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# An Assessment of Storage Terms in the Surface Energy Balance of a Subalpine Meadow in Northwest China

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**ABSTRACT**

The heat storage terms in the soil–vegetation–atmosphere system may play an important role in the surface energy budget. In this paper, we evaluate the heat storage terms of a subalpine meadow based on a field experiment conducted in the complex terrain of the eastern Qilian Mountains of Northwest China and their impact on the closure of the surface energy balance under such non-ideal conditions. During the night, the average sum of the storage terms was $-5.5$ W m$^{-2}$, which corresponded to 10.4% of net radiation. The sum of the terms became positive at 0730 LST and negative again at about 1500 LST, with a maximum value of 19 W m$^{-2}$ observed at approximately 0830 LST. During the day, the average of the sum of the storage terms was 6.5 W m$^{-2}$, which corresponded to 4.0% of net radiation. According to the slopes obtained when linear regression of the net radiation and partitioned fluxes was forced through the origin, there is an imbalance of 14.0% in the subalpine meadow when the storage terms are not considered in the surface energy balance. This imbalance was improved by 3.4% by calculating the sum of the storage terms. The soil heat storage flux gave the highest contribution (1.59%), while the vegetation enthalpy change and the rest of the storage terms were responsible for improvements of 1.04% and 0.77%, respectively.

**Key words**: subalpine meadow, complex terrain, surface energy balance, storage terms

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## 1. Introduction

A good comprehension of the mass and energy exchange between the Earth's surface and the atmosphere is fundamental for improving regional weather and global climate models (Cava et al., 2008). Over an ideal (non-vegetated) horizontal surface, the energy flux leaving the surface should be equivalent to the energy flux received at the surface; therefore, the closure of the surface energy balance has historically been accepted as a fundamental validation of data quality, particularly for validation of eddy-covariance data, with important implications for estimates of $CO_2$ flux (Kanemasu et al., 1992; Laubach et al., 1994; Twine et al., 2000; Wilson et al., 2002; Cava et al., 2008). However, many experimental studies have indicated that the available energy (net radiation minus soil heat flux) was generally larger than the sum of the vertical turbulent heat fluxes (sensible heat plus latent heat flux). Indeed, the ratio of heat fluxes to available energy, which is often called the closure ratio, ranges from 70%–90% for various ecosystems (Tsvang et al., 1991; Kanemasu et al., 1992; Aubinet et al., 2000; Wilson et al., 2002; Kanda et al., 2004; Ma et al., 2005; Foken et al., 2006; Gao et al., 2007; Cava et al., 2008; Jacobs et al., 2008).

Different hypotheses have been proposed in the literature to explain this observed systematic disparity in the surface energy balance. These hypotheses have mainly implicated instrumental errors, especially those

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related to eddy covariance techniques, including their data processing procedures (Foken and Wichura, 1996; Mauder et al., 2007; Wang et al., 2009a) and their "inabilities" in point observation of the eddy covariance system in the observation of larger eddy contributions (Steinfield et al., 2007; Mauder et al., 2008). Errors from the measurement of net radiation, especially in complex terrain (Kohsiek et al., 2007; Hiller et al., 2008), and specifically the errors in estimates of energy storage in the soil–vegetation–atmosphere system, have been investigated (Bolle et al., 1993; Braud et al., 1993; Oncley et al., 2007), as have the errors from horizontal flow convergence/divergence or a nonzero mean vertical velocity (Lee, 1988; Oncley et al., 2007). It has been found that the occurrence of an energy imbalance is, generally speaking, strongly dependent on the characteristics of the measurement site, and is more pronounced in complex terrain (Rotach et al., 2003; Cava et al., 2008).

In a study by Meyers and Hollinger (2004), the storage terms in an agricultural ecosystem in central Illinois were found to generally comprise a small fraction (<5%) of the net radiation when considered separately. However, the combination of soil and canopy heat storage and the energy stored in the carbohydrate bonds produced by photosynthesis were shown to comprise roughly 15% of the total net radiation for maize and 7% for soybean during the morning hours from 0600 to 1200 LST when the canopy was fully developed.

In a central Amazonian rainforest ecosystem in Brazil, excluding the total aboveground thermal energy storage rate from the calculation of the surface energy balance increased the average residual by approximately 25% and 20% for the dry and wet groups of days, respectively. In addition, the air, trunks and other biomass components contributed 35%, 40%, and 25% to the total daily energy storage rates of the forest, respectively (Michiles and Gielow, 2008).

