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The Sensitivity of Ground Surface Temperature Prediction to Soil Thermal Properties Using the Simple Biosphere Model (SiB2)

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ABSTRACT

Using the Simple Biosphere Model (SiB2), soil thermal properties (STP) were examined in a Tibetan prairie during the monsoon period to investigate ground surface temperature prediction. We improved the SiB2 model by incorporating a revised force-restore method (FRM) to take the vertical heterogeneity of soil thermal diffusivity (k) into account. The results indicate that (1) the revised FRM alleviates daytime overestimation and nighttime underestimation in modeled ground surface temperature ($T_{\rm g}$), and (2) its role in little rainfall events is significant because the vertical gradient of k increases with increasing surface evaporation. Since the original formula of thermal conductivity (λ) in the SiB2 greatly underestimates soil thermal conductivity, we compared five algorithms of λ involving soil moisture to investigate the cause of overestimation during the day and underestimation at night on the basis of the revised FRM. The results show that (1) the five algorithms significantly improve $T_{\rm g}$ prediction, especially in daytime, and (2) taking one of these five algorithms as an example, the simulated $T_{\rm g}$ values in the daytime are closer to the field measurements than those in the nighttime. The differences between modeled $T_{\rm g}$ and field measurements are mostly within the margin of error of ± 2 K during 3 August to 4 September 1998.

Key words: land surface processes, SiB2, sensitivity, ground surface temperature, soil thermal diffusivity, soil thermal conductivity

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1. Introduction

The thermal properties of soils have a significant effect on partitioning of energy fluxes at the ground surface. When thermal properties are related to soil temperature, the energy fluxes are more accurately associated with the transfer of heat through the soil. Thermal properties and soil temperature have a close relationship theoretically (de Vries, 1963; Farouki, 1986). Regarding the variations in thermal properties within a range of temperature, Campbell et al. (1994) found that thermal conductivity increases dramatically with temperature in moist soil. Another important factor regarding the effect of temperature on soil thermal properties is whether or not the soil is frozen (Kersten, 1949). Likewise, thermal properties also influence the variability of soil temperature spatially and temporally according to the Fourier's law. Using a soil vegetation atmosphere transfer scheme (SVATS), Peters-Lidard et al. (1998) quantified the sensitivity of the modeled

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surface ground temperature to two different formulas for thermal conductivity. The maximum differences between the two formulas in spatially averaged skin and 7.5 cm-deep soil temperatures were 1.5 K and 2.6 K, respectively.

Through analyses of micrometeorological field measurements of the BJ site during the monsoon period, Gao et al. (2004) found that the Simple Biosphere Model (SiB2) (Randall et al., 1996; Sellers et al., 1996a, b) generated a warmer ground surface during the day and a colder ground surface at night. Similar conclusions were also drawn by Doran et al. (1998) and Kim et al. (2001). Our objective in this study was therefore to investigate the sensitivity of modeled ground surface temperature ($T_{\rm g}$) to two factors. First, we applied a revised force-restore method (FRM) to the SiB2 model in order to explore the effect of vertical heterogeneity of thermal diffusivity (k) on the simulation of $T_{\rm g}$. Second, we compared several different algorithms of thermal conductivity (λ) in the SiB2 in order to understand the influence of λ involving soil moisture on $T_{\rm g}$ prediction.

2. Materials and methodology

2.1 *Field experiment*

The field experiment was conducted at the BJ site (31°37′N, 91°90′E and 4580 m above sea level) near Naqu during the intensive observation period (IOP) of GEWEX (Global Energy and Water cycle Experiment) Asian Monsoon Experiment on the Tibetan Plateau (GAME/Tibet) in 1998 (Koike et al., 1999; Choi et al., 2004). Details on field measurements (see Fig. 1), parameter settings and initial conditions used in SiB2 were presented in Gao et al. (2004), and details on the instruments and various data processing techniques are provided at the web site: http://monsoon.t.u-tokyo.ac.jp/tibet/data/iop/pbltower/doc/naqu-fx.txt.

2.2 *The traditional and revised force-restore methods (FRM) in the SiB2 model*

The traditional FRM (Bhumralkar, 1975; Deardorff, 1978), owing to its computational efficiency and reasonable physical foundation, is used in the SiB2 model for soil temperature prediction. It involves two prognostic equations based on heat conduction—one for the ground surface temperature ($T_{\rm g}$), representing the diurnal signals of the canopy and the soil surface, and the other for a deep temperature ($T_{\rm d}$) toward which $T_{\rm g}$ is relaxed, representing the annual wave, as follows.

$$\frac{\partial T_{\rm g}}{\partial t} = \frac{2(R_{\rm n} - H - LE)}{c_{\rm g}\sqrt{2k/\omega}} - \omega(T_{\rm g} - T_{\rm d}) \,, \qquad (1)$$

$$\frac{\partial T_{\rm d}}{\partial t} = \frac{R_{\rm n} - H - LE}{c_{\rm g}(365\pi)^{1/2}\sqrt{2k/\omega}} \,, \qquad (2)$$

where $R_{\rm n}$ is the net radiation flux for ground surface, and H and LE are the ground sensible heat flux and latent heat flux, respectively. $c_{\rm g}$ is the volumetric heat capacity of soil. ω is the angular velocity of the Earth's rotation. Deardorff (1978) defined the penetration depth of the annual temperature wave as $(365\pi)^{1/2}\sqrt{2k/\omega}$, which is $\pi^{1/2}$ times the e-folding depth of the annual temperature wave and has relationship with the form of heat transfer and the thermal properties of the soil. This determination is analogous to the forcing method (Arakawa, 1972; Corby et al., 1972).

However, the contribution of convective heat transport to the "fast" variation of surface temperature in moist soil is important (Hopmans et al., 2002). Therefore we investigated a revised force-restore method (Gao et al., 2008) which takes the soil heterogeneity with depth and the occurrence of convective heat transfer into account.

