water and energy flux in the alpine steppe ecosystem in Tibetan Plateau (TP) is of particular importance for elucidating hydrological cycle mechanisms in high altitude areas. In the present study, two years of actual evapotranspiration (ET) values from a semi-arid alpine steppe region (4947 m above sea level) and their environmental and biophysical controls were investigated using the energy balance Bowen ratio energy balance (BREB) method. Seasonally, ET was much lower in frozen soil period and transition period mainly because of low soil water availability. However, ample soil water supplied by rainfall during the rainy period substantially increased ET. The available energy played an important role in controlling ET in the rainy period. Also, the leaf-level stomata closure and plant leaf development through changing bulk surface conductance (G) in rainy period. Similarly, the land–atmosphere energy exchange was dominated by latent heat flux (\(\Delta E\)) in July, but was dominated by sensible heat flux (H) in December and May. Annual ET (plus sublimation) were 362.9 mm and 353.4 mm in the first and second observation year, respectively, which were close to the annual precipitation. On annual scale, the low G (3.30–3.62 mm s\(^{-1}\)), decoupling factor (Q, 0.25–0.27) and the ratio of ET to equilibrium evapotranspiration (\(ET/ET_{eq}\)) corroborated the overall water-limited conditions for the high-altitude alpine steppe. This research provides not only the ground truth data for future hydrological modeling in the data scarce region of TP but also the insights for elucidating how the environmental and biophysical stress factors control the land surface ET in high-altitude region.

1. Introduction

Land surface actual evapotranspiration (ET) is the major variable involved in understanding the interactions between soil–vegetation–atmosphere (Katul et al., 2012). The deficits between precipitation and ET to a large extent determine land surface wetness and plant activity in arid and semi-arid regions (Yang et al., 2006; Zhang et al., 2005). ET is also the key process linking the hydrological cycle with other biogeochemical processes (e.g. carbon and nutrient cycles) (Eamus et al., 2013; Zhou et al., 2015). With the recent advances in micrometeorological instrumentation, quantifying ET in different ecosystems such as forest (Igarashi et al., 2015a; Zhu et al., 2014), grassland (Aires et al., 2008; Qiu et al., 2011; Zhang et al., 2007), wetland (Keliner, 2001; Moore et al., 2013), cropland (Bezerra et al., 2012; Lei and Yang, 2010; Suyker and Verma, 2008) and desert (Ma et al., 2014a; Unland et al., 1996) through in-situ observations has significantly promoted our understanding of the interactions between soil–vegetation–atmosphere as well as the effect of ET in response to global environmental change within different ecosystems.

With an average altitude of more than 4000 m above sea level (asl) and a total area of ~2.5 × 10\(^6\) km\(^2\), the Tibetan Plateau (TP) is known as “the Earth’s Third Pole” (Qiu, 2008). The release of sensible and latent heat over this vast plateau is of primary importance in influencing the atmosphere circulations over China, eastern Asia and even the globe (Wu et al., 2015; Ye and Wu, 1998). Grassland covers more than 60% of the total area of the TP and can be divided into three main subcategories based on different climate and soil types: alpine meadow (including swamp meadow); alpine steppe; and desert steppe (Integrated Scientific Investigation Team of Qinghai-Tibetan Plateau, 1983). Specifically, the total area of alpine steppe in the TP is ~800,000 km\(^2\) (Miehe et al., 2011). It is principally distributed in the frigid and dry areas
of the central and western TP (Babel et al., 2014; Yang et al., 2009a) with altitude ranging from 4500 to 5500 m asl (Wang et al., 2015) (see Fig. 1). Although alpine steppe is characterized by low biomass productivity, due to low temperatures and relatively low precipitation, on the TP it plays a pivotal role in retarding soil desertification and maintaining ecological stability. However, the special geological, topographical and weather conditions accordant with alpine steppe make it ecologically fragile and more sensitive to climate change (Wang et al., 2015; Yu et al., 2012). For example, the experimental research by Klein et al. (2004) confirms that warming may well lead to a severe decline in grassland species diversity in the TP over a relatively short time.

The TP has experienced a much higher rate of warming in comparison with the similar Northern Hemisphere latitudes over recent decades (Yang et al., 2014). Remote sensing data from Moderate Resolution Imaging Spectroradiometer (MODIS) showed that land surface temperatures in the regions above 5000 m asl increased at a rate of up to 0.18 K yr$^{-1}$ from 2000 to 2006 (Qin et al., 2009), much higher than in lower regions. Model simulations projected that these rates of warming in the TP will become more obvious over the coming century (Liu et al., 2009). Interestingly, solar radiation in the TP decreased markedly from 1984 to 2006 due to an increase in deep convective cloud, while precipitation showed a slight increase trend in the same period over most parts of TP (Yang et al., 2014). Nevertheless, the knowledge about how climate change could impact the land surface hydrological processes (e.g. ET and runoff) over the alpine steppe ecosystem is still insufficient.

Due to the harsh environment, quantifying the nature of land–atmosphere interactions in the TP has been greatly impeded by a lack of observational data. Most meteorological stations run by the China Meteorological Administration (CMA) are located in the eastern and southern TP, with few in the west (Qin et al., 2009). Recently the establishment of the Tibetan Observation and Research Platform (TORP) (Ma et al., 2008) would focus on the long-term land surface processes and environment over the heterogeneous underlying surfaces in TP. Within the TORP, a few in-situ observations emphasized on land–atmosphere interaction over different land covers in TP have been launched, e.g. sparse grassland with sand and gravels (Chen et al., 2012), ecotone (Biermann et al., 2014; Gerken et al., 2014) and alpine meadow (Gu et al., 2005, 2008; Ma et al., 2005). However, there is still urgent need of more knowledge on the land–atmosphere energy and water vapor exchange over the alpine steppe ecosystem currently. Basing upon two-year (October 16, 2011 to October 15, 2013) in-situ measured data of an automatic climate observation system (ACOS) over the highest alpine steppe in TP, the objectives of the present study are: (1) to quantify the seasonal and diurnal variations in ET and its role in land surface water budget and (2) to clarify the environmental and biophysical controls on ET over the alpine steppe ecosystem in TP.

2. Materials and methods

2.1. Site description

Serving as one of key stations within the TORP, the Shuanghu Extreme Environment Comprehensive Observation and Research Station of the Chinese Academy of Sciences initiated the ACOS which aimed at research into long-term continuous land–atmosphere interactions over a semi-arid alpine steppe area of TP in September 2011. Located in the hinterland of Qiangtang plateau, the altitude of present ACOS is 4947 m asl. Since the glacier equilibrium line altitude in this area is roughly 5800 m asl (Yao et al., 2012), the present research represents the highest altitude in-situ observations made in the alpine steppe ecosystem. This region is characterized by a frigid semi-arid climate (Zheng et al., 2013). Climate data for the period 1981 to 2010 from the nearest CMA station at Bange (altitude is 4719 m asl, distance from in-situ ACOS is 230 km), shows that the annual mean temperature was –3.3 °C, with monthly averages of –10.1 °C for January and 9.1 °C for July, and that the average annual precipitation was 333 mm, falling mainly between June and September. However, the annual potential ET was approximately 1000 mm (Wang et al., 2013). The observation site is homogenously flat with a fetch of >1 km for the prevailing wind direction. Laboratory analysis suggests that the soil belongs to sandy loam. The average bulk density at the surface layer (0–10 cm) is 1.29 g cm$^{-3}$. The dominant vegetation type is cold-xerophytic C3 grasses such as stipa purpurea and carex moorcroftii. The mean canopy height in growing season is 0.03 m. The maximum above-ground biomass is estimated as ca. 50.1 g m$^{-2}$ in summer (Yang et al., 2009a), with patchy plant distribution over the ground surface from October to May of the ensuing year.

