# The Land Surface Climatology of the Community Land Model Coupled to the NCAR Community Climate Model

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ABSTRACT

The land surface parameterization used with the Community Climate Model (CCM3) and the Climate System Model (CSM1), the NCAR LSM (hereafter referred to as LSM1), has been modified as part of the development of the next version of these climate models. This new model is known as the Community Land Model (hereafter referred to as CLM2). In CLM2, the surface is represented by 5 primary sub-grid land cover types (glacier, lake, wetland, urban, vegetated) in each grid cell. The vegetated portion of a grid cell is further divided into patches of up to 4 of 16 plant functional types, each with its own leaf and stem area index and canopy height. The relative area of each sub-grid unit, plant functional type, and leaf area index are obtained from 1-km satellite data. The soil texture dataset allows vertical profiles of sand and clay. Most of the physical parameterizations in the model were also updated. Major model differences include: ten layers for soil temperature and soil water with explicit treatment of liquid water and ice; a multi-layer snow pack; runoff based on the TOPMODEL concept; new formulation of ground and vegetation fluxes; and vertical root profiles from a global synthesis of ecological studies. Simulations with CCM3 show significant improvements in surface air temperature, snow cover, and runoff for CLM2 compared to LSM1. CLM2 generally warms surface air temperature in all seasons compared to LSM1, reducing or eliminating many cold biases. Annual precipitation over land is reduced from 2.35 mm day$^{-1}$ in LSM1 to 2.14 mm day$^{-1}$ in CLM2. The hydrologic cycle is also different. Transpiration and ground evaporation are reduced. Leaves and stems intercept more water annually in CLM2 than LSM1, which gives rise to higher canopy evaporation. The annual cycle of runoff is greatly improved in CLM2, especially in arctic and boreal regions where the model has low runoff in cold seasons when the soil is frozen and high runoff during the snow melt season. Most of the differences between CLM2 and LSM1 are attributed to particular parameterizations rather than to different surface datasets. Important processes include: multi-layer snow, frozen water, interception, soil water limitation to latent heat, and higher aerodynamic resistances to heat exchange from ground.

## 1. Introduction

The NCAR Land Surface Model (NCAR LSM) is the land surface parameterization used with the Community Climate Model (CCM3) and the Climate System Model (CSM1). Since the documentation of this model by Bonan (1996a, 1998), the land biogeophysical parameterizations have been re-evaluated and changed as part of the development of the next version of the climate model. In particular, Zeng et al. (2002) developed a new biogeophysical parameterization called the Common Land Model. This model combines many of the features of the BATS (Dickinson et al. 1993), NCAR LSM (Bonan 1996a), and IAP94 (Dai and Zeng 1997) land models. It significantly reduces the cold summer surface air temperature bias in CCM3 and CSM1 by reducing latent heat flux and increasing sensible heat flux, improves the annual cycle of runoff, and better simulates snow mass (Zeng et al. 2002).

While the new biogeophysical parameterizations were being developed, NCAR LSM continued to be developed for carbon cycle and vegetation dynamics studies. NCAR LSM was originally developed to link the exchanges of energy, water, and $CO_2$ and was an outgrowth of earlier work with a similar model for boreal forests (Bonan 1991a,b,c, 1992, 1993a,b,c). Global simulations of NCAR LSM coupled to CCM3 showed that simple physiological and ecological assumptions result in reasonable simulation of land-atmosphere $CO_2$ exchange over a wide range of climates and ecosystems (Bonan 1995a, Craig et al. 1998). More recent work has focused on coupling with ecosystem and vegetation dynamics models. In particular, the model represents vegetation not as biomes (e.g., savanna) but rather as patches of plant functional types (e.g., grasses, trees). This is because many of the leaf physiological and plant allocation parameters used in ecological models can not be measured for biomes but can be measured for individual plant types. Plant functional types reduce the complexity of species diversity in ecological function to a few key plant types and provide a critical link to ecosystem processes and vegetation dynamics (Woodward and Cramer 1996; Smith et al. 1997). However, in NCAR LSM the types of plants in a grid cell and their abundance, leaf and stem area, and height are obtained by classifying the grid cell as one of 28 biomes. To better interface with ecological models and to take advantage of high resolution satellite data products, NCAR LSM was changed to allow plant type, abundance, leaf area, stem area, and height to be input to the model for each grid cell (Oleson and Bonan 2000; Bonan et al. 2002).

These developments in biogeophysics, carbon cycle, and vegetation dynamics have been merged into a new model of land surface processes for climate models – the Community Land Model (CLM2). This paper documents the effect of changes in model biogeophysics on the simulated climate. The carbon cycle and vegetation dynamics of the model will be described elsewhere. Three versions of the land model coupled to CCM3 are compared: LSM1 – the original NCAR LSM; LSM2 – an intermediate version of NCAR LSM that retains most of its biogeophysics, but includes new surface datasets and modifications for coupling to a dynamic global vegetation model; and CLM2 – the final model that merges the features of LSM2 with many of the biogeophysical parameterizations of the Common Land Model.