The transient conductive-convective equation is

$$\frac{\partial T}{\partial t} = k\frac{\partial^2 T}{\partial z^2} + W\frac{\partial T}{\partial z} \,,$$

where

$$W = \frac{\partial k}{\partial z} - \frac{c_{\rm w}}{c_{\rm g}} w\theta \,,$$

$c_{\rm w}$ is the heat capacity of water, w is the liquid water flux (m^3 s^{-1} m^{-2}) (positive downward), and θ is the volumetric water content (m^3 m^{-3}). W quantifies the vertical heterogeneity in thermal diffusivity and water movement in soil, both are attributable to the phase changes within the surface layer. Gao et al. (2008) indicated that $\partial k/\partial z$ is the main contributor to W, because w is usually expected to be only a few millimeters per day of evaporation flux.

Then ground surface temperature equation in SiB2 is transformed as

$$\frac{\partial T_{\rm g}}{\partial t} = \frac{2(R_{\rm n} - H - LE)}{c_{\rm g}(2k/N\omega)} - \omega\left(\frac{N}{M}\right)(T_{\rm g} - T_{\rm d}) \,, \qquad (3)$$

where

$$M = \frac{1}{\left(\dfrac{W}{2k} + \dfrac{\sqrt{2}}{4k}\sqrt{W^2 + \sqrt{W^4 + 16k^2\omega^2}}\right)} \,,$$

$$N = \frac{\sqrt{W^2 + \sqrt{W^4 + 16k^2\omega^2}}}{\sqrt{2}\omega} \,.$$

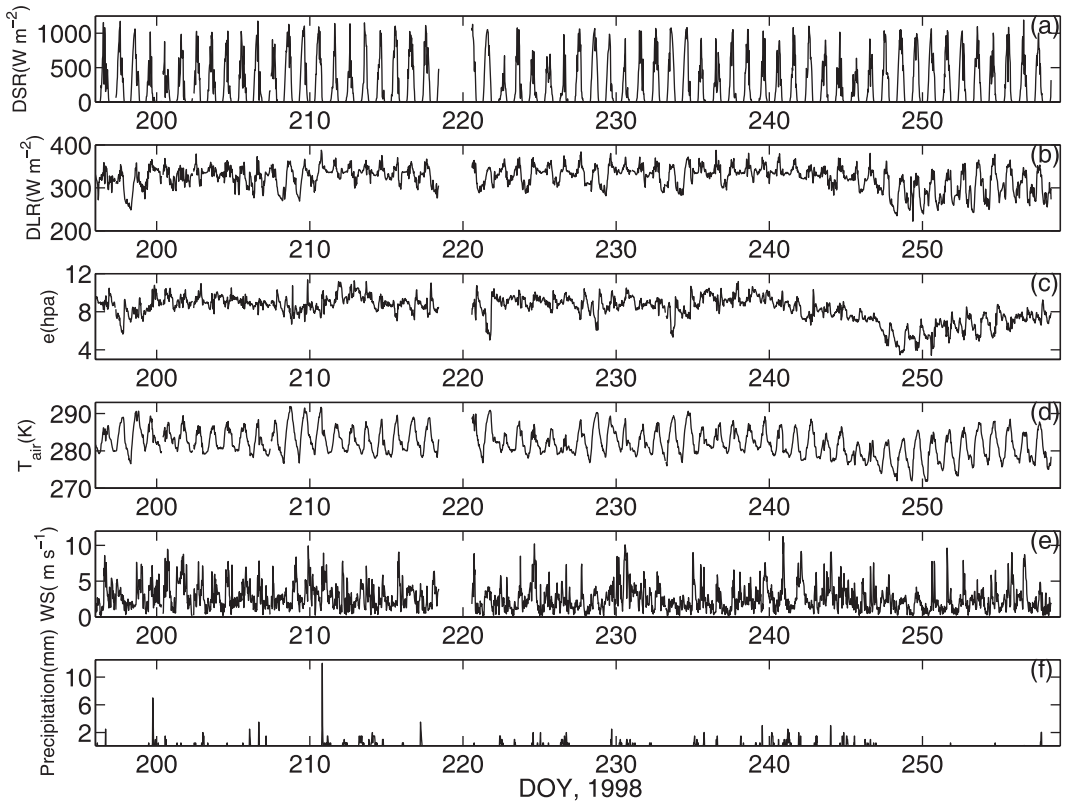

Fig. 1. Temporal variation of the atmospheric forcing quantities: short-wave downward radiation (a), long-wave downward radiation (b) measured at 1.5 m height, vapor pressure (c), air temperature (d) and absolute horizontal wind speed (e) measured at 3.5 m height, and precipitation (f) on a 30-min basis (Gao et al., 2004, modified by permission of American Geophysical Union. Copy right [2004] American Geophysical Union.)

When $W = 0, N = M = \sqrt{2k/\omega}$, and Eq. (3) is the same as Eq. (1) for homogeneous soil experiencing conduction-only heat transfer.

Regarding the revised FRM in SiB2, we assumed that heat capacity (c_g) is only related to the wetness (=volumetric water content/saturated water content) in surface layer, while the thermal conductivity (λ) at surface and depth at $z = 0.02$ m are related to the wetness in surface layer and the wetness in root zone, respectively. Thus the vertical heterogeneity of $k(= \lambda/c_g)$ is determined by the vertical gradient of k between surface and 0.02 m soil depth. In SiB2, energy transfer (owing to phase changes) by the interception water stores from within the top soil layer is also a contributor to ground temperature variation. This means that, due to the vertical gradient of thermal diffusivity (k), convective heat transfer becomes as important as conduction. Therefore the replacement of the original governing equation in the SiB2 model with the revised one is reasonable. Deep temperature, T_d, is still determined by Eq. (2) because the limited convection in the deep layer is neglected in its "slow" variation.

2.3 *Soil thermal conductivity* (λ) *in SiB2 model*

The thermal conductivity theoretically depends in a rather complicated way on its composition, especially its water content. When moisture is added, a thin water film develops that increases the effective contact area between the soil particles, thereby increasing thermal conductivity. There are, therefore, many studies for determining soil thermal conductivity. In SiB2, the original formula of λ is proposed by Camillo and Schmugge (1981), and it will be replaced by five different algorithms of λ in turn to compare the simulations of T_g. Table 1 summarizes these five representative formulas for λ, each of which involves water content in the soil.