![Fig. 1. The geographical sited (black square) and photo of the in-situ observational instruments over an alpine steppe area in the Tibetan Plateau.](image_url)
2.2. Measurements

Four air temperature/humidity sensors (HMP45C, Vaisala Inc., Finland) were installed at heights of 0.7 m, 1.5 m, 2 m and 4 m. Four wind speed sensors (010C-1, MetOne Inc., USA) were also stationed at these heights. Wind direction (020C-1, MetOne Inc., USA) was measured at a height of 4 m. Air pressure (PTB100, Vaisala Inc., Finland) was measured at 0.5 m above the land surface. Downward and upward long- and short-wave radiation (CNR4, Kipp & Zonen Inc., Netherlands) were measured at a height of 0.7 m. Soil temperatures (109, Campbell Scientific Inc., USA) were measured at the land surface (0 cm) and 5 cm deep. A soil heat flux plate (HF010SC, Hukseflux, Netherlands) was buried at a depth of 0.03 m. All observations were averaged over 10-min from ca. 4 s readings and stored in the CR1000 datalogger (Campbell Scientific Inc., USA). The rainfall was measured by an automatic-storage tipping-bucket rain gauge (SML-3, Shanghai Meteorological Instrument Factory Inc., China).

Since we did not measure the leaf area index (LAI) in field from October 16, 2011 to October 15, 2013, we used normalized difference vegetation index (NDVI) data based upon 250 m resolution MODIS 16-day NDVI composite images (MOD13Q1) (aimed at the pixel that contained our observation site) during this period. These images were from the NASA Land Processes Distributed Active Archive Center (http://reverb.echo.nasa.gov) at the USGS/Earth Resources Observation and Science Center.

2.3. Flux calculation and fetch estimating

All raw 10-min data were firstly processed into a half-hourly average. Since the soil heat flux plate only provided soil heat flux readings at the depth of 0.03 m, the integrating soil temperature method (Zhang et al., 2011) was applied to estimate heat storage on the land surface to 0.03 m deep. Given that advection may be neglected in a homogenous area with appropriate fetch (see flux footprint analysis below), and that heat storage in plants would be very small in an area covered by very short grasses, the land surface energy balance equation can be simplified as:

\[ R_n - G - H - \lambda E = 0 \]  

(1)

where \( R_n \) is the net radiation (W m\(^{-2}\)), \( G \) is the soil heat flux into the land surface (W m\(^{-2}\)), \( H \) is the sensible heat flux (W m\(^{-2}\)) and \( \lambda E \) is the latent heat flux (W m\(^{-2}\)).

The Bowen ratio can be expressed as:

\[ \beta = \frac{H}{\lambda E} = \frac{\rho_a C_p R_n[T_2 - T_1 + \tau(z_2 - z_1)]}{\rho_a C_p K_r(\varepsilon_2 - \varepsilon_1)/\gamma} = \frac{\gamma T_2 - T_1 + \tau(z_2 - z_1)}{\varepsilon_2 - \varepsilon_1} \]  

(2)

where \( \beta \) is the Bowen ratio, \( \rho_a \) is the mean air density at constant pressure (kg m\(^{-3}\)) and \( C_p \) is the specific heat capacity of air at constant pressure (J kg\(^{-1}\) K\(^{-1}\)). \( K_r \) and \( \gamma \) are the turbulent exchange coefficients for \( H \) and \( \lambda E \), respectively, and usually assumed to be equal. \( \tau \) is the adiabatic lapse rate, which is normally taken as 0.01 K m\(^{-1}\). \( \gamma \) is the psychrometric constant (kPa K\(^{-1}\)). \( T_1 \) and \( T_2 \) are the air temperature (K) at \( z_1 \) and \( z_2 \), respectively; and \( \varepsilon_1 \) and \( \varepsilon_2 \) are vapor pressure (kPa) at \( z_1 \) and \( z_2 \), respectively. Normally, the larger height difference between the two levels, the larger temperature (\( T_2 - T_1 \)) and humidity (\( \varepsilon_2 - \varepsilon_1 \)) difference, and the influence of measuring error therefore decrease. As a result, Foken (2008) recommended that the ratio of two heights of measuring the temperature and humidity should be 4–8. Keeping this in mind, the \( z_1 \) and \( z_2 \) in the present study were chose as 0.7 m and 4 m, respectively.

Combining Eqs. (1) and (2), the latent heat flux \( \lambda E \) on snow free days was derived using the Bowen ratio energy balance (BREB) method for 30-min intervals, thus:

\[ \lambda E = \frac{R_n - G}{(1 + \beta)} \]  

(3)

It should be noted that the \( G \) in Eq. (1) is difficult to estimate on snow cover days because of the uncertainties surrounding heat flux in the snow. Consequently, for days with snow cover, the BREB method was replaced with the flux-profile method (Brutsaert, 2005) for calculating the \( \lambda E \). To identify whether the observation site is snow free or not, the albedo, defined as the ratio of upward shortwave radiation to downward shortwave radiation was used (daily mean albedo was calculated as average of half-hourly albedo from 10:00 to 14:00, local time). Based on the clear variation of albedo observation (see Fig. 2), a threshold of 0.26 was set to differentiate the snow free (<0.26) and snow cover (>0.26) days. Then, the latent heat of vaporization, \( \lambda_v \) (J kg\(^{-1}\)), for snow free days and the latent heat of sublimation, \( \lambda_s \) (J kg\(^{-1}\)), for snow cover days were calculated using Eqs. (4) and (5) (Zhu et al., 2014), and the results were then used to calculate the \( ET \) or sublimation from \( \lambda E \):

\[ \lambda_v = (2500 - 2.4(T_a - 273.15)) \times 10^3 \]  

(4)

\[ \lambda_s = (2834.1 - 0.149(T_a - 273.15)) \times 10^3 \]  

(5)

where \( T_a \) is the air temperature measured at 2 m high.

