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ORIGINAL ARTICLE

Quantifying the parameters that control turbulent land–atmosphere energy exchange over the Dunhuang Gobi land surface, northwest China

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Abstract Land–atmosphere interactions in the northwest arid and semi-arid areas of China are important because of their influence on energy exchange at both regional and global scales. In this paper, regular meteorological observations, as well as flux eddy covariance methods, are used to measure land–atmosphere energy exchange. Experiments were conducted over the Dunhuang Gobi, a typical arid region of northwest China, during the summer of 2008. Surface fluxes were measured directly using an eddy-correlation system, while the vertical gradients of wind and temperature were also observed at four levels using a meteorological tower. Based on the observation data, the integral mean of surface albedo over the Gobi was found to be 0.238. In addition, the mean values of momentum, heat roughness lengths, and soil thermal conductivity were, 0.045, 0.005 cm, 0.199 and 0.201 W $(m\ K)^{-1}$, respectively. The Monin–Obukhov similarity functions for momentum and heat were analyzed and new empirical formulae were developed.

Keywords Albedo · Roughness · Similarity functions · Gobi

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Introduction

Observations of the fluxes associated with different surface conditions are important to provide reliable data for land surface modeling. In most of the present numerical atmospheric models, the handling of lower boundary physical processes, such as surface momentum and heat fluxes, is achieved via a land surface model that is coupled with the atmospheric numerical model. Dickinson (1995) pointed out that improving land surface process models would improve the simulation and prediction abilities of the atmospheric models. The key to improving the land surface models is a reasonable parameter scheme.

Desert and Gobi are the main land surfaces in the arid region of northwest China (Zhang and Wei 2004). As result of its unique albedo, roughness, and surface energy exchange, which differ to those of other surfaces, Gobi has a significant influence on local, regional, and even global climate. An understanding of Gobi–air interactions, as well as the environmental controls on surface energy exchanges, is therefore necessary if the influence of such interactions on climate and weather is quantified (Liu et al. 2009), and the response of Gobi to climate variability is anticipated. Consequently, the accurate description of land surface processes over desert and Gobi in the arid region of northwest China is becoming increasingly important. Furthermore, the study of land surface process parameters in arid northwest China will improve understanding of the hydrological and energy cycles in this region, and can aid in the economic development of the region. Gobi that is close to the edge of an oasis has its own special land surface processes that protect and maintain the oasis. Study in these areas will help to increase knowledge of the physical processes over Gobi land surface.

The purpose of this study was to collect the data required to investigate the parameter schemes associated with land surface processes on the Gobi surface. These data will be used to improve and verify the parameter schemes. The objective was to identify the most reasonable parameter schemes and to better understand the interaction between the lower atmosphere and the Gobi surface. In this paper, the calculation and verification procedures were presented.

Data and methodology

The observations over the Gobi surface were conducted by the Cold and Arid Regions Environmental and Engineering Research Institute at the Chinese Academy of Science, China. The data analyzed in this paper were gathered at the Shuangdunzi Gobi micrometeorology station between 7 and 28 August 2008. The station is located at 40°10′N, 94°31′E, and has a bare soil surface typical of this arid region. The site is located about 5 km from the edge of the oasis and at an elevation of 1150 m above sea level; it has a mean annual air temperature of 10.8 °C and mean annual precipitation of 40 mm (Zhang et al. 2005). The area is flat, with the uppermost soil layer being mainly pebbles, and this is underlain by sand (Wang et al. 2010). The land surface is rarely dry and soil moisture is roughly equal to 0 during the observation period. Vegetation cover at this station is 0; soil texture contains sand 82 %, clay 6 %, silt 12 %; and its color is a little yellow.

Observations made at the micrometeorological station include wind speed and direction, air temperature, and humidity (from a tower with four levels at 1, 2, 8, and 18 m above the surface); four radiation components, surface and soil temperatures, humidity, and heat fluxes at various depths below the ground surface. A CSAT3 3D sonic anemometer and KH20 temperature and relative humidity probe were used to calculate the latent heat (LE) and sensible heat (HE) fluxes with installing at a height of 3 m above the surface. The radiometer was mounted at 1.5 m above the surface. The instruments' detail is shown in Table 1.