2.3.1 *de Vries (1963) and Ochsner et al. (2001)*

The method of de Vries (1963) is a physical model in which the thermal conductivity is a function of the geometric arrangement of the phases in the soil matrix. f_i is the multiplication factor dependent on the thermal conductivity of ith phase and the shape of the

enclosure. $f_0 = 1$ for the continuous fluid surrounding the solid particles (i.e., air for dry soil or water for moist soil). Other values of f_i are calculated from

$$f_i = \frac{1}{3}\sum_{j=1}^{3}\left[1+\left(\frac{\lambda_i}{\lambda_0}-1\right)g_j\right]^{-1}.$$

We assumed that water is the continuous fluid at low water content. Following Ochsner et al. (2001), the critical water content was assigned to be 0.15 m^3 m^{-3} and the thermal conductivities of water (λ_0) and of air-filled pores were assigned to be 0.596 W m^{-1} K^{-1} and 0.099 W m^{-1} K^{-1}, respectively. The shape factor $g_1 = 0.144$ for the soil solids and the value of g_1 for the air-filled pores is given by

$$g_1 = 0.333-(1-S_\mathrm{r})(0.333-0.035), \quad \text{when } \theta >= 0.15,$$
$$g_1 = 0.013+\frac{\theta}{0.15}(g_{1\mathrm{c}}-0.013), \quad \text{when } \theta < 0.15,$$

where $g_{1\mathrm{c}}$ is the value of the first g_1 expression at the critical water content 0.15 m^3 m^{-3}. θ is the volumetric water content and S_r is the wetness (= volumetric water content/saturated water content) of soil.

2.3.2 *Johansen (1975), and Côté and Konrad (2005)*

The Johansen method, as detailed in Côté and Konrad (2005), is a similar interpolation approach in which the estimation of soil thermal conductivity is based on the thermal conductivity of totally dry soil (λ_dry) and of saturated soil (λ_sat).

The Kersten number (K_e), as a function of the degree of saturation, follows the empirical relationship studied by Kersten (1949). It is the normalized thermal conductivity. Here

$$K_\mathrm{e} = \frac{\kappa S_\mathrm{r}}{1+(\kappa-1)S_\mathrm{r}},$$

where κ is an empirical parameter dependent on the various soil types in the unfrozen or frozen state. In this study we assigned κ to be 3.55.

The determination of the thermal conductivity of saturated unfrozen soil is only related to the solid particles (λ_s) and water (λ_w) by the formula

$$\lambda_\mathrm{sat} = \lambda_\mathrm{s}^{1-\varepsilon}\lambda_\mathrm{w}^{\varepsilon}$$

In contrast to λ_sat, the determination of λ_dry proposed by Côté and Konrad (2005) is

$$\lambda_\mathrm{dry} = \chi \times 10^{-\eta * \varepsilon},$$

where χ (W m^{-1} K^{-1}) and η (no unit) are material parameters accounting for the particle shape effect, ε is the porosity of the soil. In this study, χ and η were assigned to be 1.7 W m^{-1} K^{-1} and 1.8, respectively.

2.3.3 *Campbell (1985)*

This method is one of most typical empirical correlations to estimate soil thermal conductivity. In this method, coefficients A, B, C, D and E are obtained by curve fitting and are related to soil types and compositions. Their formulas are

$$A = \frac{0.57+1.73\varphi_\mathrm{q}+0.93\varphi_\mathrm{m}}{1-0.74\varphi_\mathrm{q}-0.49\varphi_\mathrm{m}} - 2.8\varphi_\mathrm{s}(1-\varphi_\mathrm{s}),$$
$$B = 2.8\varphi_\mathrm{s}\theta,$$
$$C = 1+2.6m_\mathrm{c}^{-1/2},$$
$$D = 0.03+0.7\varphi_\mathrm{s}^2,$$
$$E = 4,$$

Table 1. Five formulas for determining soil thermal conductivity (λ).

No.	Formula	Reference
1	$\lambda = \dfrac{\sum_{i=0}^{n} f_i X_i \lambda_i}{\sum_{i=0}^{n} f_i X_i}$	(de Vries, 1963; Ochsner et al., 2001)
2	$\lambda = K_\mathrm{e}(\lambda_\mathrm{sat}-\lambda_\mathrm{dry})+\lambda_\mathrm{dry}$	(Johansen, 1975; Côté and Konrad, 2005)
3	$\lambda = A+B\theta-(A-D)\exp\left[-(C\theta)^E\right]$	(Campbell, 1985)
4	$S_\mathrm{r} = a_1[\sinh(a_2\lambda+a_3)-\sinh(a_4)]$	(adapted after Becker et al., 1992; Becker and Fricke, 1997)
5	$\lambda = \dfrac{1}{3}\left[\dfrac{1-\varepsilon}{2\lambda+b}+\dfrac{S_\mathrm{r}\varepsilon}{2\lambda+\lambda_\mathrm{w}}+\dfrac{(1-S_\mathrm{r})\varepsilon}{2\lambda+\lambda_\mathrm{a}}\right]^{-1}$	(adapted after Sundberg, 1988; Tarnawski and Leong, 2000)

Table 2. Correlation coefficients in Becker's methodology (sand only).

Soil type and state		a_1			a_2			a_3			a_4		
		low	mid	high	low	mid	high	low	mid	high	low	mid	high
sand	frozen	26.0	10.0	15.0	1.84	1.66	1.18	−1.0	−2.2	−1.8	−0.735	−1.625	−0.44
	unfrozen	6.4	6.8	6.8	5.55	2.77	3.47	−3.2	−2.9	−7.5	−2.0	−1.5	−2.0

where φ_q is the volume fraction of quartz, φ_m is the volume fraction of other minerals, and $\varphi_s = \varphi_q + \varphi_m$ is the volume fraction of solids. Meanwhile coefficient C is highly correlated with the clay fraction, m_c.