## 2. Methods

Simulations of 17-years length were performed with each of the three land models coupled to a version of CCM3. CCM3 is a spectral atmospheric model with T42 truncation (approximately 2.8° horizontal resolution), 18 vertical levels, and a 20-minute time step (Kiehl et al. 1996, 1998). Simulations used observed sea surface temperatures for the period September 1978-December 1995. The models were initialized with temperatures of 10°C, no snow or canopy water, and volumetric soil water content of 0.3 $mm^3$ $mm^{-3}$ over land. Lakes and wetlands were initialized to 4°C. Glaciers were initialized to -23°C and 1000 kg $m^{-2}$ of snow. Only the last 12 years of the simulations (i.e., for the period 1984-1995) were analyzed to allow a 5-year spin-up of soil water and temperature. The control simulation with LSM1 replicates the temperature and precipitation biases reported by Bonan (1998).

### *a. LSM1*

LSM1 is the NCAR LSM as described by Bonan (1996a, 1998). The model simulates the exchange of energy, water, momentum, and carbon between the surface and the atmosphere. Vegetation effects are included by allowing for 12 plant functional types (PFTs) that differ in plant physiology (leaf optical properties, stomatal physiology, leaf dimension) and vegetation structure (height, roughness length, displacement height, root profile, monthly leaf and stem area). Multiple PFTs can co-occur in a grid cell so that, for example, a mixed broadleaf deciduous and needleleaf evergreen forest consists of patches of broadleaf deciduous trees, needleleaf evergreen trees, and bare ground. Each patch, while co-occurring in a grid cell, is a separate column upon which energy, water, and carbon calculations are performed. Thus, plants do not compete for light and water. The abundance of PFTs in a grid cell is specified from one of 28 different biomes (Table 1). Lakes and wetlands, if present, form additional patches. Soil effects are included by allowing thermal and hydraulic properties to vary depending on sand and clay content. Soils also differ in color, which affects soil albedo. Required surface input data for each grid cell include a biome type (which determines the patch fractions for each PFT), the fraction of the grid cell covered by lakes, the fraction covered by wetlands, soil texture (percent sand, silt, and clay), and soil color.

Bonan (1996a) documents the model, and Bonan (1998) describes the climatology of the model coupled to the CCM3. Comparisons with tower flux data show the model reasonably simulates surface fluxes in several boreal forest (Bonan et al. 1997) and tundra (Lynch et al. 1999a) sites. The model has been used to study land-atmosphere $CO_2$ exchange (Bonan 1995a; Craig et al. 1998), the effect of lakes and wetlands on climate (Bonan 1995b), the effect of vegetation and soil (Kutzbach et al. 1996) and lakes and wetlands (Coe and Bonan 1997; Carrington et al. 2001) on the African monsoon in the middle Holocene, the effect of soil water on floods and droughts in the Mississippi River basin (Bonan and Stillwell-Soller 1998), and the effects of temperate deforestation on climate (Bonan 1997, 1999). The model has been extensively used for arctic studes (Lynch et al. 1998, 1999a,b, 2001; Tilley and Lynch 1998; Lynch and Wu 2000; Wu and Lynch 2000; Beringer et al. 2001).

In their documentation of the Common Land Model, Zeng et al. (2002) used a new soil color dataset ostensibly derived from the NCAR LSM dataset. One major difference, however, is that the 9th soil color class used in the Sahara Desert and Arabian Peninsula, which raised soil albedo, was eliminated based on analysis of satellite-derived surface albedo. The high soil albedo better matched clear sky top of the atmosphere albedo, but introduced a pronounced cold bias into the simulation (Bonan 1998). For consistency among models and to allow comparison with Zeng et al. (2002), we used the new soil color dataset for the LSM1, LSM2, and CLM2 simulations. Neither of these albedo datasets is very reliable, and it may be necessary to use new satellite-derived surface albedo datasets available in the future.

*b. LSM2*

In LSM1, the geography of PFTs and the structure of vegetation (the height, roughness length, displacement height, and leaf and stem area of each PFT) are based on biomes. The type of biome determines the composition of the vegetation (i.e., the PFTs and their abundance). The PFT determines vegetation structure. With the advent of 1-km satellite data products that allow separate specification of vegetation composition and structure, it is desirable to abandon the biome classification of land cover and separately specify vegetation composition and structure for each grid cell. This allows for a more accurate depiction of spatial heterogeneity in land cover. The PFT determines plant physiology while vegetation structure is direct input to each grid cell for each PFT. This also allows the model to interface with models of ecosystem processes and vegetation dynamics such as the LPJ dynamic global vegetation model (Sitch 2000; Cramer et al. 2001; McGuire et al. 2001). LPJ also uses PFTs to simulate the carbon cycle and vegetation dynamics, changing over time the structure and composition of patches of PFTs within a grid cell in response to disturbance (e.g., fire) and climate change. LSM2 is a restructuring of LSM1 to meet these objectives. Oleson and Bonan (2000) describe this methodology for a region of the boreal forest. Bonan et al. (2002) describe the global implementation.

In LSM2, a grid cell is divided into 5 primary land cover types: glacier, lake, wetland, urban and vegetation (Figure 1). An urban land cover is included so that future versions of the model can study urbanization, but currently the urban cover is zero. The vegetated portion of a grid cell is further divided into patches of up to 4 of 16 PFTs, each with its own leaf area index, stem area index, and canopy top and bottom heights. Bare ground is represented as no PFTs.