The flux footprint calculated using the BREB method was evaluated using Hsieh et al. (2000) model, which is based upon a combination of Lagrangian stochastic dispersion model results and dimensional analyses. Taking the measurement height as the geometric mean (1.67 m) of the abovementioned two heights (0.7 m and 4 m) of the temperature and humidity sensors, the 90% ratio of fetch to measurement height as calculated by the footprint model of Hsieh et al. (2000) is 124, demonstrating that the 90% flux contribution area was limited to a distance of 204 m away from the tower. Since the land surface’s homogeneity in this fetch is stable enough to satisfy the K theory principle which states that large homogenous surfaces are essential for guaranteeing vertical turbulence flux, the BREB method therefore provided reasonable estimation of the \( \lambda E \) flux in the area studied.

2.4. Quality control and gap filling

To ensure the high reliability of the half-hourly \( \lambda E \) calculated using BREB, three steps of quality control were used. First, obvious outliers were removed when \( \lambda E > 700 \) W m\(^{-2}\) or \( \lambda E < -200 \) W m\(^{-2}\). Second, the BREB method fails when \( \beta > 1 \) due to the denominator approaching zero in Eq. (3) (Brutsaert, 2005). We therefore filtered out the \( \lambda E \) using typical criteria of \( -1.3 < \beta < -0.7 \) (Kurc and Small, 2004; Unland et al., 1996). Third, to determine whether the signs of \( \lambda E \) estimated by BREB method are in keeping with the flux-gradient relationships, Perez et al. (1999) proposed the following criteria:

If \( R_n - G > 0 \), then when \( \Delta e > 0 \), then \( \beta > -1 \) and \( \lambda E > 0 \); when \( \Delta e < 0 \), then \( \beta < -1 \) and \( \lambda E < 0 \)

(6)

If \( R_n - G < 0 \), then when \( \Delta e > 0 \), then \( \beta < -1 \) and \( \lambda E > 0 \); when \( \Delta e < 0 \), then \( \beta > -1 \) and \( \lambda E < 0 \)

(7)

where \( \Delta e \) is the vapor pressure difference between the lower and upper measurement levels. The \( \lambda E \) were filtered out if they failed to meet the criteria in (6) or (7). Totally, 18.2% of the half-hourly \( \lambda E \) during these two years were excluded using these three quality control processes. Most gaps occurred in the late afternoon, early morning and occasionally at night. While this does not result in substantial errors in the calculation of daily \( ET \), these gaps were then filled using the well-known crop coefficient method on a day-by-day basis with the aid of reference evapotranspiration (\( ET_{ref} \)) (Allen et al., 1998). Specifically, we first calculated the daily
crop coefficient using the remainder $\lambda E$ for a specific day, then filled the half-hourly gaps by multiplying the half-hourly $ET_{ref}$ and the crop coefficient for that day.

2.5. Deviation in bulk surface parameters

Land surface temperature ($T_s$, K) is derived from the observed long-wave radiation data:

$$\varepsilon \sigma T_s^4 = R_{\text{down}} - (1 - \varepsilon)R_{\text{up}}$$

in which $\varepsilon$ is the surface emissivity, $\sigma$ is the Stefan–Boltzmann constant ($5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$), $R_{\text{up}}$ and $R_{\text{down}}$ are the observed upward and downward long-wave radiation (W m$^{-2}$), respectively.

For partial vegetation cover, the bulk surface conductance ($G_s$) was obtained by inverting the Penman–Monteith equation (Monteith and Unsworth, 2013), thus:

$$\frac{1}{G_s} = \frac{1}{G_a} \left(\frac{A}{\gamma \beta} - 1\right) + \frac{\rho_s C_p VPD}{\gamma \lambda E}$$

(9)

where $A$ is the slope of the vapor pressure curve at air temperature (kPa K$^{-1}$), $\gamma$, $\beta$, $\rho_s$ and $C_p$ are remains as Eq. (2); VPD is the vapor pressure deficit (kPa) and $\lambda E$ is the observed latent heat flux (W m$^{-2}$). Note the unit of $G_s$ calculated by Eq. (9) and $G_a$ in Eq. (10) is expressed using m s$^{-1}$; this can be transferred into mm s$^{-1}$ in the text which follows by multiplying by 1000 so as to be in accordance with other research.
\( G_s \) in Eq. (9) is the aerodynamic conductance as derived from the Monin–Obukhov similarity theory (Monteith and Unsworth, 2013), thus:

\[
\frac{1}{G_s} = \frac{k}{u^*} \ln \left( \frac{Z_m - \psi_m}{Z_m - \psi_h} \right) 
\]

(10)

\[
u = \frac{kU}{\ln \left( \frac{Z_m - \psi_m}{Z_m - \psi_h} \right)}
\]

(11)

where \( k \) is the von Karman constant (0.4), \( u^* \) is the frictional wind speed (m s\(^{-1}\)), \( Z_m \) and \( Z_h \) are the air temperature (\( T_m \)) and wind speed (\( U_m \)) measurement height (m), respectively. In present study we used the temperature and wind speed data measured at 2 m high, that is, \( Z_m = Z_h = 2 \) m. \( Z_m \) is the roughness length for momentum transfer (m) and can be assumed to be \( h/8 \) (Brutsaert, 2005).

According to our fieldwork, the mean canopy height in the rainy periods (see Section 3.1 for periods dividing) is \( h = 0.03 \) m, \( Z_m = h/8 = 3.75 \times 10^{-3} \) m. For frozen soil and transition periods, the land surface can be assumed as bare soil, the \( Z_m \) being given as \( 5 \times 10^{-4} \) m (Wierenga, 1993). \( \psi_m \) and \( \psi_h \) are stability correction functions for momentum and heat transfer, respectively. In neutral conditions, \( \psi_m = \psi_h = 0 \); in non-neutral conditions, these two variables can be calculated using universal functions in Högström (1996) and the mathematical form of the correction terms in Paulson (1970), thus:

For stable conditions:

\[
\psi_m = -5.3(Z_m - Z_{om}) / L
\]

(12)

\[
\psi_h = -8.0(Z_h - Z_{oh}) / L
\]

(13)

For unstable conditions:

\[
\psi_m = 2 \ln \left( \frac{1 + x}{1 + x_0} \right) + \ln \left( \frac{1 + x^2}{1 + x_0^2} \right) - 2 \tan^{-1} x + 2 \tan^{-1} x_0
\]

(14)

\[
\psi_h = 2 \ln \left( \frac{1 + y}{1 + y_0} \right)
\]

(15)

where \( x = (1-19Z_m/L)^{0.25}, x_0 = (1-19Z_{om}/L)^{0.25}, y = (1-11.6Z_h/L)^{0.5}, \) and \( y_0 = (1-11.6Z_{oh}/L)^{0.5} \). \( L = T_m a^2 / (kg T) \) is the Obukhov length (m), where \( g = 9.8 \) m s\(^{-2}\) and see below for \( T_m \).