In addition to the regular meteorological observations, surface fluxes were recorded during 30 min, and turbulence measurements were made using a 3D sonic anemometer and a krypton hygrometer at a height of 3 m; i.e., the eddy-correlation system (Moore 1986). Fluctuations in vertical wind, air temperature, and vapor were recorded by the eddy-correlation system at a sampling rate of 10 Hz. For a standard period (e.g., 30 min), the sensible or latent heat flux can be calculated using the following equation:

$$Q_s = \rho \frac{1}{T} \int sw\mathrm{d}T \approx \rho \frac{1}{N} \sum s'w' = \rho \overline{s'w'} \tag{1}$$

The eddy-correlation system is used to measure fluctuations in certain parameters and to obtain their flux by correlation with each other.

$$\begin{aligned} u_* &= \left[\overline{u'w'}^2 + \overline{v'w'}^2 \right]^{1/4} \\ T_* &= \overline{t'w'}/u_* \\ H &= C_p \rho \overline{w't'} \end{aligned} \tag{2}$$

where u_* and T_* are the friction velocity and turbulent temperature scale, respectively; u', v', w', and t' are fluctuations in the wind components and air temperature at the reference height; H is the sensible heat; C_p is the specific heat at constant pressure; and ρ is the air density. The temperature was corrected for the effect of humidity on sonic-measured temperatures, but temperature was not converted to potential temperature because the instruments were located very close to each other and to the surface.

The data logger saved two types of data at the same time (Dunhuang local time). The observation system was powered by five solar panels and a battery. This analysis considers data obtained during 7–28 August 2008. All data, both regular observations and eddy-correlation data, used to calculate the parameter schemes were processed by removing outliers. Furthermore, WPL correction (Webb et al. 1980) was also carried out on the eddy data. These data were used to determine the most appropriate parameter scheme to use for the Dunhuang Gobi surface.

Table 1 Specification and performance of the sensors

Observation items	Instrument (manufacturer)	Height\depths (cm)
3D wind speed	CSAT3 (CAMPBELL)	300
Air humidity	KH20 (CAMPBELL)	300
Radiation components	CNR-1 (Kipp and Zonen)	150
Wind speed gradient	WindSonic (GIll)	1\2\8\18 (m)
Air temperature gradient	PT100 (VAISALA)	1\2\8\18 (m)
Soil temperature gradient	107-L (CAMPBELL)	0\5\10\20\40\80\180
Data logger	CR5000 (CAMPBELL)	100

Results and discussion

Albedo

Land surface albedo is defined as the ratio of reflected radiation from the surface to incident radiation upon the surface. It is generally calculated from the observed radiation component as follows:

$$\alpha = S_u/S_d \quad (3)$$

where S_u is the upward shortwave as surface-reflected radiation, while S_d is the downward shortwave as solar radiation. The maximum albedo was calculated to be 0.56 and the minimum 0.22; the diurnal variation follows a 'U'-shaped profile, and is larger in the morning and evening than in the afternoon.

The higher albedo at both sunrise and sunset is due to observation errors when the values of S_u and S_d are low. It seems unlikely. That mean albedo calculated by the arithmetic mean method would cause the large observational errors. Therefore, the albedo was calculated by integral method; i.e., the ratio of the areas surrounded by the downward and upward shortwave band curves and the X axis such as in Fig. 1.

$$\alpha = \frac{\int S_u \mathrm{d}t}{\int S_d \mathrm{d}t} \quad (4)$$

Figure 1 shows the distribution of the mean diurnal variation of S_u, S_d, and α over the Dunhuang Gobi. These profiles are typical of this region: increasing and decreasing at sunrise and sunset, respectively, and with a maximum at noon. Using Eq. (4), the mean albedo was calculated to be 0.238. Fitted equation between albedo and local time was established as $\alpha = 0.003t^2 - 0.067t + 0.604$.