As described by Bonan et al. (2002), 0.5° maps of the abundance of 7 primary PFTs (needleleaf evergreen or deciduous tree, broadleaf evergreen or deciduous tree, shrub, grass, crop) were derived from the 1-km IGBP DISCover dataset (Loveland et al. 2000) and the 1-km University of Maryland tree cover dataset (DeFries et al. 1999, 2000a,b). Temperature and precipitation were used to distinguish arctic, boreal, temperate, and tropical plants, $C_3$ and $C_4$ grasses, and evergreen and deciduous shrubs (Table 2). Monthly leaf area index for each PFT in each 0.5° grid cell was obtained from 1-km Advanced Very High Resolution Radiometer (AVHRR) red and near infrared reflectances for April 1992 to March 1993 (Bonan et al. 2002). Stem area index, canopy top height, and canopy bottom height were based on the LSM1 values prescribed for each PFT (Bonan et al. 2002).

Physiological parameters for the 16 PFTs were obtained from the 12 LSM1 PFTs (Bonan 1996a) so that although the list of PFTs expanded no new physiologies were introduced. Two crop PFTs are available to account for the different physiology of crops, but only one is currently active because the 1-km land cover dataset does not distinguish crop varieties. Coupling with the LPJ dynamic global vegetation model (Sitch 2000; Cramer et al. 2001; McGuire et al. 2001) necessitated three changes in plant physiology from LSM1 (Table 2). First, roughness length and displacement height were changed to fractions of canopy top height because plant height changes during vegetation dynamics. These ratios were obtained from LSM1 values prescribed for each PFT and are similar to the values of 0.1 and 0.7 often cited for roughness length and displacement height, respectively (Bonan 2002). Second, coupling with LPJ revealed an inappropriate scaling of leaf stomatal conductance to the canopy. In LSM1, leaf processes are scaled to the canopy using sunlit and shaded leaves, which vary in absorbed photosynthetically active radiation. The formulation in LSM1 for shaded leaves was found to be unrealistic, allowing for net carbon gain at high leaf area index. In LSM2, the canopy scaling is replaced by an assumption similar to that of SiB2 whereby only sunlit leaves photosynthesize (Sellers et al. 1992, 1996). Third, values of $V_{max25}$ were correspondingly increased from LSM1 values to maintain realistic canopy photosynthesis (Table 2).

An additional feature of LSM2 is that soil texture (percent sand and clay) varies with depth according to the IGBP soil dataset (Global Soil Data Task 2000). This was motivated by a desire to include dust emissions as a component of the land model. Preliminary simulations with a dust emission parameterization found better entrainment of dust into the atmosphere in the Sahara Desert, a high dust source region, with the sandier top soil layers of the IGBP dataset rather than a uniform soil profile as in LSM1.

The surface dataset for LSM2 includes: the glacier, lake, wetland, and urban portions of the grid cell (vegetation occupies the remainder); the fractional cover in the vegetated portion of the grid cell of the 4 most abundant PFTs; monthly leaf and stem area index and canopy top and bottom heights for each PFT; soil color; and soil texture. These fields are aggregated to the CCM3 T42 grid from high-resolution surface datasets (Table 3). In contrast to LSM1, there is no irrigation of crops. This is because LSM1 recognizes irrigated crops as a biome, but LSM2 only recognizes a crop PFT.

Table 4 summarizes the differences between LSM1and LSM2. The primary difference is related to surface datasets – the representation of sub-grid land cover, vegetation structure, and soil texture. Biogeophysical parameterizations are the same except for canopy scaling and leaf physiology.

*c. CLM2*

In contrast to LSM2, which differs primarily from LSM1 in surface datasets, CLM2 uses the same surface datasets as LSM2 but differs from LSM2 in biogeophysical parameterizations. Many of the parameterizations are from the Common Land Model (Zeng et al. 2002), reconciled with the goal of including the carbon cycle and vegetation dynamics (Table 5). Major model differences from LSM1 include the LSM2 changes and: ten layers for soil temperature and soil water with explicit treatment of liquid water and ice; a multi-layer snow pack with up to 5 layers depending on snow depth; a runoff parameterization based on the TOPMODEL concept (Beven and Kirkby 1979); new formulation of ground and vegetation fluxes; and vertical root profiles from Zeng (2001).

Several differences in biogeophysical parameterizations between LSM1 and CLM2 explain many of the differences in simulated climate when coupled to CCM3. Both models have the same maximum canopy water storage (0.1 mm per unit leaf and stem area), but LSM1 restricts interception to 20% of precipitation while CLM2 intercepts more precipitation for leaf and stem area greater than about 0.5 $m^2$ $m^{-2}$ (Figure 2). At leaf and stem area index greater than about 4.5 $m^2$ $m^{-2}$, CLM2 allows more than 90% of precipitation to be intercepted (if storage capacity is not exceeded).

In both models, dry soils restrict transpiration by reducing photosynthesis and stomatal conductance. However, the relative influence of soil water varies greatly between the models (Figure 3). In LSM1, soil water does not restrict photosynthesis and stomatal conductance until soil is near wilting point. The CLM2 parameterization causes greater reduction for a similar water content.