\( Z_{om} \) in Eq. (10) is the roughness length for heat transfer (m). An efficient parameterization scheme for \( Z_{oh} \) of Yang et al. (2008, 2009b) has been widely used in TP (Biermann et al., 2014; Chen et al., 2010), i.e.

\[
Z_{oh} = \frac{70 \nu}{u^*} \exp \left( -7.2u^* / |T_r|^{0.25} \right)
\]

(16)

where \( \nu \) is the fluid kinematic viscosity and equals to \( 1.328 \times 10^{-5} \) \( (P_o/P_m)^{0.79} \) with \( P_o = 101.3 \) kPa, \( T_o = 273.15 \) K, \( P \) is air pressure (kPa), \( T_o \) is air temperature (K); \( T_r \) is the frictional temperature (K) and equals to \( -(\theta_m - \theta_h) / \ln (Z_m/Z_h) - \psi_h \), where \( \theta_h = T_h \) in Eq. (8) and \( \theta_m = T_m + Z_{om} g / (c_p) \) are potential temperatures (K). Since the \( T \) depends on the \( Z \), Eqs. (10)-(16) have to be solved iteratively for getting \( G_s \) (see detailed procedure in Yang et al. (2009b, page 2477)).

The degree of atmosphere–vegetation interaction can be estimated using a dimensionless value called the decoupling factor \( \Omega \) (Jarvis and McNaughton, 1986), thus:

\[
\Omega = (A/\gamma + 1) / (A/\gamma + 1 + G_s / G_o)
\]

(17)

The value of \( \Omega \) takes the range of 0–1. When \( \Omega \to 1 \), the land surface is completely decoupled from overhead conditions, hence the \( ET \) is mainly controlled by available energy; when \( \Omega \to 0 \), the \( ET \) is mainly controlled by bulk surface conductance and atmospheric humidity deficit (Baldocchi and Xu, 2007).

To quantify whether atmospheric demand or terrestrial moisture supply is the limiting factor, the \( ET \) is compared to the equilibrium evapotranspiration, \( ET_{eq} \), thus:

\[
\frac{ET}{ET_{eq}} = \frac{\Delta + \gamma \Delta}{\Delta + H + \Delta E}
\]

(18)

Eq. (18), in fact, normalizes the site \( ET \) values against the \( ET_{eq} \) which is primarily determined by available energy. In some cases, the potential evapotranspiration could be defined as 1.26 times of \( ET_{eq} \) (Baldocchi et al., 1997; Baldocchi and Xu, 2007), while this value is actually rare to meet in practical observations over dry environment.

In terms of the bulk surface parameters, we first filtered out the half-hourly \( G_s \), \( \Omega \) and \( ET/ET_{eq} \) values into three categories: (i) during periods of rainfall; (ii) within 1 h after the rain; and (iii) on snow cover days. The midday averages of \( G_s \), \( \Omega \) and \( ET/ET_{eq} \) were then calculated using half-hourly values from 10:00 to 15:00, local time. The days whose all midday half-hourly \( G_s \), \( \Omega \), \( ET/ET_{eq} \) values were excluded (mainly in winter) were neither gap-filled nor taken into account during the calculation of periodical average values.

Following Allen et al. (1998), the daily \( ET_{ref} \) were also calculated to provide another background of atmospheric demand. Note that although the \( ET_{ref} \) is based on a hypothetically optimal-watered vegetated surface, the bulk surface conductance is fixed as 14.3 mm s\(^{-1}\) (Allen et al., 1998) rather than 0. Theoretically, the \( ET_{ref} \) may be intending to close to the potential evapotranspiration, in practice, the observed daily \( ET \) may be able to exceed the calculated \( ET_{ref} \) if the water supply is not short (due to the different properties between the practical and hypothetical vegetated surfaces).

3. Results

3.1. Meteorological conditions

Fig. 2 illustrates the main meteorological conditions during the observation period. Both the highest (lowest) daily values of \( T_m \) and \( T_h \) occurred in July (January). Due to the high altitude, the minimum (maximum) daily \( T_m \) was 249.6 K (284.3 K) in 2012 and 252.6 K (283.6 K) in 2013, respectively. Conversely, wind speed at 2 m above ground was higher in winter and lower in summer. The daily mean net radiation \( (R_n) \) fluctuated significantly from day to day because of frequent cloudy and rainy weather conditions. The daily albedo showed obvious variation in winter because of snow disturbance. It reached \( >0.7 \) for fresh snow cover days (Fig. 2). The albedo values were, however, much lower in summer, due both to the growth of vegetation and to higher soil moisture content. Rain fell mainly from June to September. In 2012, there were 85 rainy days (defined as at least 0.1 mm precipitation) during these months, with total precipitation of 325.4 mm; in 2013 there were 62 rainy days with a total rainfall of 289.7 mm for the same months. The annual maximum daily rainfall was 23.3 mm in 2012 and 35.8 mm in 2013, respectively. The NDVI remained at ca. 0.1 from November to May of each ensuing year, and then gradually increased from June, peaking in July or August with values of ca. 0.25.

In general, routine meteorological elements showed little interannual variation during the two-year observation cycle, except for rain. This variation occurred in August 2013, when the 70.4 mm which fell was only 61.4% of the 114.7 mm for August 2012 (Fig. 2). According to the above analysis of the temperature and rain regimes, we could divide each observational year into three periods (n.b. in this study we define the first observational year as from October 16, 2011 to October 15, 2012 and the second...
observational year as from October 16, 2012 to October 15, 2013), namely: a frozen soil period (from October 16 to April 15 of the ensuing year), whose soil temperatures are mainly <273.15 K; a rainy period (from June 1 to September 30), characterized as the period when most of the precipitation falls; and two, short transition periods (from April 16 to May 31, and from October 1 to October 15), characterized by low precipitation and very little frozen soil.

3.2. Seasonal variations in evapotranspiration

The majority of daily ET values were <0.3 mm (Fig. 3) for the frozen soil period. During the snow cover days, the snow sublimation varied from 0.1 mm d\(^{-1}\) to 1.4 mm d\(^{-1}\). Total sublimation was 16.5 mm and 12.1 mm for the first and second years, respectively. Overall, the daily ET in frozen soil period and transition period were lower than the daily ET\(_{\text{ref}}\) (Fig. 3). With the onset of the rainy period, the ET increased dramatically and approached or even exceeded the ET\(_{\text{ref}}\). The mean daily ET in rainy period were 2.1 mm d\(^{-1}\) in both observational years, much greater than that of the two other periods (Table 1). The maximum daily ET values were 4.1 mm d\(^{-1}\) and 4.2 mm d\(^{-1}\) for the first and second years, respectively. At the end of the rainy period, ET decreased very gradually in response to water deficiency and vegetation senescence. Overall, the total ET in the rainy period accounted for 71.5% (first year) and 72.6% (second year) of annual ET & sublimation, respectively. This proportion decreased to 11.9–14.1% for the frozen soil period and 14.4–15.5% for the transition periods (Table 1).