2.3.4 *Becker et al. (1992), Becker and Fricke (1997)*

Based on measured thermal conductivity data available in the literature, another empirical methodology was developed by Becker et al. (1992). In this formula, a_1, a_2, a_3 and a_4 are coefficients dependent upon soil type in either the frozen or unfrozen state. Their values are given in Table 2 (Becker and Fricke, 1997).

In this study, a_1, a_2, a_3 and a_4 were assigned to be 6.8, 2.77, −2.9 and −1.5, respectively, according to the soil type of the BJ site (unfrozen sand).

2.3.5 *Sundberg (1988), and Tarnawski and Leong (2000)*

This method is called the self-consistent approximation (SCA). It assumes a uniform effective medium surrounding solid grain, a statistical proximity between the various phases of the composite material (proportional to their volume fractions), and water surrounding a solid phase as a continuous medium. Because of overestimation to soils using the SCA model, Sundberg (1988) modified the model by introducing a thermal contact resistance coefficient between solids, as follows.

$$b = \alpha \cdot \lambda_s,$$

where

$$\alpha = K_e \cdot (\lambda_{sat} - \lambda_{dry}) + \lambda_{dry}, \quad K_e = (A \cdot \log S_r + 1),$$

$$\lambda_{sat} = \frac{1}{\left[\dfrac{\beta}{\lambda_s} + \dfrac{1-\beta}{\lambda_w}\right] \cdot \lambda_s}, \quad \lambda_{dry} = \frac{1}{\left[\dfrac{\beta}{\lambda_s} + \dfrac{1-\beta}{\lambda_a}\right] \cdot \lambda_s},$$

$$\beta = 1 - 0.12833\varepsilon + 0.06461\varepsilon^2 + 0.06491\varepsilon^3 .$$

The constant A is 0.95, and the empirical constant, β, representing a solid phase at the soil micro-porous level, is a function of porosity, ε. λ_a is the thermal conductivity of air.

3. Results

3.1 *Effect of thermal diffusivity* (k) *on* T_g *prediction*

To determine whether the revised FRM has any advantage over the traditional one in T_g prediction, we compared simulated T_g using these two methods in the SiB2 model.

The results show that, even though the revised FRM still overestimated T_g during the day and underestimated T_g at night, the included angle between modeled and measured T_g was 0.016°, which was smaller than that between modeled T_g by original FRM and field measurements, 0.022° (see Fig. 2). With respect to the difference in modeled T_g, which indicates T_g (the original) minus T_g (the revised), the total amplitudes > 1.5 K occurred during 28–30 July, 16–18 August, 21–23 August, and 5–15 September corresponding to the little rainfall events. Among these, the peak and trough values during 28–30 July were 1.33 K, −1.04 K, 2.58 K, −1.11 K and 3.2 K, respectively, and during 16–18 August they were −0.6 K, 1.12 K, −1.19 K and 1.64 K, respectively. The peak and trough values during 21–23 August were −0.76 K, 1.57 K, −1.54 K, 2.92 K and −0.9 K, respectively, and during 5–15 September they were −0.84 K, 0.89 K, −1.45 K, 1.57 K, −1.98 K, 2.18 K, −2.22 K, 2.16 K, −2.08 K, 2.84 K, −2.64 K, 3.31 K, −2.93 K, 3.26 K, −2.17 K, 3.18 K, −2.58 K, 2.56 K, −2.48 K and 3.55

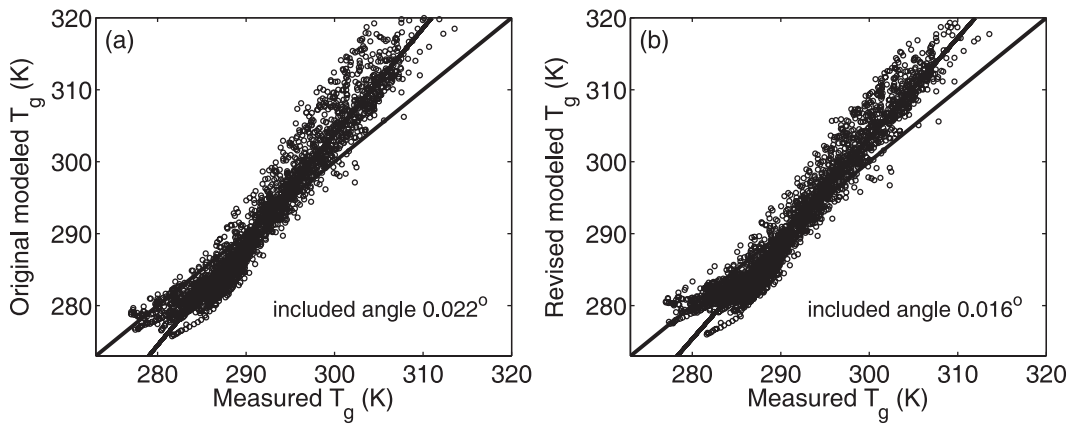

Fig. 2. Scatterplots of the modeled T_g by the original FRM relative to measurements (a), the modeled T_g using the revised FRM relative to field measurements (b).