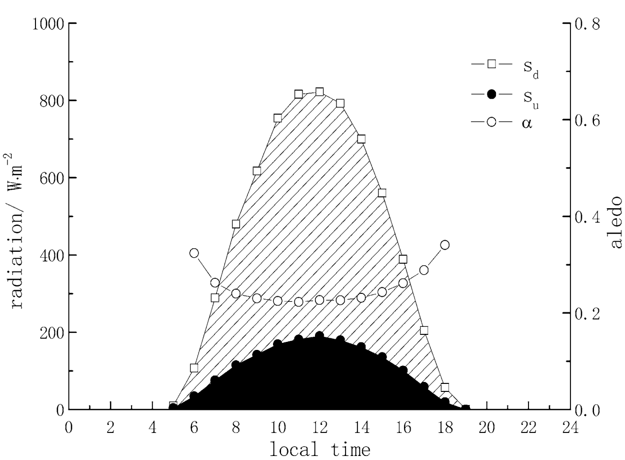

Fig. 1 Mean diurnal variation of *upward* and *downward* shortwave radiation. S_d downward shortwave, S_u upward shortwave, α albedo

Roughness length

The momentum roughness length z_{0m} is a parameter in the vertical wind profile equations used to model the horizontal mean wind speed near the ground. In the log wind profile, it is equivalent to the height at which the wind speed is zero. Based on the Monin–Obukhov (M–O) similarity theory (Monin and Obukhov 1959), the roughness length z_{0m} and z_{0h} can be calculated as follows:

$$u = \frac{u_*}{k}\left[\ln((z-d)/z_{0m}) - \psi_m(\zeta)\right] \quad (5)$$

$$T - T_s = \frac{T_*}{k}\left[\ln((z-d)/z_{0h}) - \psi_h(\zeta)\right] \quad (6)$$

where u and T are wind velocity and temperature, respectively, at the observation height z ($z = 3$ m); T_s is the land surface temperature; d is the displacement height (for Gobi $d = 0$); z_{0m} and z_{0h} are the roughness lengths of momentum and temperature, respectively; and ψ_m and ψ_h are the integral forms of momentum and the temperature profile function, respectively. $\zeta = (z - d)/L$ is the M–O stability parameter, and L is the Obukhov length, where

$$L = -\frac{u_*^3 T}{\kappa g \overline{w'T'}} \quad (7)$$

here, k is the von Karman constant and equals 0.4, and g is the gravitational constant. When $\zeta > 0$, the atmospheric stratification is stable, when $\zeta < 0$ it is unstable, and when $\zeta = 0$ it is neutral. To calculate the roughness length as accurately as possible, only those data with a neutral status ($-0.05 \le \zeta \le 0.05$) were used, where $\psi_m = \psi_h = 0$. The value of z_{0m} is so sensitive in calculation. Therefore, the least squares fitting method is presented. By mathematical manipulations the following formulae can be obtained, the slope α is what is needed (Fig. 2):

$$\begin{aligned} y &= u\kappa \\ x &= u_* \\ \alpha &= \ln\frac{z}{z_{0m}} \end{aligned} \quad (8)$$

The calculation of roughness length can be converted to the slope of the function of a single variable. This improves the result and removes the influence of anomalous values during calculating. The least squares fit of the roughness length computed from the data under neutral status was 0.045 cm. This is on the same order of magnitude as that found by previous research over the Gobi land surface (He et al. 2009; Hu et al. 1990).

So far, there have been few observations of, or research into, the temperature roughness lengths. The same method was used to change the format of Eq. (6), and calculated the temperature roughness length to be 0.005 cm. The