Aerodynamic resistances to heat exchange from ground also differ between models. CLM2 uses a lower roughness length for bare ground than LSM1 and distinguishes between momentum and thermal roughness. The result is that CLM2 has a higher aerodynamic resistance to heat fluxes from bare ground (Figure 4). Within canopy aerodynamic processes also differ. CLM2 uses an aerodynamic resistance to heat exchange between the ground and canopy air that is greater than that of LSM1 (Figure 4).

CLM2 differs from LSM1 in its representation of snow. CLM2 uses a multi-layered snow. Heat and moisture transfer in the snowpack is based on temperature and water gradients between snow layers and on the physical properties of snow. LSM1 uses a single snow layer, blending the thermal properties of snow into the first soil layer and melting snow by solving the surface energy balance with a ground temperature of 0°C. In addition, thermal conductivity and heat capacity in CLM2 vary with the density of snow, but are constants in LSM1. Snow thermal conductivity in CLM2 is less than LSM1, especially at low bulk density. Offline simulations of LSM1 and CLM2 show CLM2 better simulates snow, especially during the melt season (Figure 5).

The single snow/soil parameterization used in LSM1 results in warmer soil temperatures than a multi-layer snowpack. With the lower thermal conductivity of snow compared to soil (typical values are 0.3 versus 2 W $m^{-1}$ $K^{-1}$), the blending approach of LSM1 captures the insulating effect of snow, reducing the overall thermal conductivity as snow depth increases. However, the high heat capacity of soil compared to snow (typical values are 2 versus 0.5 MJ $m^{-3}$ $K^{-1}$) ensures a high overall heat capacity that is not representative of snow. In addition, the ground heat flux in LSM1, computed in relation to the temperature gradient between the surface and first layer, decreases as the snowpack thickens. The warmer soil temperature is seen in simulations with LSM1 for 20 cm and 50 cm snowpacks using either a blended first layer or multi-layer heat transfer in snow (Table 6). With a 20 cm deep snowpack, the blended snow/soil layer is warmer (-7.0°C) than the surface layer of snow (-8.6°C). This difference is greater with a 50 cm snowpack.

Because CLM2 uses the same surface datasets as LSM2, it employs the same plant functional types. These plant types are defined in terms of leaf and stem optical properties (Table 7), plant morphology (Table 8), and photosynthetic parameters (Table 9).

*d. Observations*

Version 3.01 of the Willmott and Matsuura monthly terrestrial air temperature and precipitation climatology (Willmott and Matsuura 2000) was used to test the models. These datasets were created from the Global Historical Climatology Network (GHCN version 2) and Legates and Willmott's (1990a,b) station records of monthly air temperature covering the period 1950-1996. Station data were interpolated to a

0.5° grid using a distance-weighting method, with climatologically aided interpolation (Willmott and Robeson 1995) and adjustment of temperature for elevation (Willmott and Matsuura 1995).

Observed monthly snow cover was obtained from the NSIDC (1996) Northern Hemisphere weekly climatological snow cover dataset for the period 1971-1995. This climatology was derived from NOAA-NESDIS weekly snow charts derived from manual interpretation of AVHRR, GOES, and other visible-band satellite data, and then revised as in Robinson et al. (1993). The NOAA charts were digitized to an 89 × 89 grid and then interpolated to the final grid. Only grid cells at least 50% covered with snow were categorized as snow covered. Snow cover extent was defined as the area covered by at least 1-cm snow depth for the NOAA data (Foster et al. 1996). This corresponds to a snow water equivalent threshold of 2.5 mm in the models (assuming a snow density of 250 kg $m^{-3}$).

Observed runoff was obtained from the UNH-GRDC (University of New Hampshire-Global Runoff Data Center) 0.5º monthly climatological composite runoff fields of Fekete et al. (2000). These fields were generated by combining observed river discharge information with output from a climate-driven water balance model. This preserves the accuracy of the discharge fields while disaggregating the discharge spatially and temporally to enable comparisons with climate model output.

## 3. Results and Discussion

### *a. Surface air temperature*

Bonan (1998) describes the surface climatology of CCM3 with LSM1. The tropics between 15°S and 15°N and the middle latitudes of U.S. and Europe are generally well simulated throughout the year with temperatures for the most part within ±2°C of observations. The transition seasons (spring, autumn) are better simulated than winter or summer. A prominent winter bias is that surface air temperature is several degrees warmer than observations in a broad band of North America extending from central Canada northwest to Alaska and in Asia extending from the Caspian Sea to eastern Siberia. This warm bias is present in versions of CCM3 without LSM1. A prominent summer bias is that much of the Northern Hemisphere land is several degrees too cold. Prominent year-round temperature biases include the Andes region of South America, India, and the Tibetan Plateau, which are several degrees too cold throughout the year.

The Sahara Desert and Arabian Peninsula are also several degrees too cold throughout the year, possibly due in part to the assumed high surface albedos.