3.3. Diurnal variations in latent heat flux

To clarify, the December, May and July were chose as representative examples to investigate the diurnal variation of \(\Delta E\) and \(H\) during the frozen soil period, transition period and rainy period, respectively (Fig. 4). Visibly, the \(\Delta E\) increased during the morning and peaked at noon, then decreased gradually, with stable, low values at night. The mean nocturnal \(\Delta E\) values in the December and May were close to zero, while reached ca. 15 W m\(^{-2}\) for July. On average, the peak value \(\Delta E\) at noon reached >200 W m\(^{-2}\) for July, but only ca. 70 W m\(^{-2}\) and 10 W m\(^{-2}\) for May and December, respectively. The diurnal variations of \(H\) in each month are similar to \(\Delta E\) in patterns, but obviously different in values. In contrast to the \(\Delta E\), the maximum of December and May average half-hourly \(H\) could be more than 300 W m\(^{-2}\) and 200 W m\(^{-2}\), respectively. While the highest July average half-hourly \(H\) were only ca. 150 W m\(^{-2}\). On the whole, \(H\) were much larger than \(\Delta E\) in December and May, while became reverse in July (Fig. 4).

3.4. Midday mean bulk surface parameters

Fig. 5 illustrates the variations in midday mean \(G_c\), ET/ET\(_{eq}\) and \(\Omega\) in snow free days at the observation site from October 16, 2011 to October 15, 2013. Generally, \(G_c\) showed evident variation during the year. It was mainly <1 mm s\(^{-1}\) in the frozen soil period due to low soil water availability. In the transition period, although plant had not begun to grow (NDVI were as low as those in winter), the possible melted snow that were stored in the soil during winter and occasional rain events replenished the soil water availability, hence the \(G_c\) increased gradually. With the onset of the rainy period, the increase in soil moisture and plant physiological activity led to higher \(G_c\), with maxima of 24.6 mm s\(^{-1}\) and 27.8 mm s\(^{-1}\) in the first and second years, respectively (Fig. 5). Average rainy period \(G_c\) were 8.04 mm s\(^{-1}\) and 6.87 mm s\(^{-1}\) for the first and second year, respectively, which were much higher than that for frozen soil period and transition period (Table 1).

During the frozen soil period, \(\Omega\) always remained ca. 0.1 and \(ET/ET_{eq}\) were usually lower than 0.2 due to extremely low soil water availability (Fig. 5). As expected, \(\Omega\) and ET/ET\(_{eq}\) increased gradually during the transition period and peaked in the rainy period. The average of \(\Omega\) and ET/ET\(_{eq}\) in transition period of first (second) year are 0.22 (0.23) and 0.28 (0.30), respectively, while reached 0.56 (0.49) and 0.72 (0.63) in rainy period, respectively (Table 1). The annual maximum \(\Omega\) was >0.80 in both years (Fig. 5). Similarly, the ET/ET\(_{eq}\) exceeded 1 to reach the equilibrium evapotranspiration during the days with ample soil moisture. However, the maximum midday mean ET/ET\(_{eq}\) was only 1.18 during two-year observation. The annual average ET/ET\(_{eq}\) were 0.35 and 0.34 for first and second year, respectively. And the annual average \(\Omega\) were 0.27 and 0.25 for the first and second years, respectively (Table 1), indicating the ET were tightly coupled with \(G_c\) and VPD rather than available energy on annual scale.

It should be noted that the term ‘rainy period’ does not mean very frequent rain events all the time, although \(\Omega\) values were on the whole larger in this period. For example, there was only one rain event, of 5.7 mm, from July 28, 2013 to August 15, 2013 (Fig. 2). ET decreased fast in response to this scarce-precipitation-period (Fig. 3). The \(G_c\) (down from 8.51 mm s\(^{-1}\) in July 28, 2013 to 1.75 mm s\(^{-1}\) in August 15, 2013) and \(\Omega\) values (down from 0.58 to 0.27 over the same period) substantially decreased over these 19 days (Fig. 5), demonstrating a rapid transfer process from available energy control to bulk surface conductance and VPD control for ET. However, \(G_c\) and \(\Omega\) recovered to high values after this interval due to rainfall supplementing soil moisture in the subsequent days.

Fig. 3. The daily ET (blue circles), sublimation (red circles) and ET\(_{eq}\) (gray area) values from October 16, 2011 to October 15, 2013 at the observation site. The F, T and R denote the frozen soil period, transition period and rainy period, respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
4. Discussions

4.1. Uncertainties in ET derived from the BREB method

First proposed by Bowen in 1926, the BREB method has been widely employed to measure the ET from various ecosystem types ranging from open water (Burba et al., 1999; Rohli et al., 2004) to the Brazilian Pantanal Vochysia forest (Sanches et al., 2011), spruce-fir-beech forest (Bernhofer, 1992), short-rotation poplar (Fischer et al., 2013), cotton field (Bezerra et al., 2012), irrigated alfalfa (Todd et al., 2000), sparse sorghum (Kato et al., 2004), irrigated groundnut (Kar and Kumar, 2007), vineyard (Heilman Table 1

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<tbody>
<tr>
<td></td>
<td>Frozen soil period</td>
<td>Transition period</td>
</tr>
<tr>
<td>$T_a$ (°C)</td>
<td>262.3</td>
<td>271.1</td>
</tr>
<tr>
<td>$T_s$ (°C)</td>
<td>263.5</td>
<td>274.5</td>
</tr>
<tr>
<td>VPD (kPa)</td>
<td>0.22</td>
<td>0.30</td>
</tr>
<tr>
<td>Rainfall (mm)</td>
<td>2.4</td>
<td>24.7</td>
</tr>
<tr>
<td>$ET_{plus \ sublimation}$</td>
<td>51.2</td>
<td>52.2</td>
</tr>
<tr>
<td>$R_n$ (W m$^{-2}$)</td>
<td>45.6</td>
<td>116.2</td>
</tr>
<tr>
<td>$G_s$ (mm s$^{-1}$)</td>
<td>4.97</td>
<td>4.2</td>
</tr>
<tr>
<td>$G_s$ (mm s$^{-1}$)</td>
<td>0.93</td>
<td>2.46</td>
</tr>
<tr>
<td>$G_s$ (mm s$^{-1}$)</td>
<td>0.08</td>
<td>0.22</td>
</tr>
<tr>
<td>$ET/ET_{eq}$</td>
<td>0.13</td>
<td>0.28</td>
</tr>
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</table>

Fig. 4. The December, May and July average of the half-hourly $\lambda E$ and $H$ for the local time zone in each year at observation site.

