Treatment of Undercanopy Turbulence in Land Models

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ABSTRACT

In arid and semiarid regions most of the solar radiation penetrates through the canopy and reaches the ground, and hence the turbulent exchange coefficient under canopy C_s becomes important. The use of a constant C_s that is only appropriate for thick canopies is found to be primarily responsible for the excessive warm bias of around 10 K in monthly mean ground temperature over these regions in version 2 of the Community Climate System Model (CCSM2). New C_s formulations are developed for the consistent treatment of undercanopy turbulence for both thick and thin canopies in land models, and provide a preliminary solution of this problem.

1. Introduction

The land surface component of climate models must address issues of the energy balance of soils and canopies. If the temperatures of canopies and underlying soil are determined separately, as in many land models, then it is necessary to parameterize the energy and moisture fluxes from the soil surface to canopy air space, and the same fluxes from the canopy to its air space. The sum of these fluxes in turn balances the sensible and latent heat exchanges to the overlying atmosphere. If the canopy is dense enough to transmit no solar radiation to the underlying soil, as assumed in the Biosphere–Atmosphere Transfer Scheme (BATS; Dickinson et al. 1993), the temperature of the underlying soil adjusts toward that of the canopy, and for a weak enough undercanopy turbulence, the energy exchanged is largely from the longwave radiation term. On the other hand, for a very sparse canopy, most of the solar energy penetrates to the underlying soil, and the longwave radiation exchange becomes negligible. This solar heating drives the soil surface toward a higher temperature than that of the canopy, with this temperature difference inversely proportional to the undercanopy turbulent transfer rate. Hence, if this rate for a sparse canopy is underestimated, the daytime soil temperature may be erroneously high, possibly affecting other land processes (e.g., surface energy, water, and carbon fluxes). Since observational constraints on the expected difference between canopy and soil temperatures are largely lacking, such error is only easy to diagnose if the undercanopy turbulent exchange has been severely underestimated.

The recently developed Common Land Model (CLM; Zeng et al. 2002; Dai et al. 2003) considers the radiative transfer through canopy but still uses the original undercanopy conductance formulation from BATS intended for a thick canopy. The initial climate simulation with CLM (Zeng et al. 2002) allowed for fractional vegetation cover (FVC) and a relatively large stem and dead leaf area index (SAI) (above unity for natural vegetation). The bare soil fraction was simulated without reference to canopy effects and the undercanopy soil temperature was simulated with enough shading to substantially reduce the solar heating. Hence, the soil temperature was not obviously in excess, even in semiarid regions. However, a newer version of CLM (referred to as the Community Land Model, or CLM2) was reformulated to be more compatible with datasets already in use at the National Cen-

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ter for Atmospheric Research (NCAR) that assumed land surfaces to be 100% vegetated (except over desert), with the sparse vegetation of semiarid regions represented by very small LAIs and SAIs. For instance, for the climate model grid cells in Arizona (a semiarid region), the FVC in CLM2 was assumed to be 1.0, the LAI varied from 0.05 to 0.3 for shrubs, and the SAI varied from 0.0 to 0.6. When Bonan et al. (2002) implemented this version into the NCAR Community Climate System Model (CCSM2), the daytime soil temperature was computed to be up to tens of degrees higher than the near-surface air temperature. Barlage and Zeng (2004) have addressed this issue from the perspective of a more realistic FVC dataset. This paper considers this issue further and provides initial parameterizations to more realistically determine undercanopy soil temperatures over the whole range of LAIs and SAIs that might be assumed in a climate model.

2. Methods

Land models can parameterize sensible fluxes from soil to canopy air space and from canopy to its air space, H_g and H_f, respectively, as (e.g., Zeng and Dickinson 1998)

$$H_g = \rho_a C_p C_s u_*(\theta_g - \theta_{af}) \quad \text{and} \tag{1}$$

$$H_f = \rho_a C_p C_f I_{ld} L_t u_*^{0.5}(\theta_f - \theta_{af}), \tag{2}$$

where ρ_a is the air density; C_p the specific heat of air; u_* the friction velocity above the canopy; L_t the sum of the leaf and stem area indexes; I_{ld} is the inverse square root of the characteristic plant surface dimension in the direction of wind flow and is typically 5 ($m^{-1/2}$) (e.g., Dickinson et al. 1993; Bonan 1996); θ_g, θ_f, and θ_{af} are the potential temperatures of the ground, canopy, and canopy air, respectively; and C_s (or C_f) is the turbulent transfer coefficient between the underlying soil (or canopy surface) and the canopy air. It needs to be emphasized that u_* does not represent the friction velocity undercanopy; instead, it is used to roughly represent the magnitude of the wind velocity incident on the leaves (Dickinson et al. 1993; Bonan 1996). The factors ($C_s u_*$) in (1) and ($C_f I_{ld} L_t u_*^{0.5}$) in (2) are further referred to as the undercanopy and canopy conductances, respectively. Following Dickinson et al. (1993), it is assumed in CCSM2 (Bonan et al. 2002; Zeng et al. 2002) that