CLM2 generally warms surface air temperature in all seasons compared to LSM1 (Figure 6). Only two regions are cooler: North Africa year-round and northern regions of North America and Eurasia in winter. The warming introduced by CLM2 reduces or eliminates many cold biases found in LSM1 (Figure 7). In particular, the prominent Northern Hemisphere summer cold bias of LSM1 is virtually eliminated and in fact a warm bias has been introduced, especially in the U.S. Another prominent difference is that the year-round cold bias over the Tibetan Plateau has been reduced (Figure 7). This is due in part to the warmer climate of CLM2 (Figure 6), but also to cooler temperatures in the new observation dataset. The tropics between 15°S and 15°N are well simulated throughout the year with temperatures for the most part within ±2°C of the observations. The Amazon region of South America is a prominent exception. Here, temperatures are too warm throughout the year, with largest bias in the dry season. Several poor aspects of the LSM1 simulations noted in Bonan (1998) remain. In particular, the Alaskan and Asian winter warm biases still occur. The Sahara Desert and Arabian Peninsula is still several degrees too cold throughout the year, as is the Andes region of South America.

These temperature differences between models and biases with observations are evident in regional analyses of monthly surface air temperature. Temperatures in spring, summer, and autumn are well simulated in arctic and boreal latitudes, with the summer warming of CLM2 improving the simulation compared to LSM1 (Figure 8). The winter warm bias of both models is evident in Alaska and Northwest Canada, West Siberia, and East Siberia. The summer warming of CLM2 introduces a warm bias of a few degrees in middle latitudes compared to LSM1 (Figure 9). Other times of the year are generally well simulated compared to observations. Results in the tropics are mixed (Figure 10). CLM2 greatly improves simulated temperature in India. A small warm bias has been introduced in Central America. In Central Africa, the warmer CLM2 temperatures are consistent with observations during the first half of the year and warmer than observations in the second half. Air temperatures are consistently warmer than observations throughout the year in the Amazon. In the Sahara Desert, southern South America, South Africa, and Australia, CLM2 is consistently warmer than LSM1 throughout the year and generally reduces temperature biases seen in LSM1 (Figure 11).

Many of the temperature differences between CLM2 and LSM1 can be attributed to their different parameterization of biogeophysical processes rather than to different surface datasets. Comparison of CLM2 and LSM2, which use the same surface datasets but different biogeophysical parameterizations, show general patterns of winter cooling in northern regions of the Northern Hemisphere and year-round warming elsewhere (Figure 12). This accounts for much of the temperature difference seen in the comparison of CLM2 and LSM1 (Figure 6). Indeed, differences in surface air temperature between LSM2 and LSM1 (Figure 13) are generally smaller than those of CLM2 and LSM2 (Figure 12). However, the changes associated with LSM2 negatively impact North Africa year-round and northern Asia in winter.

LSM2 cools surface air temperature in North Africa compared to LSM1 and accounts for much of the cold temperature bias compared to observations (Figure 13). This is due to increased sand content that creates a drier soil and increases surface albedo. This is seen in offline simulations of LSM2 that used either the LSM1 soil texture or the new IGBP soil texture (Table 10). These simulations were forced with 3-hour atmospheric data for the period from 1979-1995 (Bonan et al. 2002). With an increase in sand content, the soil dries, absorbed solar radiation decreases, and the ground surface cools.

CLM2 biogeophysics cools much of the northern portions of the Northern Hemisphere in winter (Figure 12). In northern Eurasia, the change from the blended snow/soil representation of snow in LSM1 to the multi-layer snow of CLM2 likely contributes to the cooling. Indeed, this is seen in less heat loss from the soil in CLM2 compared to LSM1 (Table 11). Similar processes contribute to the cooling over North America (Table 11). Here, however, the cooling with CLM2 is augmented by higher surface albedo and more reflected solar radiation, which is one reason why the cooling is greater in North America than in Eurasia.

This winter cooling is desirable because it helps reduce a warm temperature bias in LSM1. Indeed, the warm bias in Canada is substantially reduced compared to the LSM1 warm bias reported by Bonan (1998). However, the cooling in Eurasia due to biogeophysical parameterizations is partially offset by the change in surface datasets from LSM1 to LSM2 (Figure 13). In this region, winter surface air temperature warms by a few degrees. Uncoupled simulations of LSM2 show that this warming is attributable to the IGBP soil texture dataset (Table 12). In the region of interest, clay content increases and sand content decreases. Soil moisture increases because of the higher matric potential and poorer drainage of clay soils.

With wetter soil, the volumetric latent heat of fusion of soil increases, more energy is released in freezing soil, and the soil is warmer.

*b. Precipitation*

Precipitation is for the most part reduced in CLM2 compared to LSM1 (Figure 14). Only tropical South America and Africa have a consistent year-round increase in precipitation, with the geographic location of this increase changing seasonally in relation to the seasonal migration of the intertropical convergence zone. Increased precipitation in these regions accentuates wet biases in the model (Figure 15). Regional analyses of monthly precipitation also show the general reduction in precipitation (Figure 8, Figure 9, Figure 10, Figure 11). CLM2 reduces annual precipitation by 4-11% in arctic and boreal latitudes, 11-17% in middle latitudes, 5-21% in the tropics, and 6-35% in arid regions compared to LSM1 (Table 13). Annual precipitation over land is reduced from 2.35 mm day$^{-1}$ in LSM1 to 2.14 mm day$^{-1}$ in CLM2, compared to 2.01 mm day$^{-1}$ for the observations.

*c. Snow cover*

In North America, CLM2 has higher snow cover than LSM1, especially during the melt season, and better reproduces observations (Figure 16). This is also seen in simulations for an individual watershed (Figure 5) and likely represents improvements in snow physics associated with the multi-layer snowpack of CLM2. In addition, CLM2 generally has lower snow albedos than LSM1. Less snow cover in LSM1 during the melt season is also seen in Eurasia, although here both models compare favorably with observations (Figure 16).