In a grassland ecosystem in the Netherlands, the soil heat flux estimated using a harmonic analysis technique improved the energy budget closure by 9% and the photosynthesis budget by 3%. When the rest of the storage terms were taken into account, an additional 3% in the energy closure was achieved, with the enthalpy storage in the air layer between the EC system and the ground surface appearing to give the most important contribution (2%) and the other storage terms accounting for 1% of the improvement (0.5% from atmospheric moisture storage, 0.5% from crop enthalpy storage, and 0.1% from canopy dew water enthalpy storage). By considering these storage terms, a closure of 96% was obtained (Jacobs et al., 2008).

The present study was conducted to evaluate the heat storage terms of a subalpine meadow in the complex terrain of the eastern Qilian Mountains in Northwest China and their impact on closure of the surface energy balance under such non-ideal conditions.

## 2. Field experiment

### 2.1 *Site description*

The field experiment was conducted at the Wushaoling meteorological observation station of the China Meteorological Administration (Fig. 1), located in the eastern Qilian Mountains in Northwest China (37°12′N, 102°52′E; elevation: 3045 m). The eastern Qilian Mountains are located in the transitional zone of the Tibetan Plateau, the Loess Plateau, and the arid region of Northwest China, which is one of the key areas subject to global environmental change (Lu et al., 2006). Due to the climate and complex terrain, the prevalent vegetation types in the area are mountainous pastures and forests. The vegetation density in the area varies with terrain, soil, water, and climate factors (Zhou et al., 2007).

The observation site was a small mesa approximately 30 m in length and 20 m in width, surrounded by higher slopes in the east and west and lower valleys in the north and south. The underlying surface of the observation site was flat and covered with subalpine meadow. During the field experiment (23 July to 17 August 2006), the grass cover was allowed to grow naturally and reached heights of 0.36–0.42 m (mean height = 0.4 m) and a green grass yield of 2.7–3.2 kg m$^{-2}$ (mean yield = 3 kg m$^{-2}$). The soil in the study area is subalpine meadow soil (Yang, 2002).

The area has a temperate continental mountainous climate. During winter, atmospheric circulation is

**Fig. 1.** Photograph of the observation site.

controlled by the Mongolia anticyclone, which results in cold and dry conditions and little precipitation. During summer, the atmospheric circulation is controlled by the continental cyclone, which results in warm and wet conditions with a dramatic diurnal difference in temperature. The difference in precipitation between summer and winter is large, with the majority of the precipitation occurring during summer. The average annual rainfall at the observation site is approximately 406 mm and the average annual temperature is $-0.1°$C.

### 2.2 *Instrumentation*

The incoming ($R_{\rm gi}$) and outgoing ($R_{\rm go}$) shortwave radiation fluxes were measured using two aspirated pyranometers (Kipp & Zonen, model CM3) at a height of 1.5 m. At the same height, the incoming ($R_{\rm li}$) and outgoing ($R_{\rm lo}$) longwave radiation fluxes were measured using two longwave radiometers (Kipp & Zonen, model CG3). The outgoing longwave radiation was used to evaluate the grass cover surface temperature $T_{\rm s}$:

$$T_{\rm s}=\left(\frac{R_{\rm lo}}{\varepsilon\sigma}\right)^{1/4}, \tag{1}$$

where $\varepsilon$ is the emissivity of the grass cover [$\varepsilon$=0.99 according to Jacobs et al. (2008) and Wang et al. (2009b)] and $\sigma$ is Stefan Boltzmann's constant ($\sigma$ = 5.67× $10^{-8}$ W $K^{-4}$ $m^{-2}$). The net radiation at a height of 1.5 m, $R_{\rm nz}$, was determined by summing the total radiation budget:

$$R_{\rm nz}=R_{\rm gi}+R_{\rm go}+R_{\rm li}+R_{\rm lo}\,. \tag{2}$$

In our studies, the net radiation could also be demonstrated using a net radiometer (Kipp & Zonen, model CNR1).

