

A mechanism for the influence of vegetation on the response of the diurnal temperature range to changing climate

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Abstract. We propose a new mechanism that could contribute to the observed decrease in the diurnal temperature range (*DTR*) over the last century: the physiological behavior of vegetation in response to climate. Using a physiologically based land surface model, we analyze the influence of vegetation on the response of the *DTR* to perturbations in the state of the climate and vegetation. Increasing down-welling long wave radiation and surface air temperature together, conditions that could occur as a result of doubling of atmospheric CO₂, produced little change in the *DTR*. Changes in the state of the vegetation (i.e. amount, physiological capacity, stress) produce changes in the *DTR* of the order or larger than observed. Results emphasize that *DTR* modeling studies need to consider vegetation responses and suggest that recently reported increases in vegetation over the last decade could contribute to the observed decreases in the *DTR*.

1. Introduction

Karl and colleagues [e.g. Karl *et al.*, 1984; Easterling *et al.*, 1997] reported that during the period 1951 to 1993, the diurnal temperature range (*DTR*) on land has generally decreased as mean surface temperatures have increased. The decrease in *DTR* results mostly from larger increases in minimum relative to maximum temperatures.

A number of mechanisms have been proposed that could cause the *DTR* to decrease as the mean temperatures increase. These include changes in cloud cover, atmospheric water vapor, tropospheric aerosols, atmospheric turbulence and soil moisture. Increases in the first three could lead to reductions in the incoming surface short wave radiation during the day, and to increases in the down-welling surface long wave radiation at night. Atmospheric turbulence and soil moisture can affect the heat and water vapor fluxes from the surface, which are generally much larger during the day than at night. These mechanisms have been evaluated in several modeling studies [e.g. Cao *et al.*, 1992; Verdecchia *et al.*, 1994; Mearns *et al.*, 1995]. Hansen *et al.*, [1995] examined the role of CO₂, aerosols, and clouds on the *DTR* and mean temperatures, and argued that in climate simulations, forcings that increase the mean temperature generally do not affect the *DTR* to the extent that is observed. However, we are aware of no study that has considered the role vegetation plays in determining trends in the *DTR*.

The highly interactive nature of general circulation models (GCMs) as well as weaknesses in their parameterizations make it difficult to identify the mechanisms underlying predicted changes in the *DTR*. The purpose of this study is to examine how the diurnal temperature cycle of vegetated land surfaces responds to changes in external forcing and the biophysical state of the vegetation. We use the SiB2 land surface model [Sellers *et al.*, 1996a] in an off-line mode with prescribed meteorology for a number of scenarios highlighting the impact of vegetation on *DTR*. Off-line simulations do not account for feedback between the surface fluxes and the driving variables, and as such allow examination of how the vegetation component of the climate system responds to given climate forcing. Some GCM results are also presented.

2. MODEL

2.1 Land Surface Model

SiB2 (Simple Biosphere Model Version 2) land surface model couples energy, momentum, CO₂ and water fluxes in a consistent way for use in GCMs, as a replacement for and an advancement over the "bucket-type" models [Sellers *et al.*, 1997]. SiB2 includes parameterizations of canopy physiological responses (photosynthesis, stomatal conductance) and was designed to utilize satellite measurements for many of the important vegetation boundary conditions such as fraction of short wave radiation absorbed, albedo and roughness.

In our GCM simulations (CSU GCM, see Randall *et al.* [1996]) transpiration from vegetated surfaces generally accounts for up to 60% of the land surface latent heat flux (λE) over the year, the rest coming directly from the soil and from evaporation of precipitation intercepted by the canopy and soil. Physiological control of transpiration arises from the response of canopy conductance (g_c) to environmental conditions. The expression describing g_c in SiB2 is relatively simple, but it is based on physiological and ecological principles and is consistent with observations [Collatz *et al.*, 1991, Collatz *et al.*, 1992, Sellers *et al.*, 1992]:

$$g_c = \frac{m h_s A(T_c, c_i, PAR, V_m, W) FPAR}{c_s \bar{k}} + b(W) LAI, \quad (1)$$

where A is leaf net photosynthesis, which is a function of canopy surface temperature (T_c), leaf internal CO₂ partial pressure (c_i), photosynthetically active solar radiation (PAR), nutrition (V_m), and soil water stress (W); h_s and c_s are the surface relative humidity and CO₂ partial pressure respectively; m is a parameter expressing the overall sensitivity of g_c to the rest of variables in this term of the equation and b is the minimum stomatal conductance approached when photosynthesis is not active, e.g. at night and during dormant periods. $FPAR$ is the fraction of incident PAR absorbed by the canopy, \bar{k} is the mean light extinction coefficient of the canopy [Sellers *et al.*, 1992], and LAI is the leaf