*d. Surface energy fluxes*

Analyses of regional energy fluxes generally show reduction in latent heat and increase in sensible heat for CLM2 compared to LSM1. This is likely the cause of the warm season warming in CLM2. Indeed, Bonan (1998) attributed the large Northern Hemisphere summer cold bias of LSM1 to high latent heat flux. The reduction in latent heat arises from a decrease in transpiration, large increase in evaporation of intercepted water, and large reduction in soil evaporation. This is illustrated for arctic and boreal latitudes by east Siberia (Figure 17). Here, the change in transpiration is negligible compared to the large summer increase in canopy evaporation and decrease in ground evaporation. With a warmer surface, sensible heat and net longwave emission to the atmosphere increase. Similar changes occur in middle and tropical latitudes with the exception that transpiration is substantially reduced. Central U.S. illustrates the reduction in transpiration, increase in canopy evaporation, decrease in ground evaporation, and increases in sensible heat and net longwave loss also found in eastern U.S. and central Europe (Figure 18). The Amazon Basin has similar changes, also seen in Central America, India, and the Congo (Figure 19). Arid regions (e.g., southern Africa) also show reduced transpiration and ground evaporation and increased canopy evaporation with CLM2 (Figure 20). In these regions, both latent and sensible heat decrease with CLM2 despite relatively little change in absorbed solar radiation. Instead, more energy is returned to the atmosphere as longwave radiation.

CLM2 and LSM1 differ substantially in the partitioning of latent heat into evaporation of intercepted water, transpiration, and ground evaporation. Leaves and stems intercept more water annually in CLM2 than LSM1 (Table 13), which gives rise to the higher canopy evaporation. This is a result of a greater fraction of precipitation intercepted for the same leaf and stem area (Figure 2). Greater interception means less precipitation reaches the ground to recharge the soil, contributing to drier soils in CLM2 than LSM1.

The reduction in transpiration with CLM2 is the result of several processes. The canopy integration scheme in CLM2 reduces transpiration compared to LSM1. Five-year simulations of CLM2 coupled to CCM3 and using either the LSM1 or CLM2 canopy integration parameterization show the reduction in transpiration due to the CLM2 parameterization contributes to the summer warm biases in central U.S. and Europe. This is also seen in regional analyses, which generally show reduced transpiration in LSM2 (which uses the same canopy integration as CLM2) compared to LSM1 (Figure 17, Figure 18, Figure 19, Figure 20). However, the difference between CLM2 and LSM2 is generally greater than the difference between LSM2 and LSM1. One process that likely contributes to this is greater interception in CLM2, which results in drier soil. This is accentuated by the different parameterization of soil water influence on stomata, which increases the degree to which soil water reduces transpiration in CLM2 compared to LSM1 (Figure 3). Five-year simulations with CCM3 and CLM2 that used either the LSM1 soil water factor or the CLM2 soil water factor show the CLM2 parameterization reduces latent heat, increases sensible heat, and warms surface air temperature in central U.S., central Europe, and the tropics in the June-August season.

The reduction in ground evaporation arises from drier soil and several parameterization differences with LSM1. LSM1 uses a surface resistance to reduce saturated soil evaporation for soil water limitation. This resistance is similar to the effect of soil water on stomata, increasing only with very dry soil. CLM2 reduces ground evaporation not through a surface resistance by rather through the moisture gradient with the atmosphere, decreasing saturated soil specific humidity as the soil dries. This is parameterized as a non-linear function of soil water. In addition, the aerodynamic resistance governing ground evaporation is higher in CLM2 than LSM1 (Figure 4). For bare ground, this is because the roughness length of soil is less in CLM2. For vegetation, this is because of different assumptions about turbulent processes within a canopy.

*e. Hydrologic cycle*

The increased interception in CLM2 results in a different hydrologic cycle compared to LSM1. In LSM1, 5-13% of precipitation is intercepted annually (Table 13). CLM2 intercepts 12-39%. In the Amazon and Congo, more than one-third of the annual precipitation does not reach the soil. Elsewhere, interception generally ranges from 20-25% of annual precipitation. Although both models have the same maximum water storage capacity (0.1 mm per unit leaf and stem area index), LSM1 limits interception to at most 20% of precipitation during any time-step while CLM2 allows a considerably greater fraction of precipitation to be intercepted (Figure 2). Although it is difficult to say which parameterization is correct, annual interception in CLM2 as a percent of annual precipitation is higher than the generally quoted values of 10-20% (Bonan 2002).

In addition to substantially reducing the water reaching the ground, CLM2 differs from LSM1 in the fate of this water. A higher fraction of water reaching the ground is lost as surface runoff or sub-surface drainage in CLM2 (Table 13). This reduces the soil water available for evapotranspiration. This is especially evident in central U.S., central Europe, and the Amazon, where runoff is 2-3 times higher in CLM2. Dry soil in these regions likely contributes to the warm temperature bias (Figure 7).