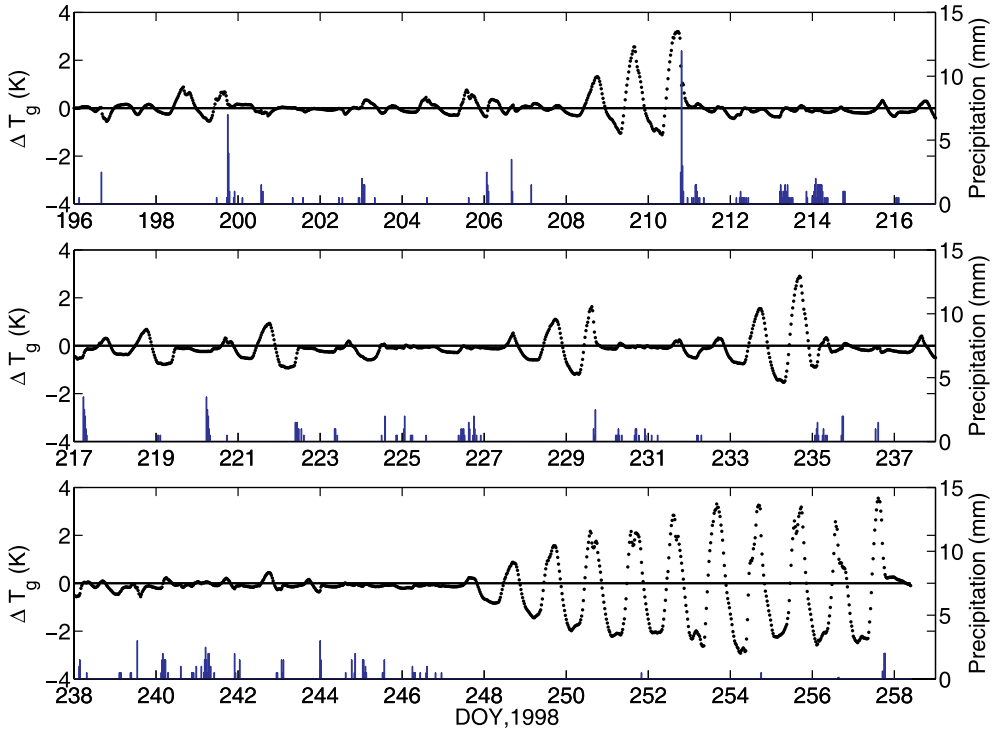

Fig. 3. The difference in the modeled T_g between the original FRM and the revised FRM (dotted line) relative to precipitation measurements (bar).

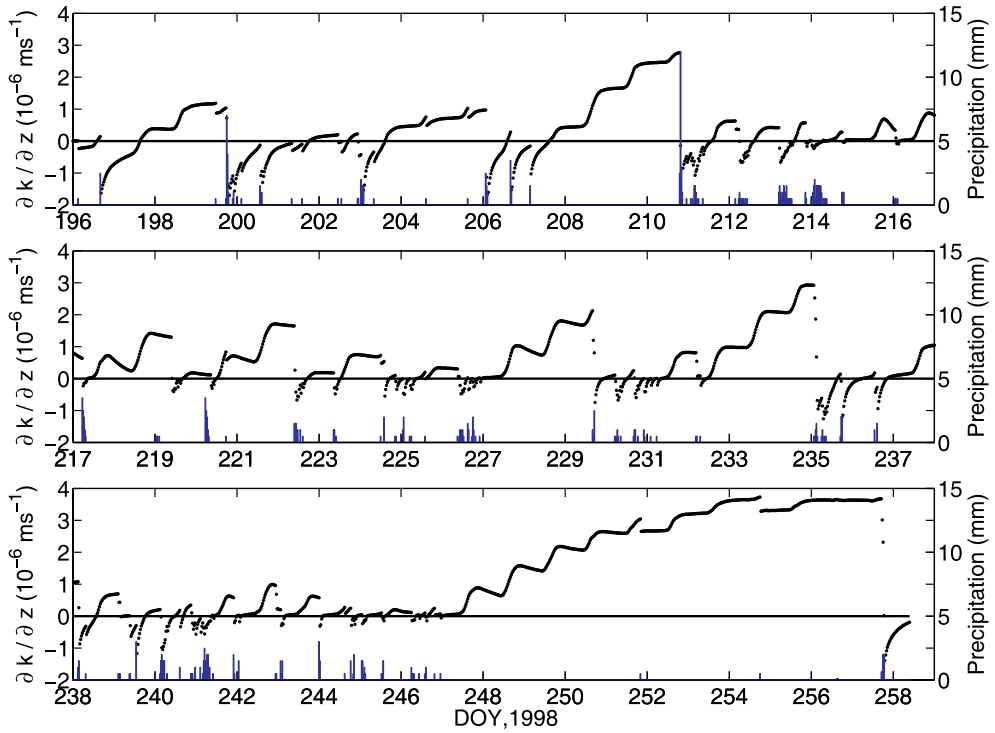

Fig. 4. The vertical gradients of k in the revised FRM (dotted line) relative to precipitation measurements (bar).

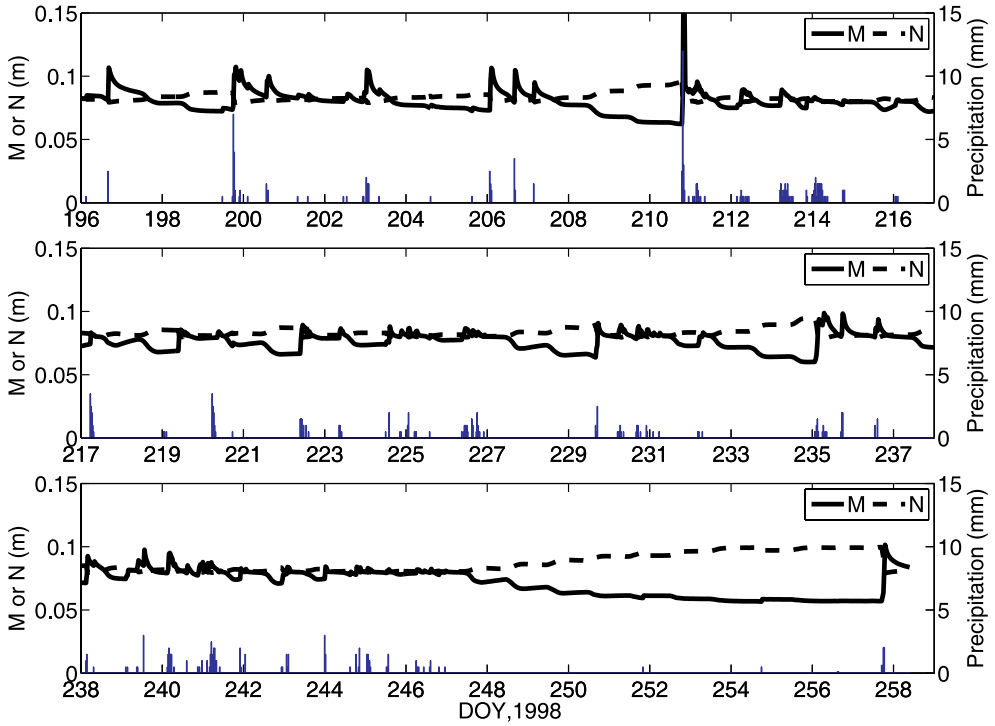

Fig. 5. Parameters M (solid line) and N (dashed line) in the revised FRM relative to precipitation measurements (bar).