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area index. The parameters m , b and V_m are vegetation type-dependent and time-invariant; in contrast, $FPAR$, \bar{k} , and LAI vary seasonally, largely independent of vegetation type. $FPAR$ and LAI can be obtained from satellite data. T_c , c_i , h_s , c_s , and W are state variables calculated by the model and must be solved for numerically. Equation (1) illustrates that g_c is highly sensitive to both prescribed parameters and to the state of the climate.

SiB2 has been tested against observations and matches reasonably well with measured fluxes for a number of diverse vegetation types [e.g. *Colello et al.*, 1997, *Denning et al.*, 1996, *Rocha et al.*, 1996]. Other land surface models of this type produce similar responses to meteorological forcing, as long as soil moisture is not limiting [*Koster and Milly*, 1997]

2.2 DATA

Observations show that northern high latitude regions have generally experienced large decreases (3-5 K) in DTR between 1950 and 1993 [*Easterling et al.*, 1997] so we choose for our off-line analysis a boreal region in north-central Canada. The meteorological data were collected by others at a Black Spruce forest site near Thompson, Manitoba as part of NASA's BOREAS Program.

The input data include short wave incoming radiation (SW), long wave incoming radiation (LWD), temperature (T_m) water vapor pressure (e_m), wind speed and precipitation measured above the canopy. A period of 11 rainless days in July were selected. Peak midday SW ranged from 450-700 $W m^{-2}$, T_m ranged from 282 to 302 K, e_m from 0.7 to 2.2 kPa and wind speed from near 0 to 6 $m s^{-1}$. Soil moisture was not limiting. The vegetation parameters used in the control simulations are given in *Sellers et al.*, [1996b] for evergreen needle leaf forest. $FPAR$ was derived from satellite measurements [*Sellers et al.*, 1996b] for July in north-central Canada and corresponded to a leaf area index of 2 consistent with the average reported for spruce and pine stands in this region [*Dang et al.*, 1997].

A number of simulations were run in which we altered the forcing meteorology, model biophysics, or model parameters, in order to investigate the sensitivity of the DTR to the landsurface parameterization. We also include a brief comparison with results from coupled SiB2 - GCM climate simulations.

3. RESULTS AND DISCUSSION

Off-line SiB2 simulations using our standard parameter data set for boreal forests and observed meteorological conditions produced peak λE and sensible heat flux (H) at midday of about 300 $W m^{-2}$ and peak gross photosynthetic rates of about 11 $\mu mol m^{-2} s^{-1}$ comparable to values reported for forests of Spruce and Pine during BOREAS [*Pattey et al.*, 1997, *Jarvis et al.*, 1997, *Joiner et al.*, 1999].

3.1 Mean Temperature versus DTR Change

Hansen et al., [1995] reported that increasing radiative forcing in GCM simulations always increases mean temperatures but generally causes weak responses in the DTR . They defined a measure of the change in the DTR relative to the change in the mean as:

$${}^n\Delta DTR = \frac{0.5\Delta DTR}{|\Delta T_a|}, \quad (2)$$

Their study reported observed and simulated surface air temperatures, while here we utilize the canopy air space

Table 1. Temperature differences from control (shown in parentheses). Differences in T_a , canopy surface and surface air temperatures are shown for the off-line runs, CSU GCM and *Hansen et al.*, [1995] respectively. Asterisks following ${}^n\Delta DTR$ values indicate magnitudes comparable to observations.

	ΔT_{can}	ΔDTR	${}^n\Delta DTR$
SiB2 Off-line			
$T_m + LWD$	2.9(297.0)	-0.2(14.2)	-0.03
T_m	2.3	0.9	0.19
LWD	0.5	-1.0	-0.94*
Fixed g_c	-0.3	0.4	0.57*
$T_m + LWD + g_c (2XCO_2)$	3.0	0.0	0.00
1/2 V_{max}	0.2	0.4	0.97*
low $FPAR$	0.3	0.5	0.71*
CSU GCM			
2 X CO_2	2.6	-0.3	-0.05
Low Stress	-0.8	-0.7	-0.43*
<i>Hansen et al.</i> , (1995)			
2 X CO_2	2.8	-0.4	-0.08
2 X $CO_2 + SW(+2\%)$	3.0	-0.3	-0.05
Observed	0.5	-0.5	-0.5