The soil temperature was measured at depths of 0, 10, 20, and 30 cm using temperature probes (Campbell Scientific Inc., model 109). The soil moisture content was measured at depths of 10, 20, and 30 cm using a water content reflectometer-Time Domain Reflectometry (TDR) system (Campbell Scientific Inc., probe type CS616). The soil heat flux was measured using three Hukseflux soil heat flux plates (Campbell Scientific Inc., model HFP01) buried at a depth of 2 cm at three sites with circumferences of 1.5 m from the center at an angle of 120° by spatial averaging of the three sensors. During measurement, the aspects of differing thermal properties between each sensor and its environment were dealt with using a higher accuracy self-calibrating type of heat flux sensor according to the manufacturer's specifications.

The water vapour and $CO_2$ fluxes were measured using the eddy-covariance method. The tower was equipped with a three-dimensional sonic anemometer to measure the wind speed, sonic temperature, and wind direction (Campbell Scientific Inc., model CSAT3) and a $CO_2$ and $H_2O$ open-path analyzer (Infrared Gas Analyser, LI-COR Inc., model LI-7500) to measure the $CO_2$ and water vapour concentrations. The instruments were mounted at 1.8 m above ground level. In addition, the tower was equipped with an air temperature and relative humidity probe (Vaisala, Inc., model HMP45-D) enclosed in a radiation shield and a wind direction and speed sensor (R. M. Young Company, model 05106) at heights of 1.5 and 2.5 m above ground level, respectively.

All instruments were installed according to the manufacturer's specifications. For data acquisition, we used CR5000, CR1000, and CR23X-TD dataloggers (Campbell Scientific Inc.) as the storage media. A network of dataloggers was set up and managed by LoggerNet datalogger support software of Campbell Scientific Inc. In addition, LoggerNet supported the programming, communications, and data retrieval between the dataloggers and a computer. The slow response meteorological instruments were sampled at 0.2 Hz and 30-min intervals, after which the data were averaged and stored in the dataloggers for subsequent processing. The fast response instruments were sampled at 10 Hz. The raw data of the eddy covariance system were stored on a computer and processed at 30-min intervals using LoggerNet 2.

## 3. Theory and methodology

The energy balance over an ideal (nonvegetated) horizontal surface can be written as:

$$R_{\rm n}=H+L_{\rm v}E+G_{\rm o}, \tag{3}$$

where $R_{\rm n}$ (W $m^{-2}$) is the net radiation flux, $G_{\rm o}$ (W $m^{-2}$) is the surface soil heat flux, $H$ (W $m^{-2}$) is the sensible heat flux and $L_{\rm v}E$ (W $m^{-2}$) is the latent heat flux at the interface between the Earth's surface and the atmosphere. As these measurements are usually performed at a certain height or depth away from the surface, the energy balance closure test should show an imbalance in favour of net radiation. To illustrate this, a more complete energy balance equation is presented below in Eq. (4).

$$R_{\rm nz}=H_{\rm z}+L_{\rm v}E_{\rm z}+G_{\rm d}+S+X, \tag{4}$$

where $R_{\rm nz}$ is the net radiation measured at height $z$, $H_{\rm z}$ is the sensible heat flux, $L_{\rm v}E_{\rm z}$ is the latent heat flux (both measured at height $z$), $G_{\rm d}$ is the soil heat flux measured at depth $d$, $X$ (W $m^{-2}$) is the residual error associated with the measurement and data calculation processes, and $S$ (W $m^{-2}$) represents the additional storage terms in the soil–vegetation–atmosphere system. $S$ includes the soil heat storage flux above the heat flux plate ($S_{\rm s}$), the air enthalpy change ($S_{\rm a}$), the atmospheric moisture change ($S_{\rm q}$), the canopy dew water enthalpy change ($S_{\rm d}$), the photosynthesis flux ($S_{\rm p}$), and the vegetation enthalpy change ($S_{\rm v}$). $X$ is often neglected at flat and extensive homogeneous locations. The soil heat storage flux can be written as:

$$S_{\rm s}=C_{\rm s}\int_d^0\frac{\partial T_{\rm so}}{\partial t}dz, \tag{5}$$

where $d$ is the depth of the soil heat flux plate, $C_{\rm s}$ (J m$^{-3}$ K$^{-1}$) is the volumetric heat capacity of the soil, and $T_{\rm so}$ is the soil temperature. For practical calculations, Eq. (5) was reduced as follows according to the specifications of the manufacturer (Campbell Scientific Inc.):