K, respectively (see Fig. 3).

Generally speaking, surface layer evaporation often increases on days of little rainfall. Then the perpendicular heterogeneities of thermal properties become clear and the vertical gradient of thermal diffusivity (k) would increase with increasing evaporation, which is the main contributor to W in the revised FRM. The maximum gradients of k during 28–30 July, 16-18 August, 21–23 August, and 5–15 September were 2.77×10^{-6} m s^{-1}, 2.13×10^{-6} m s^{-1}, 2.93×10^{-6} m s^{-1}, and 3.67×10^{-6} m s^{-1}, respectively (see Fig. 4). Regarding parameters M and N, which are closely related to W, their variations started decreasing and increasing, respectively, at the beginnings of these four little rainfall periods and did not stop until the vertical homogeneity appeared (see Fig. 5).

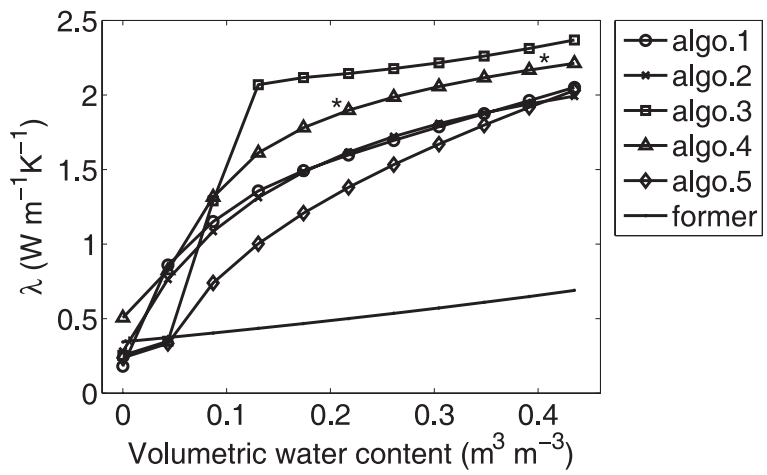

Fig. 6. Calculated thermal conductivity as a function of volumetric water content by the five different algorithms and the former formula in SiB2 [data with marker * is from Garratt (1994)].

Thus, the description by the revised FRM is much closer to the real soil condition than that by the traditional FRM, because the involved convective heat transfer significantly influences the variability of T_g.

3.2 *Effect of thermal conductivity on T_g prediction*

As previously mentioned, in the revised FRM the vertical heterogeneity of thermal diffusivity (k) necessarily affects the variation of T_g. However, the revised FRM still did not satisfactorily improve the modeled T_g regarding the daytime overestimation and nighttime underestimation. Therefore, we continued to explore the effect of thermal conductivity (λ) on T_g. Five algorithms were compared, including the classical theoretical algorithm, the modified self-consistent approximation algorithm, and the empirical algorithms utilizing curve fitting.

To illustrate the necessity of the study about λ, the original formula of λ in SiB2 was first compared with these five formulas as a function of volumetric water content. Taking the data from Garratt (1994) as reference (* in Fig. 6), the large underestimation of λ with water content variation by the former formula is 2 W m^{-1} K^{-1} at most. By contrast, the estimations of λ by these five algorithms more closely reflected the trend of the reference data. All five algorithms yielded a positive relationship between thermal conductivity and water content, reflecting the important role of water in soil. The maximum difference of λ among them occurred between algorithms 3 and 5 and was 1 W m^{-1} K^{-1} at the same water content, but the difference between algorithms 1 and 2 was tiny at any water content (see Fig. 6).

With respect to the simulations of T_g by the five algorithms, we eliminated the "phase drift" because of the time integral. So the analyses presented were based on no "phase drift". The daytime overestimations of the simulated T_g using these five algorithms decreased obviously relative to the original formula of λ. However, the improvement in nighttime was not as significant as in daytime even though the nighttime results approached field measurements (see Fig. 7).

Since the amplitudes of the differences among the five algorithms were generally < 2 K (figure omitted), we took algorithm 1 as an example. The daytime estimations of T_g were closer to the field measurements than those at night (see Fig. 8). The differences between the simulated T_g and field measurements were greatly reduced by algorithm 1 in contrast to the former calculation of λ in SiB2. During the period from 3 August to 4 September, the amplitudes were mostly within the margin of error of ± 2 K (see Fig. 9).

In addition, three statistical analyses concerning the modeled results $(\mathrm{sim})_i$ relative to measurements $(\mathrm{obs})_i$ were calculated using the following equations:

$$\mathrm{BIAS} = \sum_{i=1}^{n} \frac{(\mathrm{sim})_i - (\mathrm{obs})_i}{n}$$

$$\mathrm{SEE} = \sqrt{\frac{\sum_{i=1}^{n} \left[(\mathrm{sim})_i - (\mathrm{obs})_i\right]^2}{n-2}}$$

$$\mathrm{NSEE} = \sqrt{\frac{\sum_{i=1}^{n} \left[(\mathrm{sim})_i - (\mathrm{obs})_i)^2\right]}{\sum_{i=1}^{n} (\mathrm{obs})_i^2}},$$

where BIAS means the slope in the 1:1 linear regression, SEE indicates the standard error of the estimate, and NSEE is the normalized SEE, denoting an estimate of relative uncertainty.