$$C_f = 0.01(\text{m s}^{-1/2}) \quad \text{and} \quad C_s = 0.004. \tag{3}$$

For a thick canopy (e.g., $L_t = 6$) and a typical value (0.3 m s^{-1}) for u_*, (3) gives a undercanopy conductance that is smaller by two orders of magnitude than the canopy conductance. Since the solar heating of the soil is also two orders of magnitude smaller, this conductance difference does not give undercanopy soil temperatures that differ much from those of the canopy and canopy air space. For a sparse canopy (say, $L_t = 0.4$), however, most of the solar energy penetrates to the underlying soil but the undercanopy conductance is still smaller than the canopy conductance. To balance this solar heating through sensible and ground heat fluxes, the soil temperature in a land model (e.g., CCSM2) that uses the above C_s must be substantially (and unrealistically) elevated.

A different C_s formulation (Brutsaert 1982; Choudhury and Monteith 1988; Shuttleworth and Gurney 1990) has been used in some other land models:

$$C_s = bk\left(1 - \frac{d}{h}\right)\exp\left[-b\left(1 - \frac{z_{oc}}{h}\right)\right] \bigg/ \left[1 - \exp\left(-b\frac{d}{h}\right)\right], \tag{4}$$

where k is the von Kármán constant (0.4); d, z_{oc}, and h are the displacement height, roughness length, and height of the canopy, respectively; and the coefficient b is prescribed. This formulation was derived by assuming an exponential vertical profile of turbulent diffusivity and omitting the effect of atmospheric stability within and below the canopy. It does not consider countergradient turbulence transfer either. This formulation has been of special interest for interpreting radiometric temperatures from remote sensing when both canopy and soil contribute (i.e., for sparse canopies) (e.g., Friedl 1995). When it is applied to land modeling, Bonan (1996) assumed $b = 3$ and Lo Seen et al. (1997) assumed $b = 2.5$. Assuming $z_{oc} = 0.1h$ and $d = 2h/3$, $C_s = 0.031$ in Bonan and 0.043 in Lo Seen et al. These values are larger than that in (3).

When canopy disappears (i.e., $L_t = 0$), H_f should approach zero (i.e., C_f should remain finite), and C_s should approach its bare soil formulation. Based on the Monin–Obukhov similarity theory that considers the surface sublayer (or the variable ratio of the roughness length for momentum over that for heat, z_o/z_{oh}) over bare soil (Zeng and Dickinson 1998),

$$C_s = k/\left[\alpha\left(\frac{u_* z_o}{\upsilon}\right)^{0.45}\right], \tag{5}$$

where the coefficient $\alpha = 0.13$, υ is the kinematic viscosity of air (1.5×10^{-5} m^2 s^{-1}), and z_o is the roughness length for bare soil. Because the denominator is equal to ln (z_o/z_{oh}) with a typical value of 2 (Zeng and Dickinson 1998), the typical value of C_s is about 0.2, which is much larger than the value of 0.004 given for a thick canopy in (3). It is also larger than 0.031 from Bonan (1996) and 0.043 in Lo Seen et al. (1997).

In the more complicated first-order closure of turbulence within and under canopy, as used in the Simple Biosphere model (SiB2; Sellers et al. 1996), leaf area density varies with height, and z_{oc}/h and d/h vary with L_t and vegetation type. Furthermore, the undercanopy resistance $r_d = C_2/u_2$ with the coefficient C_2 depending on vegetation type and L_t. Qualitatively, the wind speed at the canopy top (u_2) is proportional to u_*, even though its exact expression is not analytical and its value has to be obtained numerically at each time step in SiB2. Therefore, roughly speaking, the C_s value in (1) is approximately proportional to $1/C_2$ in SiB2 and is primarily dependent on L_t and vegetation type. Because C_s in (4) or from SiB2 does not depend on u_*, it does not converge to the bare soil formulation in (5) as canopy disappears.