The seasonality of runoff is much improved compared to LSM1. In arctic and boreal regions, LSM1 has negligible seasonal variation in runoff (Figure 8). Observations have a pronounced peak during the snow melt season, which is better captured by CLM2. In addition, runoff in CLM2 is less than LSM1 and closer to observations during the cold season. This is likely due to the accounting of frozen and unfrozen water in CLM2. Drainage does not occur if ice exists in the soil column. In contrast, LSM1 allows drainage regardless of the thermal state of the soil. Similar improvements are seen in middle latitudes regions such as central U.S., eastern U.S., and central Europe (Figure 9). In these regions, however, cold season hydrology is less important. Instead, the exponential decay of saturated hydraulic conductivity and the new 10-layer soil water parameterization in CLM2 result in phase shifts in runoff compared to LSM1. Differences between models are less apparent in the tropics, where both models reproduce the seasonality of runoff (Figure 10). CLM2 generally has more runoff than LSM1 during the rainy season. This difference is also seen in arid regions (Figure 11).

**4. Conclusions**

The surface physics and hydrology of the Community Land Model for use with the Community Climate System Model has been greatly updated from the NCAR LSM. This reflects new ideas formulated in the Common Land Model (Zeng et al. 2002) and to facilitate coupling to terrestrial ecosystems models (Bonan et al. 2002). Major model differences include: abandonment of the biome classification of surface types and inclusion of a sub-grid mosaic of land cover types and plant functional types; satellite-derived land cover, plant type, and leaf area index datasets; ten soil layers with explicit treatment of liquid water and ice; a multi-layer snow pack; runoff based on the TOPMODEL concept; new formulation of ground and vegetation fluxes; and vertical root profiles from a global synthesis of ecological studies.

CLM2 generally warms surface air temperature in all seasons compared to LSM1, reducing or eliminating many cold biases. In particular, the prominent Northern Hemisphere summer cold bias of LSM1 has been eliminated due to reduced latent heat and increased sensible heat. Reduction in latent heat arises from reduced transpiration and ground evaporation but increased evaporation of intercepted water. This reflects increased soil water limitation to transpiration, higher aerodynamic resistances to heat exchange from the ground, and greater interception in CLM2 compared to LSM1. New snow parameterizations including a multi-layer snowpack result in improved simulation of snow depth and snow cover. The annual cycle of runoff is greatly improved in CLM2, especially in arctic and boreal regions where the inclusion of cold season hydrology improves the annual cycle of runoff.

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**Figure Legends**

Figure 1. Representation of land cover heterogeneity in LSM2 and CLM2. The grid cell is divided into 5 primary land cover types. The vegetated portion is further divided into up to 4 types of plants.

Figure 2. Interception of water in LSM1 and CLM2. Both models intercept a fraction of the incoming precipitation up to a maximum storage capacity. This storage capacity is the same in both models.

Figure 3. Soil water factor in LSM1 and CLM2 in relation to soil water content for a loamy soil. This dimensionless factor reduces photosynthesis, stomatal conductance, and transpiration as the soil dries.

Figure 4. Bare ground and within canopy aerodynamic resistances to heat and water in relation to wind speed for LSM1 and CLM2 . Resistances use an atmospheric height of 45 m and neutral atmospheric conditions. Within canopy resistances are shown for tree and grass.

Figure 5. Daily snow water equivalent for LSM1 and CLM2 forced with 18-years (1966-1983) of observed atmospheric data for a grassland catchment at the Valdai water balance research station in Russia (Vinnikov et al. 1996; Schlosser et al. 1997). The site is assumed to be 90% $C_3$ grassland and 10% bare soil. The models were initialized repeating the first year of forcing until equilibrium (Yang et al. 1995). The observations are the daily average from up to 44 sites throughout the catchment (Schlosser et al. 1998).

Figure 6. Surface (2 m) air temperature difference between CLM2 and LSM1 (CLM2-LSM1) for the four seasons (December-February, March-May, June-August, and September-November). Stippling shows regions where the difference is statistically significant based on a t-test ($p < 0.05$).

Figure 7. CLM2 surface (2 m) air temperature bias compared to the Willmott and Matsuura (2000) observations for the four seasons.

Figure 8. Regionally averaged monthly surface (2 m) air temperature, precipitation, and runoff for LSM1, CLM2, and observations in arctic and boreal latitudes. Data are spatially averaged for land points.

Figure 9. As in Figure 8, but for middle latitudes.

Figure 10. As in Figure 8, but for tropical latitudes.

Figure 11. As in Figure 8, but for arid regions.

Figure 12. Surface (2 m) air temperature difference between CLM2 and LSM2 (CLM2-LSM2) for the four seasons (December-February, March-May, June-August, and September-November). Stippling shows regions where the difference is statistically significant based on a t-test ($p < 0.05$).

Figure 13. As in Figure 12, but for the LSM2-LSM1 difference.

Figure 14. Precipitation difference between CLM2 and LSM1 (CLM2-LSM1) for the four seasons (December-February, March-May, June-August, and September-November). Stippling shows regions where the difference is statistically significant based on a t-test ($p < 0.05$).

Figure 15. CLM2 precipitation bias compared to the Willmott and Matsuura (2000) observations for the four seasons.

Figure 16. Monthly snow cover for North America (including Greenland) and Eurasia for LSM1 and CLM2. Observations are from the NSIDC (1996) Northern Hemisphere climatology for the period 1971-1995.