Fig. 5. Variations in midday mean (10:00 to 15:00, local time) bulk surface conductance ($G_s$), decoupling factor ($\Omega$) and $ET/ET_{eq}$ on snow free days from October 16, 2011 to October 15, 2013, at the observation site. The upper F, T and R denote the frozen soil period, transition period and rainy period, respectively.
et al., 1994; Zhang et al., 2008), lemon orchard (Daamen et al., 1999), prairie wetland (Burba and Verma, 2001), red bed (Peacock and Hess, 2004), tall-grass prairie (Bremer and Ham, 1999), temperate steppe (Qiu et al., 2011), desert (Unland et al., 1996) and sandhill ecoregions with sub-irrigated meadows (Billesbach and Arkebauer, 2012). It is worth pointing out that the more widespread application for land–atmosphere energy exchange research currently is the eddyy covariance (EC) technique (Baldocchi, 2014). In addition to its high cost, however, the EC also requires extensive maintenance. This presents practical difficulties in this study’s alpine steppe area, which is located at an altitude of nearly 5000 m asl. The BREG method is therefore a convenient choice. While the BREG method may involve errors, it is still regarded as one of accurate way for measuring $ET_{a}$ (Burba, 2013; Wang and Dickinson, 2012). Indeed, the potential uncertainties of the BREG method should be acknowledged. The first uncertainty may stem from assuming the time-constant equality of the turbulence exchange coefficient of heat, $K_h$, and vapor, $K_v$, in Eq. (2) (Foken, 2008), especially when advection influence is significant (Bertela, 1989). But McNaughton and Labaukh (1998)’s experiment implied that the violation of $K_h=K_v$ was minor and only cause slight errors in estimating latent heat flux; this was supported by Dias and Brutsaert (1996), who argued that $K_h=K_v$ is theoretically realistic under a wide range of stable conditions. Another potential problem is the random and systematic errors given by the measurement sensor itself, which in turn may also result in some uncertainty in the estimation of the Bowen ratio. Still, a recent inter-comparison by Savage (2010) endorsed that the BREG system with HMP45C sensors is reasonable for measuring the vertical vapor pressure differences to calculate Bowen ratio. Although rigorous data filtering and gap filling using the flux-profile relationship and crop coefficient were used in this study to minimize errors in calculating the flux, we concur with the ca. ±15% uncertainty range in $ET$ estimation using the BREG method proposed by other studies (Kabat et al., 1997; Kurc and Small, 2004; Zhu et al., 2003) when compared with EC method.

4.2. Comparisons of energy partitioning with other studies in TP

It is not surprised for the obvious seasonal variations of the energy partitioning, like the present study (Fig. 4), in the alpine steppe and alpine meadow ecosystems in central and eastern TP. That is, the $E$ plays a main role in summer, while the $H$ plays a dominating role in other seasons in the land–atmosphere energy exchange process. In other words, only the rainy period could bring out large $ET$ values. The in-situ observations at Tanggula (Yao et al., 2008) in Northern TP, Haibei (Gu et al., 2005) in Northeastern TP, Qmado (Bian et al., 2002) as well as Maqu (Shang et al., 2015) in Eastern TP and Naqu (Choi et al., 2004; Ma et al., 2005) in central TP all illuminated above-mentioned seasonal transfer of the energy partitioning dominated by $H$ or $E$. Furthermore, remote-sensing-based investigations by Ma et al. (2014b) demonstrated the evaporative fraction [defined as $E/(H+E)$] in most part of TP was >0.7 in summer, while only ca. 0.3 in winter. Note that, however, this might not be the case of some regions in western TP where arid desert steppe distributed. The annual precipitation over there is usually less than 100 mm and this makes the $H$ consistently dominated in the energy partitioning regardless of the advent of rainy season, as was observed by Li et al. (2003) in Shiquanhe and Bian et al. (2003) in Gerze, both belong to the western part of TP. In addition to the climate, the land use is also another factor that impacts the energy partitioning. For example, recent research at the bare soil/gravel surface in the Qomolangma region of southern TP by Li et al. (2014a) suggested that the sensible heat flux is also the main component of energy partitioning in all seasons.

4.3. Environmental and biophysical controls on $ET$

4.3.1. Water and energy availability

The low values of annual mean $ET/ET_{eq}$ and $\Omega$ evidently showed that the alpine steppe was under water-limited condition rather than energy-limited condition on annual scale. This because the annual precipitation is less than 400 mm in the area that alpine steppe distributed (Gao and Liu, 2013). However, the wind speed and solar radiation in TP are generally larger than the lower-altitude region (Yang et al., 2014). Present water-limited condition was similar to the semi-arid grasslands in East Asia (Chen et al., 2009; Hao et al., 2007; Zhang et al., 2007), temperate grassland in Southern Canada (Wever et al., 2002; Zha et al., 2010) and Mediterranean grasslands under drought in Portugal (Aires et al., 2008) and in California (Ryu et al., 2008), where values have been documented to have ranged from 0.36 to 0.59 for $ET/ET_{eq}$ and from 0.20 to 0.33 for $\Omega$, but these values are much less than those in wet temperate grassland of Japan (Li et al., 2005) and tropical rice cropland of Philippines (Alberto et al., 2011), where water supplies are plentiful.

Considering the general shallow effective root depth and lower stomatal control compared with forest ecosystems for alpine steppe, such strong coupling between the $ET$ and soil water availability on annual scale indicates that the possible increased drought frequency and severity under a warming climate may substantially cause negative influences on the alpine steppe ecosystem. Fortunately, however, the annual precipitation showed a little raising trend in most non-humid regions of TP since 1980s (e.g. Li et al., 2014b; Yang et al., 2014). As result, the $ET$ also slightly accelerated in the arid and semi-arid regions of TP in past few decades (Yin et al., 2013; Yang et al., 2014), suggesting an intensified hydrological cycle in this high-altitude area. Moreover, the enhanced cooling effect due to such increased $ET$ was also able to mitigate the regional warming rate in TP by reducing the sensible heat that would have elevated air temperature (Shen et al., 2015).

On seasonal scale, the $ET/ET_{eq}$ and $\Omega$ increased dramatically during the rainy period, due to the ample soil water supplied by the rain, and decreased gradually with the onset of the transition and frozen soil periods (Fig. 5), suggesting that seasonal variations in $ET$ are in accordance with precipitation which in turn determines soil water availability and plant growth in the alpine steppe (Figs. 2 and 3). Low $\alpha$ and $\Omega$ values in frozen soil period and transition period are in line with the relationship between $ET_{et}$ and $ET$ (Fig. 3), and this also confirms the low $ET$ values in these two periods resulted from the restriction of water supply rather than the low atmospheric demand or available energy.

On the other hand, the annual maximum daily $ET$ rates in the present alpine steppe were within the scope of values estimated for other grasslands which range from 3 mm d$^{-1}$ to 5.5 mm d$^{-1}$ (e.g. Aires et al., 2008; Gu et al., 2008; Hao et al., 2007; Krishnan et al., 2012; Kurc and Small, 2004; Wever et al., 2002). Moreover, during the 121-day span of each rainy period, there were in total 35 and 26 days whose $\Omega > 0.7$ in first and second year, respectively (Fig. 5), suggesting the $ET$ became to be more coupled with available energy and a switch to temporary energy-limited condition did occur in these days of rainy period at the present alpine steppe. With generally cool temperatures and frequent cloud cover in the summer season, much precipitation which enlarged soil water availability is the main reason for this phenomenon.