As emphasized in Sellers et al. (1996), it is very doubtful that a first-order closure model [e.g., SiB2 or (4)] can describe transfer processes below a canopy in a credible way (also see Shaw and Pereira 1982). The formulation in SiB2 was validated over tropical forests (Sellers et al. 1989) but not over other vegetation types (e.g., shrub). Recognizing the poor understanding of undercanopy turbulence (e.g., McNaughton and Van den Hurk 1995), we have taken two different approaches to remove the ground temperature bias over semiarid regions in CCSM2 due to its deficiency in C_s.

In the first approach, C_s for any L_t is simply interpolated between the values for a thick canopy and bare soil [i.e., (3) and (5)]:

$$C_s = k/\left[\alpha\left(\frac{z_o}{\upsilon}\right)^{0.45}\right]u_*^{-0.45}W + 0.004(1 - W), \quad (6)$$

where the fractional weight W is

$$W = \exp(-L_t). \quad (7)$$

Equations (6) and (7) satisfy two conditions: at $L_t = 0$, $W = 1$ and C_s follows the bare soil formulation; and as L_t becomes sufficiently large (e.g., $L_t \geq 5$), W approaches zero and C_s approaches the thick-canopy value. Sensitivity tests in section 3 will further show that the results are insensitive to the functional form of W as long as the above two conditions are met.

An alternative C_s formulation is motivated by (4). First, two measures of the inverse of the reduction of turbulence by the canopy are defined as

$$r_{t1} = \frac{h}{d(\beta + 0.1)}\left[1 - \exp\left(-\frac{\beta d}{h}\right)\right]e^{\beta} \quad \text{and} \quad (8)$$

$$r_{t2} = \frac{h}{d\beta}\left[1 - \exp\left(-\frac{\beta d}{h}\right)\right]e^{\beta}, \quad (9)$$

where $\beta = 0.7L_t$. The small value of 0.1 incremented to β in the denominator of (8) ensures that $r_{t1} = 0$ at $L_t = 0$. In contrast, $r_{t2} = 1$ at $L_t = 0$.

We can then define undercanopy aerodynamic and sublayer nondimensional resistances as

$$r_1 = \frac{d}{k(h - d)}r_{t1} \quad \text{and} \quad (10)$$

$$r_2 = r_{t2}^{0.45} \times \ln(z_o/z_{oh})/k, \quad (11)$$

where z_o and z_{oh} are the soil roughness lengths for momentum and heat, respectively, and $\ln(z_o/z_{oh})$ is equal to the denominator in (5). Then C_s for any L_t can be computed as

$$C_s = 1/(r_1 + r_2), \quad (12)$$

which is a direct generalization of (4).

Theoretically, C_f in (2) should vary with L_t as well, and a formulation similar to (6) could be used. This variation is not as important as that of C_s, however, because H_f in (2) is proportional to C_fL_t, and is always zero at $L_t = 0$ as long as C_f is finite. Therefore, C_f is kept as constant (0.01) for simplicity.

3. Preliminary evaluation of the new formulations

We first compare the new formulations (6) and (12) in Fig. 1. For convenience, it is assumed that $\ln(z_o/z_{oh}) = \alpha[(u_*z_o/\upsilon)]^{0.45} = 2$ and $d/h = 2/3$. As L_t approaches zero, both C_s formulations correctly approach the value over bare soil as given in (5). They also essentially converge to the same value at $L_t = 7$. The C_s values are different for intermediate L_t with C_s from (6) higher for $L_t < 1$ and C_s from (12) higher for $L_t > 1$. Qualitatively, as long as these C_s values multiplied by u_* (i.e., the undercanopy conductance) are smaller than the canopy conductance (e.g., for relatively thick canopies), the impact of the C_s difference between (6) and (12) would be small.

Next, we compare these two formulations with (4). For $b = 3$ (Bonan 1996) and $b = 2.5$ (Lo Seen et al. 1997), $C_s = 0.031$ and 0.043, respectively. Figure 1 shows that these C_s values correspond to L_ts of 2.0 and 1.6 in (6). As mentioned earlier, (4) with a constant