temperature (T_a) [see *Sellers et al.*, 1996a]. For the observed increase in mean surface air temperatures of 0.5 K and decrease in the DTR of 0.5 the ${}^n\Delta DTR$ equals 0.5. *Hansen et al.*, [1995] argued that the most plausible mechanism for a decrease in the DTR under radiatively forced conditions would be an increase in low clouds over land. Climate simulations reported by *Sellers et al.*, [1996c], using the CSU GCM, which includes SiB2, revealed similar results; that radiative forcing by itself does not produce a big enough decrease in the DTR relative to the mean temperature to match observations (Table 1). In Table 1 we note which of the forcings we describe below produce changes in the ${}^n\Delta DTR$ of a magnitude comparable to or larger than observations. The calculations of ${}^n\Delta DTR$ in Table 1 are based on the differences between the scenario simulations and the control. The sign of ${}^n\Delta DTR$ depends on the direction of the particular perturbation such that a change in the opposite direction would likely produce similar responses in ${}^n\Delta DTR$ but of opposite sign.

We ran a 10-year integration, using the CSU GCM, in which we reduced physiological stresses (water and temperature) that occurred because of incompatibilities between the simulated climate and the prescribed vegetation parameters. This reduction in stress was accomplished largely by increasing the high temperature stress parameter [*Sellers et al.*, 1996a]. Though both the annual mean temperature and the DTR over land decrease there is a large decrease in ${}^n\Delta DTR$ (Table 1). The reduction in stress causes increases in λE , total atmospheric water vapor content, total cloudiness and precipitation. We conclude that the land surface parameterization in GCMs plays an important role in the diurnal temperature cycle, and must be adequately addressed in modeling studies of the DTR . The following analyses of off-line simulations allow us to examine the direct response of the land surface model to changes in climate forcing and in the parameterization of the vegetation.

3.2 Climate Forcing

Three simulations were carried out in which the meteorological driver data were altered to mimic plausible changes in climate forcing caused by a doubling of atmospheric CO_2 partial pressure. In the first case the air temperature above the canopy (T_m) was increased by 3.2 K. In the second case, long wave down-welling radiation (LWD) was increased by 23 $W m^{-2}$

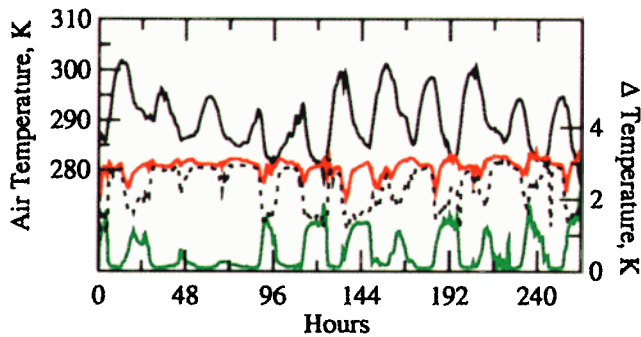


Figure 1. Canopy air space temperature for the control (solid black) and differences from control for an increase in T_m (dotted black), LWD (green) and increases in both T_m and LWD (red).

and both the T_m and LWD modifications were imposed together in the third case. These perturbations were derived from results of $2XCO_2$ GCM simulations reported in Sellers *et al.*, [1996c], and were applied at each time step (1/2 hour). Above canopy water vapor pressure was increased with T_m to maintain the relative humidity at the observed values. Doubling of atmospheric CO_2 can cause stomata to close and, therefore, reduce g_c (Equation (1), Sellers *et al.*, [1996c]), but for these runs this response was not invoked. In the next section, variable stomatal responses will be addressed.

Increasing T_m raises the daytime maximum T_a by about 2.8 K, while increasing LWD has virtually no effect at midday (Figure 1). Midday T_a is warmer than the air above, so increasing T_m reduces the temperature gradient resulting in decreased H . Since incoming radiation does not change, the decrease in H is mostly compensated for by an increase in λE (caused by the increased water vapor gradient between the surface and the warmer canopy surface) and by small increases in upwelling long wave radiation. Aerodynamic resistances are typically low during the day (< 10 s/m) so the surface is closely coupled to the air above it.