Taking the rainy period individually as an example, a brief comparison of $ET$ and bulk surface parameters between different categories of grassland ecosystems (range from humid to arid) is presented in Table 2. It can be seen that the soil water availability was still the dominated factor that controls the $ET$ in arid grasslands, while the available energy plays a much more important role in determining the $ET$ in humid and semi-humid grasslands.
Table 2

<table>
<thead>
<tr>
<th>Grasslands Period</th>
<th>Total ET (mm)</th>
<th>ET/ET$_{eq}$</th>
<th>Total ET$_{b}$ (mm)</th>
<th>Daily mean $k_E$ (W m$^{-2}$ s$^{-1}$)</th>
<th>Average of midday mean decoupling factor</th>
<th>Average of midday mean ratio of ET to equilibrium evapotranspiration (ET/ET$_{eq}$)</th>
<th>Maximum LAI</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wet temperate grassland in Japan (1300 mm)$^a$ June 16–September 27, 1999 (09:00–16:00)$^b$</td>
<td>559.3</td>
<td>0.03</td>
<td>109.7</td>
<td>6.2</td>
<td>0.73</td>
<td>0.73</td>
<td>5.6</td>
<td>Li et al. (2005)</td>
</tr>
<tr>
<td>Alpine meadow in Tibetan Plateau (567 mm)$^a$ May 21–September 20, 2002 (10:15–14:15)$^b$</td>
<td>394.7</td>
<td>0.38</td>
<td>60.2</td>
<td>60.2</td>
<td>0.76</td>
<td>0.76</td>
<td>2.4</td>
<td>Gu et al. (2005)</td>
</tr>
<tr>
<td>Mediterranean annual grassland in California (556 mm)$^a$ January 1–April 30, 2002–2007 (10:00–14:00)$^b$</td>
<td>187.8</td>
<td>0.35</td>
<td>44.4</td>
<td>60.2</td>
<td>0.51</td>
<td>0.51</td>
<td>2.3</td>
<td>Ryu et al. (2008)</td>
</tr>
<tr>
<td>Alpine steppe in Heihe River Basin (335 mm)$^a$ June 1–September 30, 2006 (11:00–15:00)$^b$</td>
<td>184</td>
<td>0.35</td>
<td>44.4</td>
<td>60.2</td>
<td>0.51</td>
<td>0.51</td>
<td>2.3</td>
<td>Chen et al. (2009)</td>
</tr>
<tr>
<td>And grazing steppe in Mongolia (187 mm)$^a$ June 1–September 30, 2003 (11:00–15:00)$^b$</td>
<td>244</td>
<td>0.35</td>
<td>44.4</td>
<td>60.2</td>
<td>0.51</td>
<td>0.51</td>
<td>2.3</td>
<td>Li et al. (2006)</td>
</tr>
</tbody>
</table>

* Multiyear annual precipitation for providing general climate backgrounds.  
* The local time for calculating the midday mean.  
* 6 years average values.  
* Manually digitized from the Figure 1 of Ryu et al. (2008).  
* Only mean LAI is available.

4.3.2. Physiological control via surface conductance

The annual average $G_o$ in the present alpine steppe were 3.62 mm s$^{-1}$ and 3.30 mm s$^{-1}$ for the first and second years (Table 1), respectively, which are close to values for semi-arid prairie in Canada (Zha et al., 2010) and typical steppe in Inner Mongolia (Chen et al., 2009), but much lower than those for alpine meadow in TP (Gu et al., 2005), peatland in Michigan (Moore et al., 2013) and, of course, the most forest ecosystems (e.g. Zha et al., 2010). It is reported that the relationship between $ET/ET_{eq}$ and $G_o$ afford an illustrative way to probe into the interactive effects of soil water availability, leaf area index and photosynthetic capacity on the $ET$ for different vegetation types (Baldocchi et al., 1997; Ryu et al., 2008). Obviously, irrespective of the differences of seasonal and interannual precipitation and $ET$, the $ET/ET_{eq}$ decreased nonlinearly with the decrease of $G_o$ in the present alpine steppe (Fig. 6). This is consistent with other experimental analysis for the relationship between the $ET/ET_{eq}$ and $G_o$ in a variety of ecosystems (e.g. Baldocchi and Xu, 2007; Gu et al., 2008; Krishnan et al., 2012; Lei and Yang, 2010). Also, present $ET/ET_{eq}$ became insensitive to $G_o$ when the latter exceeded 13 mm s$^{-1}$, this threshold is close to the value of theoretical study of McNaughton and Spriggs (1986) which stated that the $G_o$ began to significantly impact $ET/ET_{eq}$ when it was below 16 mm s$^{-1}$.

The general water-limited condition mentioned above suggests the alpine steppe is required to regulate its transpiration through plant functional (e.g. stomata control) and structural property (e.g. leaf area) to survive in this dry environment (Baldocchi et al., 1997). Therefore, despite the $G_o$ calculated by Eq. (9) is not an explicit surrogate for representing physiological parameter.

![Fig. 6. Relation between the midday mean bulk surface conductance ($G_o$) and midday mean $ET/ET_{eq}$ over different periods at the observation site. The gray squares, red triangles and blue circles represent data for the frozen soil period, transition period and rainy period for both observational years, respectively. The heavy black curve $[ET/ET_{eq} = 0.942 \cdot 0.950 \cdot e^{-0.218G_o}]$ denotes the fitted regression for all data over the two observational years. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)](image-url)
stomatal conductance, due to the potential influences of soil evaporation and/or canopy turbulence in sparsely vegetated areas (Alves et al., 1998; Kelliher et al., 1995), Wilson and Baldocchi (2000) hypothesized at least two physiological factors are able to affect the \( G_s \) in growing season: (i) the change of leaf-level stomatal conductance and (ii) the change of leaf area. The former can impact the \( G_s \) through the change of atmospheric humidity deficit, soil moisture and photosynthetic capacity. In the present study, the \( G_s \) decreased with the increase of VPD during the rainy period when plant growth rates are highest (Fig. 7). This is similar to other semi-arid grasslands (e.g. Aires et al., 2008; Krishnan et al., 2012; Ryu et al., 2008; Zhang et al., 2007) and forests (e.g. Granier and Bréda, 1996; Zhu et al., 2014). Although this \( G_s \)-VPD relationship may involve the effect of the drying of topsoil that decrease the soil evaporation, we speculated the decrease of leaf-level stomatal conductance of alpine steppe also played an important role in the decrease of \( G_s \) because vascular plants would close their stomata when exposed to the very high atmospheric humidity deficit (Baldocchi et al., 1997; Igarashi et al., 2015b; Kellner, 2001). Undoubtedly, during the frozen soil and transition periods, the \( G_s \) were not sensitive to the variations of VPD because of the low soil water availability.

Also, the effects of leaf area change on ET have been widely reported around the world (e.g. Li et al., 2005; Suyker and Verma, 2008; Zha et al., 2010; Igarashi et al., 2015a). However, it is not easy to obtain the periodical LAI at observation site because of the inclement environment. Fortunately, the NDVI values exhibit a strong linear relation with LAI when LAI is <2 (this is true for present alpine steppe) (Wang et al., 2005), and the former could be easily derived from the MODIS NDVI product. Fig. 8 shows the relation between 16-day averages of midday mean \( G_s \) values as correspondent with NDVI values in different periods. Overall, the \( G_s \) increased with the increase of NDVI values for the rainy period, indicating that plant growth also impact the \( G_s \) considerably.