Table 3. Statistical analysis of modeled T_g by the former formula and the five algorithms of λ.

No.	BIAS	SEE	NSEE
former	−0.2557	3.6730	0.0126
1	−1.1001	2.3756	0.0082
2	−1.1090	2.4083	0.0083
3	−1.3956	2.8441	0.0098
4	−1.2715	2.5341	0.0087
5	−0.9739	2.4847	0.0085

Table 3 lists a summary of results of the statistical analyses involving the modeled results and filed measurements. Because of the offsets between underestimations and overestimations, the former calculation has the lowest absolute value of BIAS. On the contrary, the five algorithms mainly underestimated T_g at night and they had relatively higher absolute values of BIAS. Except the BIAS, algorithm 1 had the lowest values of SEE and NSEE, while the former algorithm in SiB2 had the highest values of SEE and NSEE, 3.6730 and 0.0126, respectively.

4. Conclusions

The simulated ground surface temperatures using the SiB2 model are always overestimated during the day and underestimated at night (Doran et al., 1998; Kim et al., 2001; Gao et al., 2004). In view of the close link between soil temperature and thermal properties, we investigated the influences of soil thermal properties on the prediction of ground surface temperature (T_g). The effect of thermal diffusivity (k) on T_g simulation were examined by applying a revised FRM to the SiB2 model. This revised FRM, concerning both

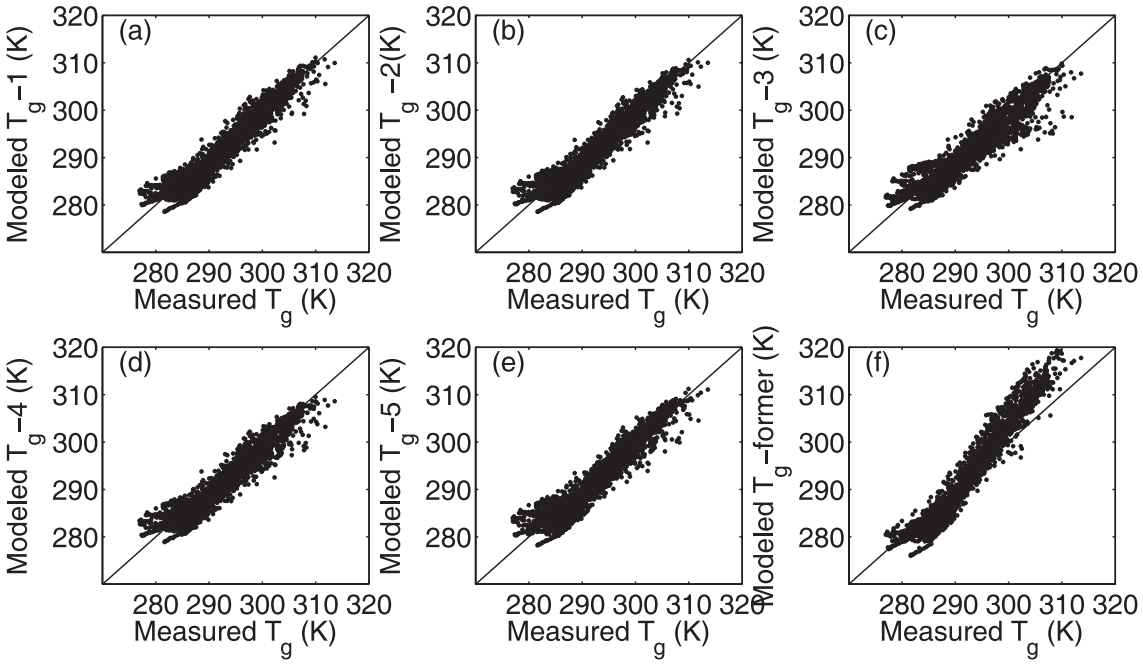

Fig. 7. Scatterplots of modeled T_g using algorithm 1 (a), algorithm 2 (b), algorithm 3 (c), algorithm 4 (d), and algorithm 5 (e) and using the former formula of λ (f) relative to measurements.