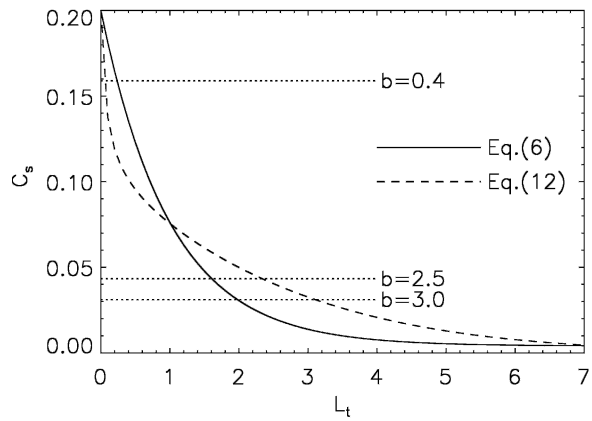

FIG. 1. Surface turbulent exchange coefficient C_s vs the sum of leaf, stem, and dead leaf area indexes (L_t) using (6) (solid line) and (12) (dashed line). Results using (4) with $b = 0.4$, 2.5, and 3.0 are also given.

coefficient b does not converge to the bare soil formulation as canopy disappears. Qualitatively, (4) could also be consistent with (6) if the coefficient b is allowed to vary with L_t. In fact, previous studies have demonstrated that b varies from 0.4–0.8 for thin canopies and to 2–4 for thick canopies (Brutsaert 1982). Taking $b = 0.4$ for a thin canopy, C_s is 0.16, which is fairly close to 0.2 from (6) (Fig. 1). A more rigorous convergence, of course, is provided by (12) through modification of (4).

Using the same assumptions in the derivation of (4), the C_f formulation can also be obtained (Bonan 1996; Lo Seen et al. 1997):

$$C_f \propto [1 - \exp(-b/2)]/b. \qquad (13)$$

As mentioned earlier, the canopy conductance is proportional to $C_f L_t$. For the variation of b by a factor of 10 ($b = 0.4$ for a thin canopy with, e.g., $L_t = 0.4$, and $b = 4.0$ for a thick canopy with $L_t = 6$), C_f varies by only a factor of 2 in (13) in contrast to the variation of L_t by a factor of 15. The variation of C_f is also much smaller than that of C_s from (6) or (12). Therefore, C_f is kept as constant (0.01) for simplicity, as mentioned in section 2.

While many land models that consider undercanopy turbulence have been evaluated using in situ observational data (e.g., Lo Seen et al. 1997), few observational data are available to directly evaluate different C_s formulations. The undercanopy conductance ($C_s u_*$) was measured using source plates beneath a maize canopy in Sauer et al. (1995), and was found to vary from 0.002 to 0.03 m s^{-1}. This range would be consistent with the range of our C_s from 0.004 to 0.2 in (6) or (12) (see Fig. 1), if we (reasonably) assume that u_*, denoting the friction velocity above the canopy to roughly represent the wind velocity incident on the leaves in (1), varies from 0.1 to 0.5. Sauer et al. (1995) also provided the results for a particular day with leaf area index being 2.46 and canopy height being 2.65 m (their Table 3): the undercanopy conductance was about 0.006 m s^{-1} (their Table 4). Using the wind speed and sensible heat flux measurements above the canopy from these tables, we can estimate u_* to be around 0.4 m s^{-1}. Then the average C_s value is about 0.015, which is fairly close to 0.02 from (6) at $L_t = 2.5$ but is much lower than those using a constant b or from (12) (see Fig. 1).

To assess the impact of different C_s formulations on the computation of ground temperature and surface fluxes, we have run the CLM2 using observational near-surface atmospheric data over a station in Arizona (32.28°N, 110.95°W) for the month of June 2002 when the monthly precipitation was zero (information available online at http://ag.arizona.edu/azmet/01.htm). The initial soil moisture at 1 June 2002 was set as the wilting point value, because there was no rain during the month of May 2002. The initial bottom layer soil temperature was set as the annual mean soil temperature measured at 0.1-m depth, and soil temperatures at other layers were interpolated (or extrapolated) using the measured soil temperatures at 0.1- and 0.5-m depths and the estimated bottom layer soil temperature. All model parameters are the same as those used for shrubland over this area in the CCSM2 whose land component is CLM2 (including $L_t = 0.1$). The first 20 days of CLM2 simulations are used to spinup ground temperature, surface fluxes, and soil temperature in the top soil layers.