Note that this positive relation exists between \( G_s \) and NDVI only validates under well soil water conditions (Chen et al., 2009; Li et al., 2005; Wever et al., 2002). Previous studies have shown that if the soil moisture gets larger than a specific-threshold (depends on soil and plant properties), ET would become not much sensitive to it, while the reverse is seen when soil moisture is lower than such threshold (Zha et al., 2010; Zhang et al., 2005, 2007). Krishnan et al. (2012) found that when the soil moisture declined to <0.08 m\(^{-3}\) m\(^{-3}\) at two grasslands of Arizona, the reduced \( G_s \) as a response water stress and higher VPD could cause obvious decrease of ET. That is, when the desired water conditions are not met, soil moisture would affect the \( G_s \) more significantly (through decreasing both stomatal conductance and soil evaporation) than the leaf area (Igarashi et al., 2015b). This explains why the data of rainy period in Fig. 8 scattered to a little extent, e.g. the 3.57 mm s\(^{-1}\) of \( G_s \) value and the corresponding 0.19 of NDVI value belong to the 16-day period from July 28 to August 12, 2013 with little rainfall.

In fact, the controlling of biophysical factors on the ET via the changes in stomatal conductance and/or leaf area mentioned above is also reflected in the partitioning of ET into soil evaporation and plant transpiration (Schlesinger and Jasechko, 2014; Wang et al., 2014). This is because the transpiration process has a strong link with the vegetation physiological activities. Recent global synthesis showed that not only the LAI but also the plant growth stage could greatly impact the transpiration, thereby its proportion in ET (Wang et al., 2014). Although the biomass of alpine steppe is limited to some extent due to its relatively short growing season length and dry condition when compared with the forest and/or cropland in low-altitude regions, the stable isotope mass balance analysis by Jasechko et al. (2013) suggested that transpiration in TP did account for nearly 80% of the ET. While this proportion has been debated subsequently (Coenders-Gerrits et al., 2014; Good et al., 2015; Wang et al., 2014), the role of biophysical factors play in controlling the ET in alpine steppe ecosystem should be highlighted. Also, more knowledge is urgently needed to elucidate how other biophysical roles (e.g. root water uptake and under-canopy turbulence processes) impact the ET over TP.

### 4.4. Relationship between ET and precipitation

The cumulative rainfall that observed over the chosen area of alpine steppe were 352.5 mm and 322.7 mm in the first and second
year, respectively; while the cumulative ET (plus sublimation) were 362.9 mm and 353.4 mm in the first and second year, respectively (Table 1). As the observation site is generally flat and precipitation would therefore seldom generate runoff, this deficit may be caused by the lack of snow observations. In order to deduce quantitatively the proportion of annual precipitation accounted for by snow, we analyzed the precipitation component over the same period for the CMA station nearest our observation site, at Bange, where both snow and rain observations had been made. The results show that 8.9% and 8.4% of the annual precipitation at Bange occurred during the frozen soil periods in the first and second year, respectively. Considering the daily air temperature in frozen soil period were obviously below freezing temperature (Fig. 2), we estimated the snow were at least ca. 34.4 mm [=352.5 × [8.9% (1–8.9%)]] and 29.6 mm [=322.7 × [8.4%/ (1–8.4%)]] at our observation site in these two years, respectively. Adding the estimated snow to the rain we measured, the annual precipitation over the two years became 386.9 mm and 352.3 mm, respectively. Analysis of the abovementioned annual water budget of total precipitation and total ET (plus sublimation) therefore clarifies that nearly all precipitation in the alpine steppe returned to the atmosphere through land surface ET and sublimation.

Generally, the deficit between precipitation and ET over land surface (with little anthropogenic activity) is reflected in temporal changes snow or soil moisture, while excess rainfall runs off and thus supports streams and/or rivers. Catchment-scale investigations in the non-humid basins of Northern China (Yang et al., 2006) as well as the Qiangtang and Qaidam basins of Tibetan Plateau (Li et al., 2014a,b) have demonstrated the annual ET variability was primarily determined by precipitation. Plot-scale observations in the semi-arid grasslands of Mongolia (Zhang et al., 2005, 2007), Southern Canada (Weyer et al., 2002; Zha et al., 2010) and central New Mexico (Kurc and Small, 2004) also showed that the ET nearly balanced with rainfall on annual scale. These suggest that, in arid and semi-arid area with little ground-water influence, precipitation is the dominant factor that drives the annual ET variability. Therefore, it is not surprising to witness such almost equality between ET (plus sublimation) and precipitation over the present alpine steppe. It should be noted that this phenomenon may be not only true in dry regions. For example, Igarashi et al. (2015a) recently found that even in the tropical teak forest the precipitation levels and length of rain season all play key roles in regulating annual ET variability through their effects on the vapor pressure deficit and plant physiological process.

Interestingly, however, it appears that the total sublimation values are slightly less than the estimated snow in winter in the present study. This maybe because a proportion of the snow melted and stored in the soil, and supplied the ET over the transition period, although no much rain and snow fell in transition period. This was similar to the phenomena in the temperate grasslands of Arizona (Krishnan et al., 2012) and the Mediterranean grasslands of California (Ryu et al., 2008). With the onset of the rainy period, the rainfall substantially increased and this made the cumulative rainfall more or less equal to the cumulative ET in July or August (Fig. 9).

5. Conclusions

Present research investigates two successive years of ET using in-situ observation from a semi-arid alpine steppe at an altitude of 4947 m asl in the TP. Annual ET (plus sublimation) were close to the annual precipitation. On a seasonal scale, ET were much higher in the rainy period (accounting for >70% of annual ET plus sublimation), while were much lower in the transition and frozen soil periods. Although direct soil moisture observations were unattainable, similar patterns in seasonal variations of surface parameters (Gc, Ω and ET/ET0) indicate that ET was generally controlled by soil water availability, which in turn determined by precipitation in the present high-altitude alpine steppe. This is especially true for ET in the frozen soil and transition periods, which experienced extremely low bulk surface conductance. However, multiple rain events enhanced soil water availability in the rainy period. This led to a closer coupling of ET to available energy, thereby witnessing the switch to temporally energy-limited in some days of rainy period.

Thus, the implications from our ET observation is twofold: (i) seasonally, in addition to the soil moisture, both available energy and plant physiological control should be elaborately taken into account when we simulate the monsoon season land surface processes in alpine steppe, although this ecosystem is characterized as semi-arid; (ii) annually, while both stomatal closure and plant structural properties (e.g. leaf area) can regulate the ET to some extent through controlling transpiration process, the commonly water-limited condition makes the alpine steppe ecosystem is quite vulnerable to the possible drought in future because of its shallow root systems and lower stomatal control.

Last but not least, although the present in-situ observation of ET could provide ground truth data for modeling the hydrological and biophysical processes in the data scarce region of TP, more comprehensive observations, e.g. eddy covariance technique, detailed soil water status and leaf stomatal conductance measurement, should be utilized in future to provide comprehensive long-term measurements of ET in alpine steppe.

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