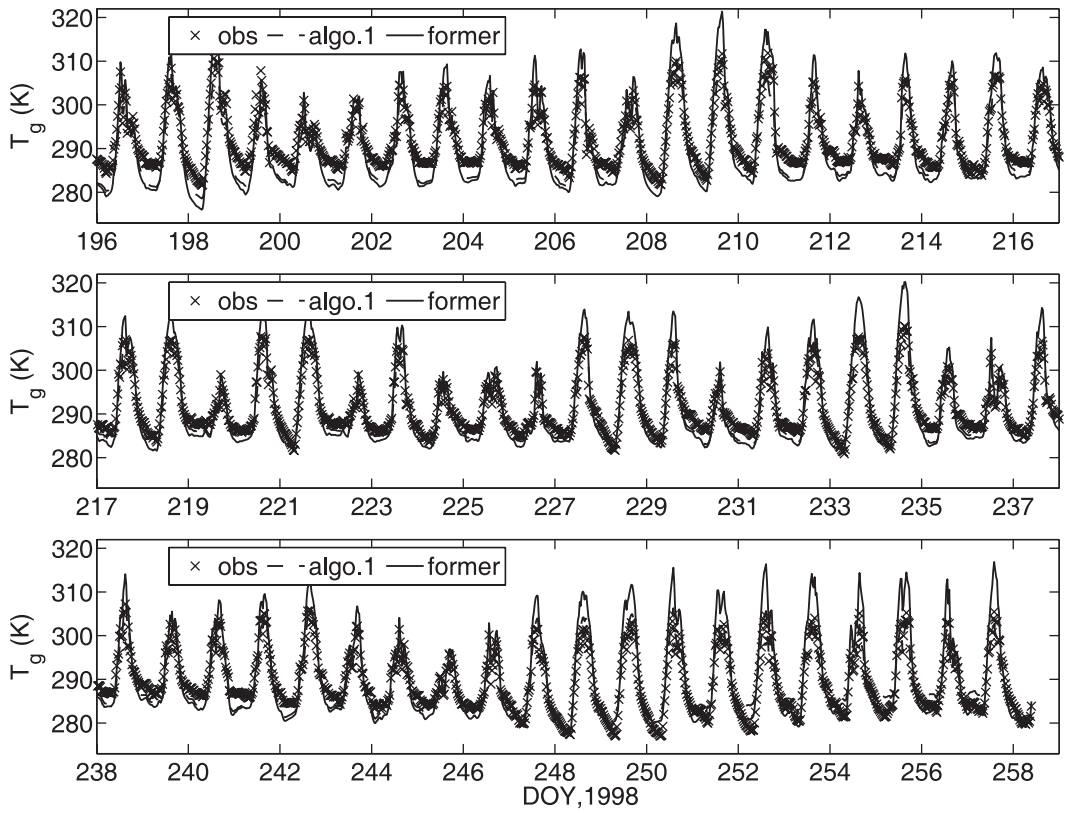

Fig. 8. Comparison of the modeled T_g using algorithm 1 (dotted line) and using the former formula of λ (solid line) relative to filed measurements (marker x).

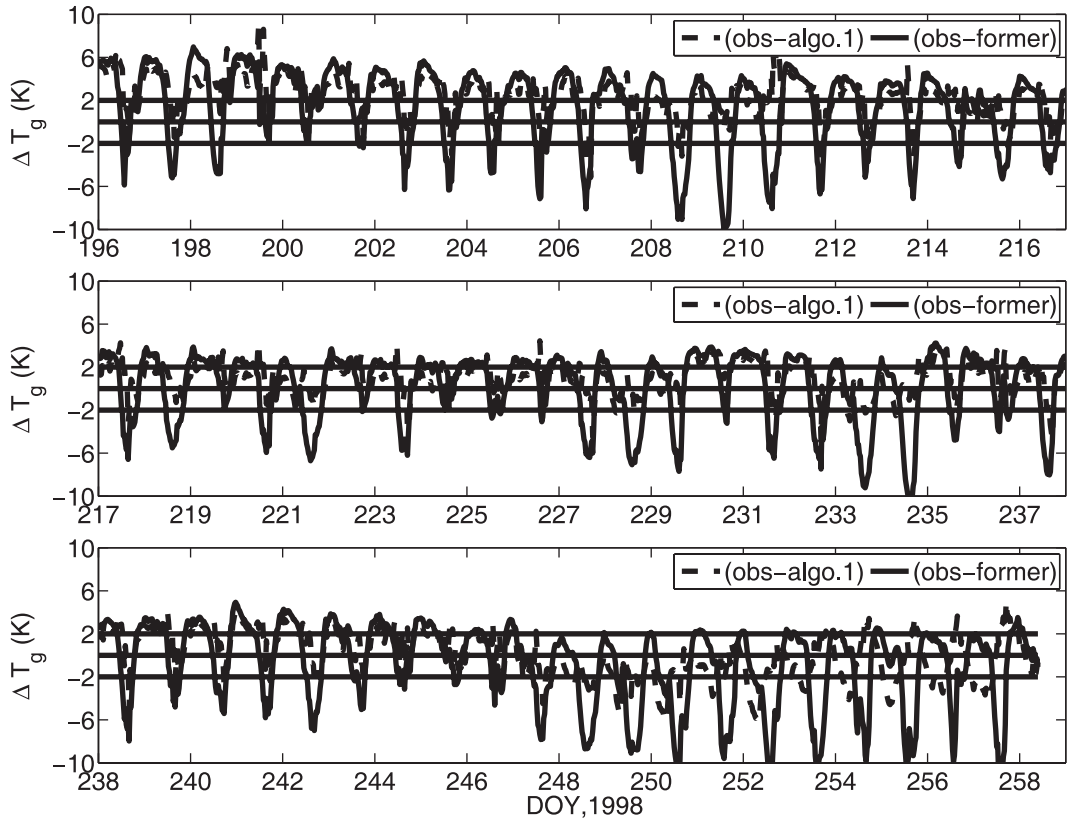

Fig. 9. Difference between the measured T_g and modeled T_g using algorithm 1 of λ (dotted line), and difference between measured T_g and modeled T_g using the former formula of λ (solid line).

conductive and convective heat transfer, takes the vertical heterogeneity of k and water movement in soil into account. The alleviations of < 3.55 K in modeled T_g were improved by the revised FRM, but the trend of daytime overestimation and nighttime underestimation was not altered radically. Therefore, we analyzed further the effect of thermal conductivity (λ) on T_g prediction by comparing former formula of λ in SiB2 and other five representative algorithms of λ. Taking the data from Garratt (1994) as a reference, the former formula greatly underestimated the soil thermal conductivity with increasing water content, and the estimations of λ by the five algorithms were closer to the trend of reference data. These five algorithms significantly improved the T_g prediction, especially in daytime. Because the differences in simulated T_g among the five algorithms were generally $< \pm 2$ K, we took algorithm 1 as an example to analyze the comparisons between measured T_g and modeled T_g in the entire time integration. The differences were mostly within the margin of error of ± 2 K during the period from 3 August to 4 September.

The results presented in this paper imply that, (1) the description using the revised FRM was much closer to the real soil conditions than that from the traditional FRM. But the general application of the revised FRM needs more evidences using other land-atmosphere models, (2) the improvements in nighttime using the five representative algorithms of λ were not as significant as those in daytime even though the nighttime results also approached the field measurements. Further explanations about the simulated T_g in nighttime are necessary to complement and support the conclusions in this study, (3) although algorithm 1 had relative advantages over others through the statistical analyses of modeled results relative to field measurements, it is difficult to determine which of those five algorithms is best from simulations at only one site, and (4) it is possible to revise the thermal conductivity parameterization in SiB2 according to different soil types and states. For frozen soil, moisture is largely affected by thawing and freezing processes which further complicates the determination of thermal conductivity (Penner, 1970; Farouki, 1981; Smith and Low, 1996; Smoltczyk, 2003).

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