First we artificially prescribe L_t from 0 (bare soil) to 7, and run the CLM2 with each L_t using the above atmospheric forcing data. The ground temperature T_g averaged over the last 10 days as a function of L_t is given in Fig. 2. In general, T_g should not change much when L_t is increased from zero (i.e., bare soil) to a very small value (e.g., $L_t = 0.1$). However, Fig. 2 shows that the use of $C_s = 0.004$ in the control run results in a substantial jump of T_g by 8 K from bare soil to $L_t = 0.1$, demonstrating that indeed the C_s formulation and small L_t in CLM2 are primarily responsible for the excessive warm bias in T_g. The use of $C_s = 0.031$ and 0.043 results in a jump of T_g by 2.2 and 1.3 K, respectively. In contrast, (6) or (12) results in a reasonable (and statistically insignificant at the 95% level) variation of T_g within 0.5 K between $L_t = 0$ and 0.1, partly due to the different albedos of bare soil versus shrubs. Differences between (6) and (12) are within 1 K (and statistically insignificant) for all L_t values. They are also close to those in the control run for $L_t \geq 3$. Results are relatively close using (6) or (12) versus $C_s = 0.031$ or 0.043 for intermediate L_t values (e.g., $L_t = 1$ to 4), but differ signifi-

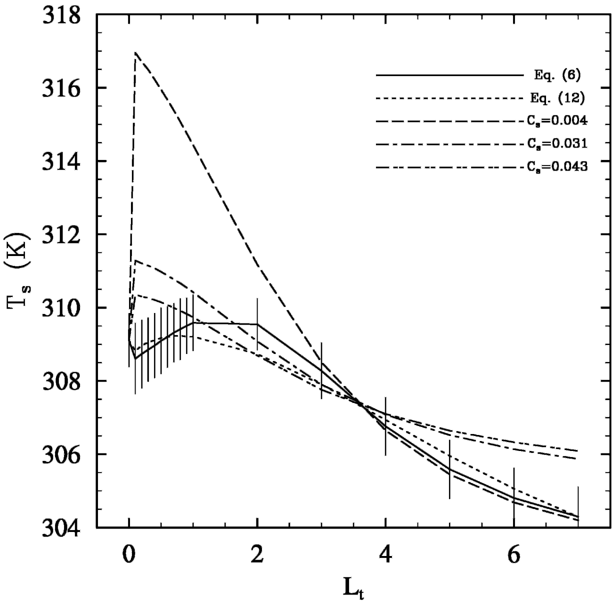

FIG. 2. The CLM2-simulated ground temperature (T_g) averaged over the last 10 days of Jun as a function of prescribed L_t from 0 (bare soil) to 7 using observed atmospheric forcing data in AZ. Vertical lines indicate two standard deviations from the 10-day mean using (6) and hence represent its 95% confidence interval. Results using (6), (12), and $C_s = 0.004$ (control simulation), 0.031 [from (4) with $b = 3.0$], and 0.043 [from (4) with $b = 2.5$] are shown.

cantly at both small (e.g., $L_t = 0.1$) and large L_t values (e.g., $L_t = 7$).

Because the exponential function in (7) is empirical, the sensitivity of the results to the exact functional form of W needs to be assessed. We have redone computations in Fig. 2 with four different W formulations that also satisfy the two conditions given in section 2:

$$W = \exp(-1.3L_t), \tag{14}$$

$$W = \exp(-0.7L_t), \tag{15}$$

$$W = 1 - \frac{\min(L_t, 5)}{5}, \quad \text{and} \tag{16}$$

$$W = 1 - \sin\left[\frac{\min(L_t, 5)}{5}\frac{\pi}{2}\right]. \tag{17}$$

Similar to (7), these W formulations result in a reasonable variation of T_g within 0.5 K between $L_t = 0$ and $L_t = 0.1$ (figure not shown). The difference of T_g using these W formulations versus (7) is statistically insignificant for all L_t values [for (14) and (15)] or for most L_t values [for (16) and (17)]. In other words, results are insensitive to the exact functional form of W and hence the simple (7) is used.

Figure 3 compares the diurnal cycles of surface variables averaged over the last 10 days with $L_t = 0.1$ as used in CCSM2 over this area. Equations (6) and (12) give nearly the same results, and, compared with $C_s = 0.004$ (control run), they reduce the averaged daily maximum ground temperature (T_g) by as much as 19 K. These T_g differences also propagate downward in the soil, and result in soil temperature differences of 7 K at 0.21-m depth. The reduction of the daily maximum T_g by (6) or (12) also leads to the reduction of net longwave radiation (LW_n) and heat flux into the soil (G) by 140 and 100 W m^{-2}, respectively. Since the net solar radiation is the same and the latent heat flux is close to zero, the reduction of LW_n and G is compensated by the increase of sensible heat flux (SH) by as much as 240 W m^{-2}. Compared with the results averaged over the last 10 days from the control simulation, (6) or (12) reduces T_g by 8 K, soil temperature at 0.21-m depth by 7 K, LW_n by 55 W m^{-2}, and G by 6 W m^{-2}. Correspondingly, (6) or (12) increases SH by 61 W m^{-2}.