$$S_{\rm s}=\frac{\Delta T_{\rm so}\,C_{\rm s}\,d}{\Delta t}, \tag{6}$$

where $C_{\rm s}$=4.2×10$^3\rho_{\rm s}(0.2+\rho_{\rm s}\theta_{\rm v})$ (J m$^{-3}$ K$^{-1}$) when considering the influence of soil water on the soil heat storage term (according to Li et al., 2007), $\rho_{\rm s}$ is the soil density (kg m$^{-3}$), $\theta_{\rm v}$ is the soil volumetric water content, and $\Delta t$ is a time averaging interval. The air enthalpy change can be written as:

$$S_{\rm a}=C_{\rm a}\int_0^h\frac{\partial T_{\rm a}}{\partial t}dz, \tag{7}$$

where $T_{\rm a}$ (K) is the air temperature, $C_{\rm a}$ (J m$^{-3}$ K$^{-1}$) is the volumetric heat capacity for moist air, and $h$ is the fixed height of the eddy covariance system. The atmospheric moisture change can be written as:

$$S_{\rm q}=L_{\rm v}\int_0^h\frac{\partial a}{\partial t}dz, \tag{8}$$

where $L_{\rm v}$ (J kg$^{-1}$) is the latent heat for vaporization, and $a$ (kg m$^{-3}$) is the atmospheric moisture density. During calculation, according to Michiles and Gielow (2008), Eqs. (7) and (8) were reduced to:

$$S_{\rm a}=\frac{\rho_{\rm a}c_p(1+0.84\bar{q})}{\Delta t}\Delta\bar{T}_{\rm a}h; \tag{9}$$

$$S_{\rm q}=\frac{\rho_{\rm a}L_{\rm v}}{\Delta t}\Delta\bar{q}h, \tag{10}$$

Where $\rho_{\rm a}$, $c_p$, and $\bar{T}_{\rm a}$ are the density, the specific heat at constant pressure of the air, and the average temperature of the air volume from 0 to $h$, respectively, and $\bar{q}$ is the average specific humidity of the air volume.

In this study, the canopy dew water enthalpy change was ignored in the surface energy balance equation because the dew water change was not measured. The photosynthetic flux is the change in the Gibbs free energy. A canopy assimilation rate of 1 mg $CO_2$ m$^{-2}$ s$^{-1}$ is equal to an energy flux of 11.2 W m$^{-2}$ (Nobel, 1974; Meyers and Hollinger, 2004), and this conversion factor was used to compute the measured photosynthesis rates from the eddy covariance measurements to an equivalent energy flux. The vegetation enthalpy change can be written as:

$$S_{\rm v}=\frac{\Delta T_{\rm s}M_{\rm gc}C_{\rm gc}}{\Delta t}, \tag{11}$$

where $\Delta T_{\rm s}$ (K), $M_{\rm gc}$ (kg m$^{-2}$), and $C_{\rm gc}$ (J kg$^{-1}$ K$^{-1}$) are the surface temperature change, the mass, and the specific heat capacity of the green grass cover, respectively. In this field experiment, we only measured the green grass mean yield based on the biometrics means (as described in section 2.2) and did not obtain the specific heat capacity of the green grass. Practical calculation enabled $M_{\rm gc}C_{\rm gc}$ to be reduced to:

$$M_{\rm gc}C_{\rm gc}=M_{\rm w}C_{\rm w}+M_{\rm dw}C_{\rm dw}, \tag{12}$$

where $M_{\rm w}$ (kg m$^{-2}$) and $M_{\rm dm}$ (kg m$^{-2}$) are the masses of water and organic dry matter of the green grass, respectively, and $C_{\rm w}$ (J kg$^{-1}$ K$^{-1}$) and $C_{\rm dm}$ (J kg$^{-1}$ K$^{-1}$) are the specific heat capacities of water and organic dry matter, respectively. In this study, the numerical values of masses used (calculated from the green grass mean yield) were $M_{\rm w}$= 2.64 kg m$^{-2}$ and $M_{\rm dm}$= 0.36 kg m$^{-2}$, and the numerical values of the specific heat capacities [according to Atzema (1993) and Jacobs et al. (2008)] were $C_{\rm w}$ = 4190 J kg$^{-1}$ K$^{-1}$ and $C_{\rm dm}$ = 1920 J kg$^{-1}$ K$^{-1}$.