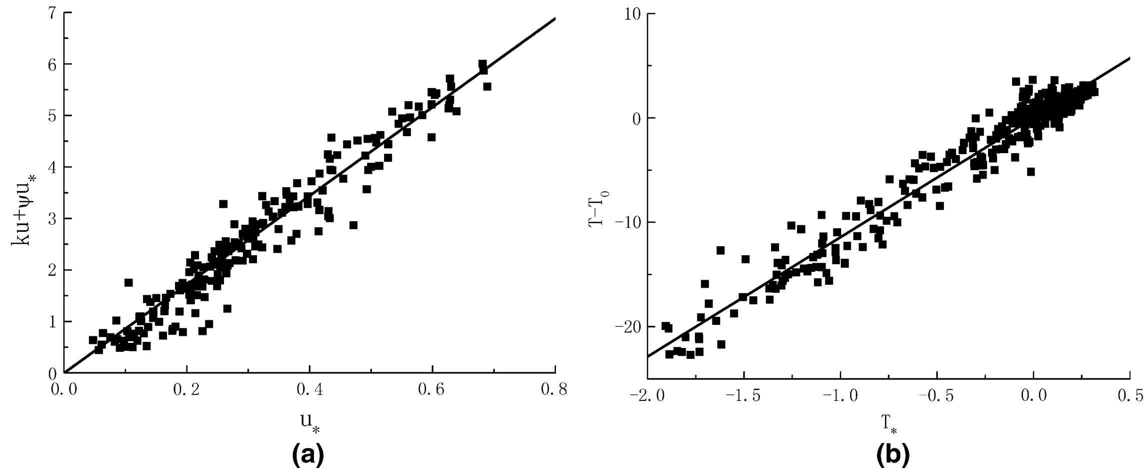

Fig. 2 Results of the least squares method used to calculate roughness lengths. **a** Momentum, **b** temperature

mean values of momentum and heat roughness lengths were also determined.

Similarity profile function

In most land surface process models, surface momentum and sensible heat fluxes are calculated using the flux-profile relationship based on the M–O similarity theory, which can be expressed as

$$\psi_{\mathrm{m}}(\zeta) = \int_{\zeta_0}^{\zeta} [1 - \varphi_{\mathrm{m}}(\zeta)]\mathrm{d}\ln\zeta \tag{9}$$

$$\psi_{\mathrm{h}}(\zeta) = \int_{\zeta_0}^{\zeta} [1 - \varphi_{\mathrm{h}}(\zeta)]\mathrm{d}\ln\zeta \tag{10}$$

where ϕ_{m} and ϕ_{h} are the forms of momentum and temperature profile functions, respectively, and can be calculated using observational data and the following equations:

$$\phi_{\mathrm{m}}(\zeta) = \frac{\mathrm{d}\bar{u}}{\mathrm{d}z}\frac{\kappa z}{u_*} \tag{11}$$

$$\phi_{\mathrm{h}}(\zeta) = \frac{\mathrm{d}T}{\mathrm{d}z}\frac{\kappa z}{T_*} \tag{12}$$

The empirical formulae of the similarity functions ϕ_{m} and ϕ_{h} were first obtained by Businger et al. (1971) using data from the Kansas experiment. Since then, there has been much research into similarity functions, and the popularly accepted forms of the similarity functions have been developed (Dyer 1974; Hoegstroem 1988; Stull 1988). The Businger–Dyer empirical formulae are the most widely accepted, and are used in many numerical models. However, are they suitable for use in extremely arid regions with a Gobi land surface?

To determine the similarity functions ϕ_{m} and ϕ_{h}, the differential equations are used by the gradient of wind speed, potential temperature, and roughness lengths given before, the similarity function can be present.

Figure 3 shows variations in the M–O similarity functions ϕ_{m} and ϕ_{h} with the atmospheric stability parameter based on the observational data. The values of the similarity functions are somewhat different to those determined by Dyer, especially under stable conditions. The scatter of both ϕ_{m} and ϕ_{h} is larger than the Dyer empirical equations under stable conditions. From this, the new curves of ϕ_{m} and ϕ_{h} can be fitted (Fig. 3), and the fitted equations are as follows:

$$\phi_{\mathrm{m}} = \begin{cases} (1 - 18.3\zeta)^{-\frac{1}{4}} & \zeta \le 0 \\ 1 + 10.1\zeta & \zeta > 0 \end{cases} \tag{13}$$

$$\phi_{\mathrm{h}} = \begin{cases} (1 - 25.3\zeta)^{-\frac{1}{2}} & \zeta \le 0 \\ 1 + 12.1\zeta & \zeta > 0 \end{cases} \tag{14}$$

Using Eqs. (9) and (10), the integrated forms of ϕ_{m} and ϕ_{h} are obtained as follows:

$$\psi_{\mathrm{m}} = \begin{cases} 2\ln(\frac{1+x}{2}) + \ln(\frac{1+x^2}{2}) - 2\arctan x + \frac{\pi}{2} & \zeta \le 0 \\ -10.1\zeta & \zeta > 0 \end{cases} \tag{15}$$

where $x = (1 - 18.3\zeta)^{\frac{1}{4}}$.

$$\psi_{\mathrm{h}} = \begin{cases} 2\ln\left[(1 + x^2)/2\right] & \zeta \le 0 \\ -12.1\zeta & \zeta > 0 \end{cases} \tag{16}$$

where $x = (1 - 25.3\zeta)^{\frac{1}{4}}$.