Figure 3 also shows that results using (6) or (12) versus those in the control simulation are statistically significant at the 95% level based on the Student's t test except for the early morning hours. Qualitatively, daytime SH in the control simulation (Fig. 3b) is too small (less than 100 W m^{-2}), and daytime G (Fig. 3a) is too large. The 30-K difference between the early afternoon T_g in the control simulation (Fig. 3c) and the observed 2-m air temperature (Fig. 3g) is also too large. Their difference of 6 K in the early morning (when the atmosphere is nearly neutral) is also unreasonable. Compared with results using (6) or (12), the use of $C_s = 0.031$ or 0.043 yields higher T_g, LW_n, and G (and hence lower SH) during the day, and yields similar results at night. Soil temperature at 0.21-m depth (Fig. 3d) using $C_s = 0.031$ or 0.043 is lower than that in the control simulation by 4 K, but is still higher by 2–3 K than (6) or (12).

Since observational ground temperature data are unavailable at this site, we have also chosen a tiger bush site in West Africa (13.20°N, 2.24°E) to further evaluate (6) versus $C_s = 0.004$. The atmospheric forcing data and observed T_g data are available (online at http://www.ird.fr/hapex/) for the period of 19 August–7 October 1992 during the Hydrology–Atmosphere Pilot Experiment in the Sahel, 1990–1992 (HAPEX-Sahel; Prince et al. 1995). All CLM2 parameters are the same as those used for the shrub tile over this area in the CCSM2 (including $L_t = 0.4$). Figure 4 compares the daily averaged T_g in the control run and the run using (6) with observations for the month of September 1992. The use of $C_s = 0.004$ produces excessive ground temperatures by as much as 15 K in the last 2 weeks of the simulation. When (6) is implemented, bias nearly dis-