The storage changes were calculated over a measurement averaging time interval of 30 min. The soil temperature and the soil moisture content of a given depth were calculated based on harmonic analysis and the linear interpolation technique [according to Darcy's Law (Darcy, 1856; Philip, 2006)], respectively. Numerical values of some parameters used in this study were: $\rho_{\rm s}$=1030 kg m$^{-3}$ (Yang, 2002), $\rho_{\rm a}$ = 1.2 kg m$^{-3}$, $c_p$=1004 J kg$^{-1}$ °C$^{-1}$, and $L_{\rm v}$ = 2.4×10$^6$ J kg$^{-1}$ (Michiles and Gielow, 2008).

## 4. Results and discussion

Data from the field experiment were calculated and analyzed in detail to assess the storage terms and realize the characteristics of the energy balance closure for the subalpine meadow in the complex terrain of the eastern Qilian Mountains in Northwest China. The main meteorological variables evaluated during the field experiment are shown in Fig. 2.

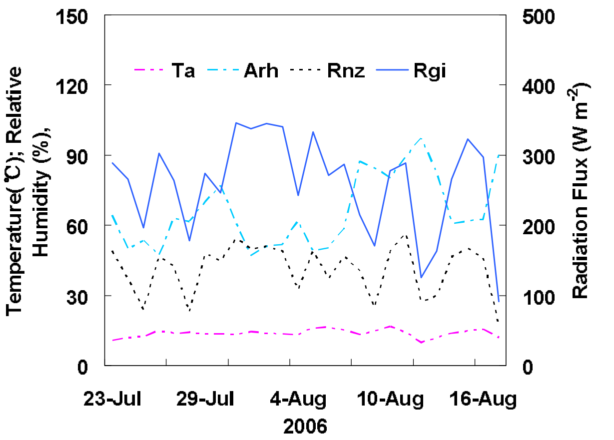


**Fig. 2.** Variation of the daily mean air temperature ($T_{\rm a}$), relative humidity ($A_{\rm rh}$), shortwave incoming radiation ($R_{\rm gi}$) and net radiation ($R_{\rm nz}$) at a height of 1.5 m for the period between 23 July and 17 August, 2006.

The day-averaged variation of the individual storage terms and their sum $S$ ($S_{\rm p}+S_{\rm q}+S_{\rm s}+S_{\rm a}+S_{\rm v}$) in the soil–vegetation–atmosphere system together with the net radiation ($R_{\rm nz}$) during the field experiment is shown in Fig. 3. During the night, the sum of the storage terms, $S$, was $-5.5$ W m$^{-2}$ on average, corresponding to 10.4% of $R_{\rm nz}$. $S$ became positive at 0730 LST. During the morning and early afternoon, between 0730 and 1430 LST, $S$ was positive, with an average of 13.2 W m$^{-2}$, corresponding to 3.5% of $R_{\rm nz}$. The daily maximum value of $S$ was reached at about 0830 LST, with values of up to 19 W m$^{-2}$ observed. Typically, at about 1500 LST, $S$ became negative again, reaching its daily minimum at approximately 2000 LST, with values as low as $-13.0$ W m$^{-2}$ observed. The average sum of storage terms, $S$, was 6.5 W m$^{-2}$, corresponding to 4.0% of $R_{\rm nz}$ during daytime. Additionally, the daily average value of the ratio $S/R_{\rm nz}$ was 7.2%. When compared to the subalpine meadow in this study, the daily variation in the storage terms for rainforest (Michiles and Gielow, 2008) and cropland (Meyers and Hollinger, 2004) was also similar. However, the value of $S$ and the ratio $S/R_{\rm nz}$ was larger in rainforest and cropland than in the subalpine meadow.

Figure 4 is a scattergram of the 30-min averages of net radiation, $R_{\rm nz}$, and the sum of the energy flux densities $H_{\rm z}+L_{\rm v}E_{\rm z}+G_{\rm d}$ for the period between 23 July and 17 August 2006. Here, periods with rain events were discarded. The linear regression forced through the origin was $y$=0.860$x$ with an $r^2$=0.966 ($N$=977), implying that there is an imbalance of 14.0%, which is higher than the results of some studies, but not an uncommon value (Wilson et al., 2002; Meyers and Hollinger, 2004; Gao et al., 2007; Wang et al., 2009a). These findings indicate that the energy budget closure is not ideal for the subalpine meadow in the complex terrain of the eastern Qilian Mountains in Northwest China when the storage terms are not considered.