Soil temperature gradient and soil thermal conductivity

The soil temperature gradient and soil thermal conductivity are important soil parameters, and are often expressed as follows:

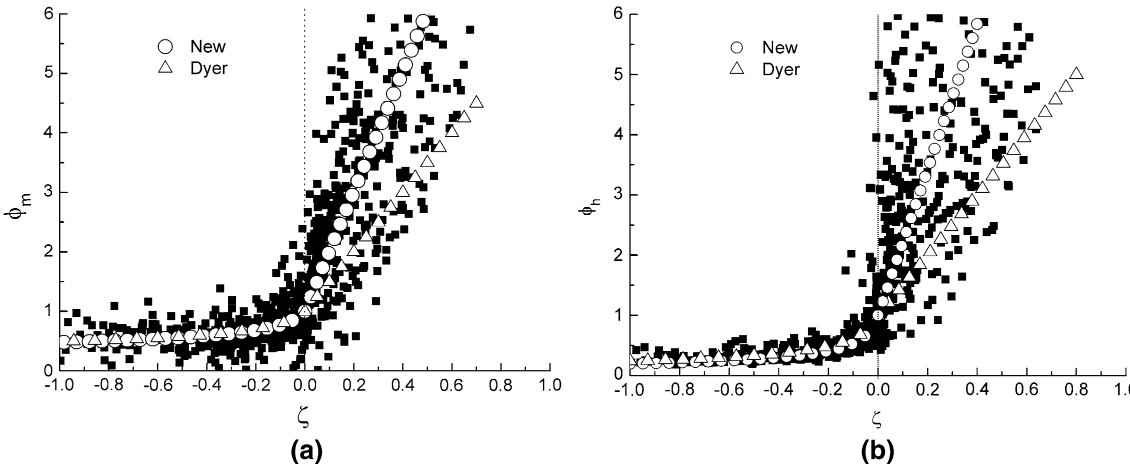

Fig. 3 The similarity function with stability parameter. *Dyer* Dyer empirical equations, *New* fitted equations. **a** Momentum, **b** temperature

$$\lambda = -G\frac{T_2 - T_1}{z_2 - z_1} \tag{17}$$

where G is the soil heat flux, λ is the soil thermal conductivity, and $(T_2 - T_1)/(z_2 - z_1)$ is the soil temperature gradient. The soil heat flux is directly influenced by the soil temperature gradient. Using Eq. (17), and the measured soil temperature (at depths of 0, 5, and 10 cm) and heat flux (at depths of 2.5 and 7.5 cm), the mean soil thermal conductivities at depths of 2.5 and 7.5 cm were 0.199 and 0.201 W (m K)$^{-1}$, respectively. Soil thermal conductivity at the two depths is similar because the sites are located close to each other and the soil composition is the same, being dry sand.

Conclusions

Observations of energy fluxes and micrometeorological environmental variables were conducted over the Dunhuang Gobi land surface during 7–28 August 2008 to provide the data required for simulation studies. The results provide a clear understanding of the effect of surface conditions on the surrounding environment.

Using the relative reflected radiation as a weighting factor, mean albedo over the Dunhuang Gobi was calculated by the integral 0.238. The roughness length was calculated using the friction velocities obtained by the eddy-correlation method. To remove the influence of anomalous roughness lengths on the mean value, a least squares method was developed. This method had a positive effect, even in the case of very scattered roughness lengths, as the values of the momentum and temperature roughness lengths were 0.045 and 0.005 cm, respectively. Combining the gradient and eddy-correlation data, an improved fit of the equations of the M–O similarity functions was also obtained. The mean soil thermal conductivities were 0.199 and 0.201 W (m K)$^{-1}$ at depths of 2.5 and 7.5 cm, respectively.

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