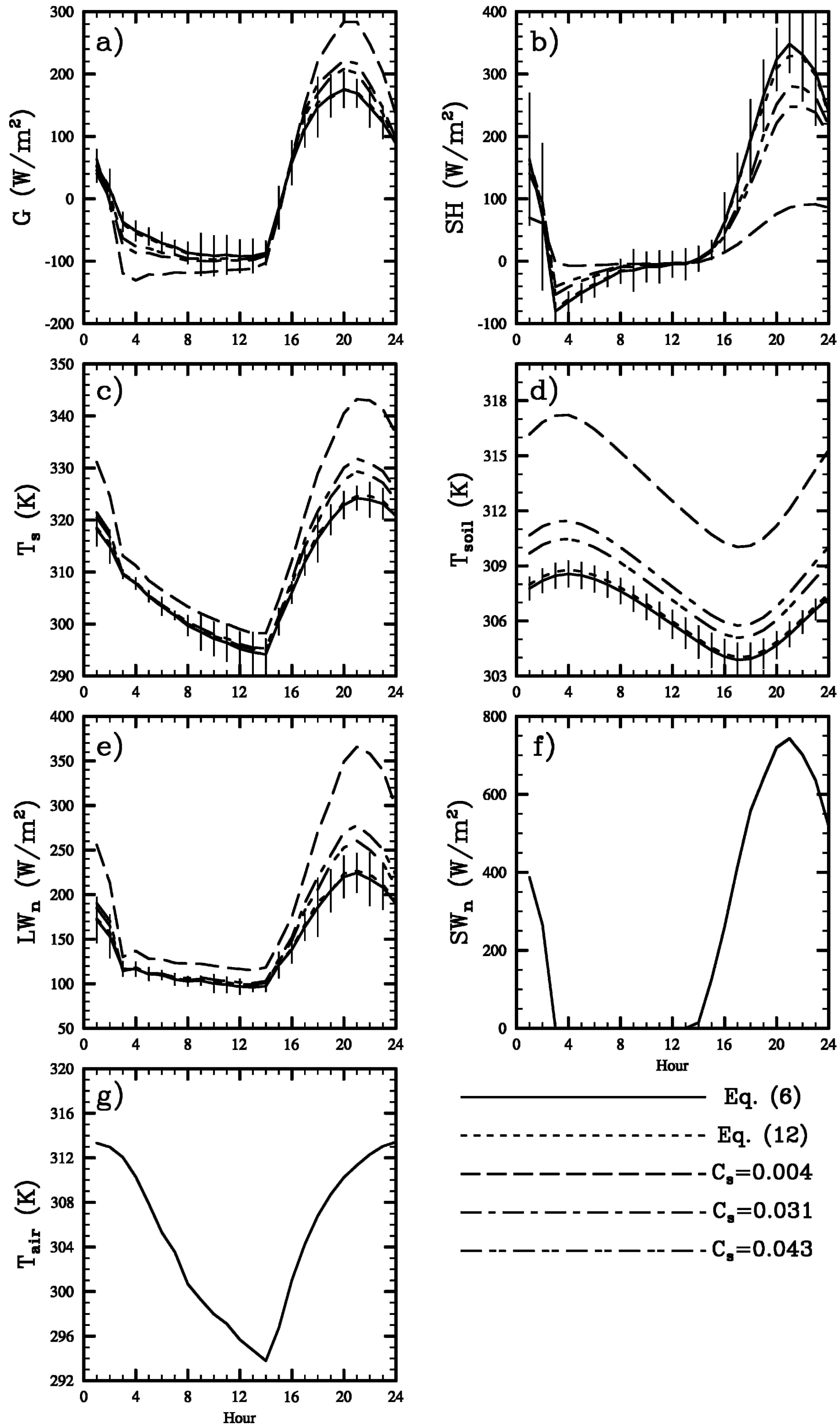

FIG. 3. The diurnal cycles averaged over the last 10 days of Jun of (a) CLM2-simulated heat flux into the soil (G), (b) surface sensible heat flux (SH), (c) ground temperature (T_g), (d) soil temperature at 0.21-m depth (T_{soil}), (e) net longwave radiation (LW_n), (f) net solar radiation (SW_n), and (g) observed 2-m air temperature (T_{air}). The observed atmospheric data in AZ are used to drive CLM2. Results using (6), (12), and $C_s = 0.004$ (control simulation), 0.031 [from (4) with $b = 3.0$], and 0.043 [from (4) with $b = 2.5$] are shown. Vertical lines indicate two standard deviations from the 10-day mean using (6) and hence represent its 95% confidence interval. Standard deviations are not needed for observed T_{air} or SW_n (which is not affected by the use of different C_s formulations).

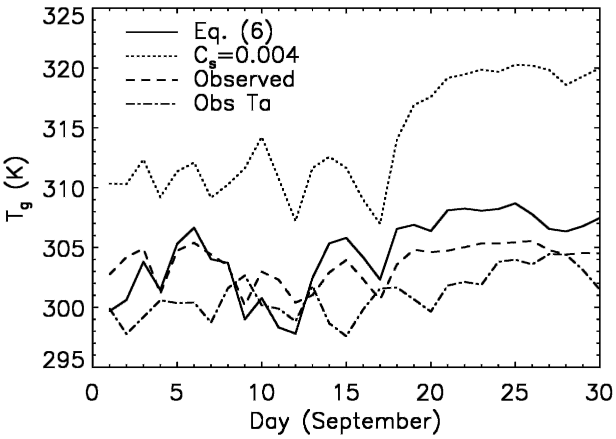

FIG. 4. Comparison of CLM2-simulated daily mean ground temperature (T_g) using (6) and $C_s = 0.004$ (control simulation) with observational data in Sep 1992 during the HAPEX-Sahel experiment. Observed surface air temperature (T_a) is also shown.

appears in the beginning of the month and is substantially reduced in the last 2 weeks. Therefore, as expected, (6) significantly improves the results compared with the control simulation.

Equation (6) is simpler but more empirical, while (12) is more physically based. Mathematically, however, both satisfy the two conditions in section 2 and give similar results. Therefore, the simpler (6) has been chosen for implementation in the newer version of CCSM (i.e., CCSM3). Just as in the above offline simulations, (6) is found to substantially reduce the ground temperature bias over arid and semiarid regions in the global land–atmosphere coupled simulations and in the climate system modeling (CCSM3) (Dickinson et al. 2006).