At night, however, T_m and LWD forcing contributes about equally to the change in T_a for nights with lower wind speeds (e.g. Figure 1, ~192 hours). At night the temperature gradient is reversed from that of the day so, in contrast to midday, by increasing T_m the temperature gradient is increased. Aerodynamic resistances are larger at night (> 100 s/m) causing the surface to be more uncoupled from the air above it and increasing T_m further increases atmospheric stability. The 23 Wm^{-2} increase in LWD represents a ~25% change in net radiation at night versus ~5% change during midday.

These responses are summarized in the comparison of ${}^{\circ}\Delta DTR$ (Equation 2) shown in Table 1. The impact of LWD forcing alone produces a significant change in the DTR relative to the mean temperature while increasing T_m or the combination of increasing T_m and LWD have only small impacts on ${}^{\circ}\Delta DTR$.

3.2 Surface Conductance Parameterization

A pair of simulations were run in which g_c was *i*) allowed to decrease in response to a doubling of atmospheric CO_2 as reported by Sellers *et al.*, [1996c] (Equation (1)) resulting in ~25% decrease in midday g_c , and *ii*) held constant over the course of the diurnal cycle as is the case for so called bucket models which do not include a diurnal physiological response [see Sellers *et al.*, 1997]. In the former case, λE decreases by as much as 10% at midday compared to the control, but canopy air

temperatures increases only by about 0.1 K because of the strong coupling between the surface and air above causing increased H to compensate for the lower λE . In the latter case in order to mimic the bucket model approach, g_c was fixed for the whole diurnal cycle at a value to match midday λE in the control. At night, the fixed g_c scenario causes excessive λE by as much as 50 $W m^{-2}$, reducing temperatures by up to 2 K and thus increasing the DTR relative to the control (Table 1). g_c response to high CO_2 combined with increased T_m and LWD has virtually no effect on the ${}^{\circ}\Delta DTR$.

3.3 Vegetation Parameters

The two most influential vegetation parameters in SiB2 are V_m , which specifies the maximum capacity of the canopy for photosynthesis and, therefore, the maximum g_c capacity; and $FPAR$, which represents the fraction of incident PAR absorbed by the canopy. V_m is analogous to the maximum g_c parameter used in other biophysical land surface models, and is reduced by water stress and extreme temperature stress. $FPAR$ is mostly a function of the amount of green leaves (LAI), solar illumination angles and canopy structure. We examine the sensitivity of the DTR to the value of the prescribed maximum capacity. Effects of prescribed $FPAR$ are also examined.

Reducing V_m by 50% causes a 30% reduction in g_c at midday, and a decrease in peak λE by about 25%. T_a at midday increases by up to 0.5 K. Strong midday stomatal closure occurs in the low V_m simulation indicating a positive feedback between stomatal closure, surface humidity and T_a (Equation (1), Collatz *et al.*, 1991). The warming occurs only during the day, causing the DTR to increase (Table 1). The large change in the DTR relative to the mean air temperature causes the ${}^{\circ}\Delta DTR$ to significantly increase. This implies that the prescribed value of V_m or conductance capacity can significantly affect ${}^{\circ}\Delta DTR$. Water and temperature stress would have the effect of reducing V_m or conductance capacity thus potentially increasing the DTR .

Reducing $FPAR$ by 50% causes peak g_c values to decrease by 33%, and reduces peak λE by over 25% from the control. This leads to an increase in T_a at midday of about 0.5 K, similar to the low V_m simulation. The sensitivity of ${}^{\circ}\Delta DTR$ to these parameters illustrates the importance of model boundary conditions in determining the response of the DTR .

The same analyses described here were performed for a tropical forest site and produced similar results (not shown).

4. Conclusions

The following general conclusions can be drawn from these results:

- An increase in LWD raises nighttime temperatures thus reducing the DTR while changes in T_m or $T_m + LWD$ increase both the maxima and minima with only small effects on the DTR . This response is largely driven by diurnal variability in aerodynamic stability and net radiation.
- Maintaining g_c constant over the diurnal cycle as in the case of bucket type models causes the DTR to be larger than when more realistic, diurnally varying g_c responses are used.
- All physiological forcings (g_c , V_m and $FPAR$) produce changes in the DTR that are large relative to changes in the mean temperatures. This implies that errors in parameterizations of landsurface models will introduce errors in predicted ${}^{\circ}\Delta DTR$, a conclusion also supported by our GCM simulation in which we reduced physiological stress. These results

suggest that reported increases in vegetation cover in the Northern Hemisphere during the 1980's [Myneni *et al.*, 1997] could have contributed to the lowered the *DTR* by increasing λE during the day, decreasing stability at night and through the influence of increased λE on atmospheric water vapor content and cloudiness.

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