A scattergram of the 30-min averages of net radiation $R_{\rm nz}$ versus the sum of the energy flux densities $H_{\rm z}+L_{\rm v}E_{\rm z}+G_{\rm d}+S$ for the period between 23 July and 17 August 2006 is presented in Fig. 5. Periods with rain events were discarded from this analysis. The linear regression forced through the origin was $y$=0.894$x$ with $r^2$=0.967 ($N$=977), implying that there is an imbalance of 10.6%. Based on these findings, we con-

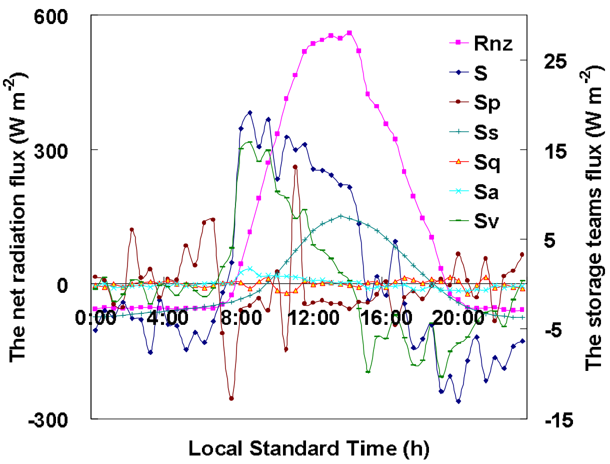


**Fig. 3.** Day-averaged variation in net radiation ($R_{\rm nz}$) and the additional storage terms ($S$ without calculating the canopy dew water enthalpy change) during the same periods as in Fig. 2. Periods with rain events were discarded.

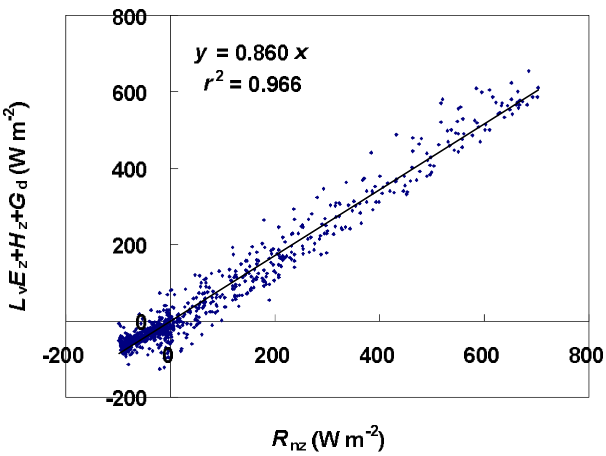


**Fig. 4.** Scattergram of the 30-min averages of the net radiation ($R_{\rm nz}$) and the sum of the energy flux densities $H_{\rm z}+L_{\rm v}E_{\rm z}+G_{\rm d}$ during the same periods as in Fig. 2.

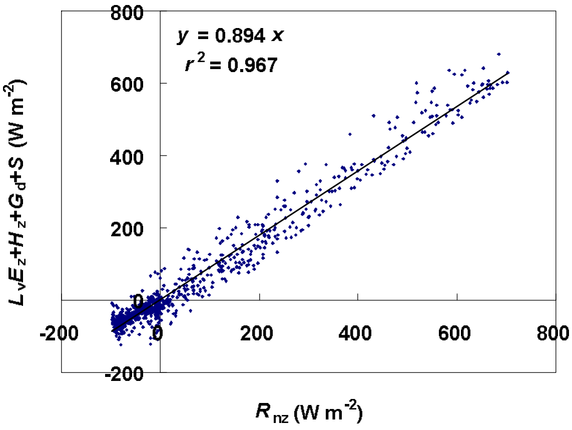


**Fig. 5.** Scattergram of the 30-min averages of the net radiation ($R_{\rm nz}$) and the sum of the energy flux densities $H_{\rm z} + L_{\rm v}E_{\rm z} + G_{\rm d} + S$ during the same periods as in Fig. 2. $S = S_{\rm p} + S_{\rm s} + S_{\rm q} + S_{\rm a} + S_{\rm v}$.

clude that considering the storage terms improves the imbalance by 3.4%. This improvement of 3.4% is lower than the results of some studies (Meyers and Hollinger, 2004; Jacobs et al., 2008; Michiles and Gielow, 2008), but it also represents a significant improvement in the surface energy budget.