4. Conclusions and further discussion

The NCAR Community Climate System Model (CCSM2) shows an excessive warm bias of around 10 K in monthly mean ground temperature over arid and semiarid regions. Sensitivity studies show that this is primarily caused by the prescription of small L_t (i.e., the sum of leaf, stem, and dead leaf area indexes) (Barlage and Zeng 2004) and the use of a constant surface exchange coefficient C_s that is adequate for thick canopies only. A similar problem may also exist in other land models that consider the radiative transfer through canopy and use a constant C_s. Two different formulations [i.e., (6) and (12)] have been developed for the computation of C_s for any L_t values. They approach the same C_s values at the extreme cases of bare soil (i.e., $L_t = 0$) and thick canopy with $L_t = 7$, but differ for intermediate L_t values. Sensitivity tests show that (6) or (12) leads to similar ground temperature T_g, and significantly improves the T_g simulation compared with the control run (with $C_s = 0.004$). The use of $C_s = 0.031$ or 0.043 also improves the T_g simulation compared with the control run, but it still leads to a significant increase in T_g from $L_t = 0$ to $L_t = 0.1$. Because of the results presented here, (6) has been implemented in the most recent version of CCSM (i.e., CCSM3), and is found to substantially reduce the T_g bias over arid and semiarid regions (Dickinson et al. 2006). Furthermore, while the C_s values for bare soil and thick canopy are model dependent, our approach of obtaining C_s for any L_t as an interpolation between the values for a thick canopy and bare soil is probably applicable to all land models that consider canopy and its underlying soil separately.

So far the ratios of d/h and z_{oc}/h in (4) and (12) have been taken as constant, corresponding to their values for thick canopies. In reality, both vary with L_t (Lindroth 1993; Shaw and Pereira 1982) and frontal area index (Raupach 1994), which is related to fractional vegetation cover (Zeng et al. 2000) and the ratio of canopy thickness versus width (Schaudt and Dickinson 2000). In particular, z_{oc}/h is not monotonic with L_t; rather, it reaches its peak for an intermediate L_t (e.g., Sellers et al. 1989). While C_s in (6) is not directly affected by these two ratios, the undercanopy conductance ($C_s u_*$) is still dependent on them through their impact on the computation of above-canopy friction velocity u_*. As canopy disappears (i.e., as $L_t \to 0$), z_{oc} and d should approach their bare soil values, which are much smaller than the canopy height h, so that d/h and z_{oc}/h become very small. This does not affect the asymptotic behavior of the C_s formulation in (12), because, at $L_t = 0$, $r_{t1} \equiv 0$ in (8) and $r_{t2} \equiv 1$ in (9) no matter whether d/h approaches zero or a finite value. For the C_s formulation in (4), however, if d/h and z_{oc}/h are assumed to become very small as $L_t \to 0$, then

$$C_s \sim ke^{-b}/(d/h). \tag{18}$$

Qualitatively, the decrease of d/h and z_{oc}/h with the decrease of L_t would lead to the increase of C_s from (4), in better agreement with (6). Quantitatively, however, C_s from (18) does not converge to the bare soil value in (5). For instance, taking $b = 3$ and $d/h = 0.01$, C_s from (18) would be larger than the bare soil value by an order of magnitude.

Besides (6) or (12), another formulation may also be developed through revision of the more complicated first-order closure model of undercanopy turbulence (Sellers et al. 1996). It is not pursued here largely because it is not analytical and because the simple (6) or (12) already largely removes the excessive warming of the ground over arid and semiarid regions, and is roughly consistent with limited observational data of Sauer et al. (1995). A focused field program is needed to provide the comprehensive observational data for the evaluation and further improvement of (6) or (12). In particular, the functional form of W and the 0.004 factor in (6) could be better calibrated with appropriate data. Multilayer canopy models (e.g., Pyles et al. 2003, and references therein) will also be useful for the further improvement of (6) or (12).

Figures 1 and 2 suggest that measurements need to be made at both small (e.g., $L_t = 0.1$) and large L_t values (e.g., $L_t = 7$). The design of such an experiment also needs to pay attention to the treatments of horizontal heterogeneity, atmospheric convective turbulence, and radiative transfer through canopy, because they are different in various land models and all directly affect the C_s formulation. For instance, in the two-source model (i.e., computing the canopy and ground temperatures separately) of Kustas and Norman (1999), a single ground temperature is used for bare soil and undercanopy soil, and hence their surface resistance [i.e., $1/(C_s u_*)$] formulation may not be appropriate for a land model that considers shaded and unshaded soils separately (e.g., CLM2). Similarly, the convective gustiness due to unstable atmospheric boundary layer large eddies is directly considered in the surface resistance formulation in Kustas and Norman (1999), while this convective gustiness is directly considered in the computation of scalar wind in CLM2, which in turn affects u_* and hence surface resistance. Evidently, scientists in field measurements, remote sensing, and land surface modeling need to work together to develop such a program.

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