To obtain better insight into the importance of the various storage terms, the percentage improvement in the energy budget in response to various storages is presented in Table1. The results indicate that the soil heat storage term, $S_{\rm s}$, gives the highest contribution to this improvement (approximately 1.59%), followed by the vegetation enthalpy storage term, $S_{\rm v}$ (approximately 1.04%). The rest of the storage terms, $S_{\rm p} + S_{\rm q} + S_{\rm a}$, led to an improvement of approximately 0.77%.

When compared with low altitude grassland (Jacobs et al., 2008), the total contribution of the storage terms to the surface energy closure in the subalpine meadow evaluated here was smaller (Table 1). However, the contribution of the vegetation enthalpy storage term in the subalpine meadow is bigger than in the low altitude grassland. This relates to differences in mass of the green grass cover in the two areas. The mean yield of green grass is 3 kg m$^{-2}$ in the subalpine meadow but only 1.7 kg m$^{-2}$ in the low altitude grassland due to the grass cover being mowed weekly (Jacobs et al., 2008). The contribution of other storage terms ($S_{\rm p}$, $S_{\rm q}$ and $S_{\rm a}$) in the subalpine meadow is smaller than in the low altitude grassland.

Contrasting the importance of the various storage terms to the surface energy balance in the two grasslands, the soil heat storage term plays the most important role in the subalpine meadow. The second is the vegetation enthalpy change and the third is the photosynthesis flux. In the low altitude grassland, however, the most important role is that played by photosynthesis flux, the second is the air enthalpy change, and the third is the atmospheric moisture and vegetation enthalpy change. By all accounts, the importance of the various storage terms to the surface energy closure is different in the two grasslands, even if the soil heat storage term is not considered. This indicates that the contribution of the various storage terms to the energy closure in middle and high altitude grassland needs more in-depth study.

Compared with agricultural (Meyers and Hollinger, 2004) and forest ecosystems (Michiles and Gielow, 2008), the total contribution of the storage terms to the surface energy closure in the grassland ecosystem was smaller. In particular, the total contribution of aboveground thermal energy storage is obviously greater in agricultural and forest ecosystems. The contribution of the soil heat storage term in different ecosystems also requires more in-depth study.

## 5. Conclusions

In this paper we have evaluated the heat storage terms of a subalpine meadow in the complex terrain of the eastern Qilian Mountains in Northwest China and their impact on the closure of the surface energy balance under such non-ideal conditions based on a field experiment. The following main conclusions can be drawn from the study:

(1) During the night, the average sum of the storage terms was $-5.5$ W m$^{-2}$, corresponding to 10.4%

**Table 1.** Comparison of the contribution of the storage terms to the surface energy closure in two grasslands.

| | Storage term | | | | | | Elevation |
|---|---|---|---|---|---|---|---|
| | $S_{\rm p}$ | $S_{\rm s}$ | $S_{\rm q}$ | $S_{\rm a}$ | $S_{\rm v}$ | $S_{\rm d}$ | |
| Subalpine meadow | 0.60% | 1.59% | 0.03% | 0.14% | 1.04% | — | 3045 m |
| Grassland | 3.0% | — | 0.5% | 2.0% | 0.5% | 0.1% | 7 m |

Note: $S_{\rm p}$ = photosynthesis flux, $S_{\rm s}$ = soil heat storage flux,$S_{\rm q}$ = atmospheric moisture change, $S_{\rm a}$ = air enthalpy change, $S_{\rm v}$=the vegetation enthalpy change, $S_{\rm d}$=the canopy dew water enthalpy change. Grassland data was from Jacobs et al. (2008).

of the $R_{\mathrm{nz}}$. This value became positive at 0730 LST and negative again at around 1500 LST, with a maximum value of 19 W m$^{-2}$ observed at about 0830 LST. During daytime, the average sum of the storage terms was 6.5 W m$^{-2}$, corresponding to 4.0% of $R_{\mathrm{nz}}$.

(2) There is an imbalance of 14.0% in the subalpine meadow when the storage terms are not considered in the surface energy balance.

(3) The imbalance of the energy budget for the subalpine meadow is improved by 3.4% when calculating the sum of the storage terms. $S_{\mathrm{s}}$ give the highest contribution (1.59%), while $S_{\mathrm{v}}$ and $S_{\mathrm{p}}+S_{\mathrm{q}}+S_{\mathrm{a}}$ result in improvements of 1.04% and 0.77%, respectively.

(4) The importance of the various storage terms to the surface energy closure is different in low and high altitude grasslands.

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