Figure 17. Regionally averaged monthly surface energy fluxes for LSM1, LSM2, and CLM2 in east Siberia. Latent heat is partitioned into transpiration, evaporation of water intercepted by the canopy, and ground evaporation. Net longwave is the net loss to the atmosphere.

Figure 18. As in Figure 17, but for central U.S.

Figure 19. As in Figure 17, but for the Amazon.

Figure 20. As in Figure 17, but for southern Africa.

Table 1. Surface types, associated PFTs, and fractional cover used in LSM1. NET, needleleaf evergreen tree; NDT, needleleaf deciduous tree; BET, broadleaf evergreen tree; BDT, broadleaf deciduous tree; TST, tropical seasonal tree; ES, evergreen shrub; DS, deciduous shrub; ADS, arctic deciduous shrub; CG, $C_3$ grass; WG, $C_4$ grass; AG, arctic grass; C, crop; B, bare.

| | Patch 1 | | Patch 2 | | Patch 3 | |
|---|---|---|---|---|---|---|
| Surface Type | PFT | Cover | PFT | Cover | PFT | Cover |
| Glacier | B | 1.00 | - | - | - | - |
| Desert | B | 1.00 | - | - | - | - |
| Needleleaf evergreen forest, cool | NET | 0.75 | B | 0.25 | - | - |
| Needleleaf deciduous forest, cool | NDT | 0.50 | B | 0.50 | - | - |
| Broadleaf deciduous forest, cool | BDT | 0.75 | B | 0.25 | | |
| Mixed forest, cool | NET | 0.37 | BDT | 0.37 | B | 0.26 |
| Needleleaf evergreen forest, warm | NET | 0.75 | B | 0.25 | - | - |
| Broadleaf deciduous forest, warm | BDT | 0.75 | B | 0.25 | | |
| Mixed forest, warm | NET | 0.37 | BDT | 0.37 | B | 0.26 |
| Broadleaf evergreen forest, tropical | BET | 0.95 | B | 0.05 | - | - |
| Broadleaf deciduous forest, tropical | TST | 0.75 | B | 0.25 | - | - |
| Savanna | WG | 0.70 | TST | 0.30 | - | - |
| Forest tundra, evergreen | NET | 0.25 | AG | 0.25 | B | 0.50 |
| Forest tundra, deciduous | NDT | 0.25 | AG | 0.25 | B | 0.50 |
| Forest crop, cool | C | 0.40 | BDT | 0.30 | NET | 0.30 |
| Forest crop, warm | C | 0.40 | BDT | 0.30 | NET | 0.30 |
| Grassland, cool | CG | 0.60 | WG | 0.20 | B | 0.20 |
| Grassland, warm | WG | 0.60 | CG | 0.20 | B | 0.20 |
| Tundra | ADS | 0.30 | AG | 0.30 | B | 0.40 |
| Shrubland, evergreen | ES | 0.80 | B | 0.20 | - | - |
| Shrubland, deciduous | DS | 0.80 | B | 0.20 | - | - |
| Semi-desert | DS | 0.10 | B | 0.90 | - | - |

| | | | | | | |
|---|---|---|---|---|---|---|
| Irrigated crop, cool | C | 0.85 | B | 0.15 | - | - |
| Crop, cool | C | 0.85 | B | 0.15 | - | - |
| Irrigated crop, warm | C | 0.85 | B | 0.15 | - | - |
| Crop, warm | C | 0.85 | B | 0.15 | - | - |
| Wetland, forest | BET | 0.80 | B | 0.20 | - | - |
| Wetland, non-forest | B | 1.00 | - | - | - | - |

Table 2. Plant functional types for LSM2 and CLM2 and their derivation from 1-km land cover data. Two types of crops are allowed to account for the different physiology of crops, but currently only one type is specified in the surface datasets. Roughness length and displacement height are as a fraction of canopy top height.

| 1-km Land Cover and Tree Cover Data | Plant Functional Type | $V_{max25}$ (μmol $CO_2$ $m^{-2}$ $s^{-1}$) | $z_0$ | d |
|---|---|---|---|---|
| Needleleaf evergreen tree | Needleleaf evergreen tree, temperate | 51 | 0.055 | 0.67 |
| | Needleleaf evergreen tree, boreal | 43 | 0.055 | 0.67 |
| Needleleaf deciduous tree | Needleleaf deciduous tree, boreal | 43 | 0.055 | 0.67 |
| Broadleaf evergreen tree | Broadleaf evergreen tree, tropical | 75 | 0.075 | 0.67 |
| | Broadleaf evergreen tree, temperate | 69 | 0.075 | 0.67 |
| Broadleaf deciduous tree | Broadleaf deciduous tree, tropical | 40 | 0.055 | 0.67 |
| | Broadleaf deciduous tree, temperate | 51 | 0.055 | 0.67 |
| | Broadleaf deciduous tree, boreal | 51 | 0.055 | 0.67 |
| Shrub | Broadleaf evergreen shrub, temperate | 17 | 0.120 | 0.68 |
| | Broadleaf deciduous shrub, temperate | 17 | 0.120 | 0.68 |
| | Broadleaf deciduous shrub, boreal | 33 | 0.120 | 0.68 |
| Grass | $C_3$ grass, arctic | 43 | 0.120 | 0.68 |
| | $C_3$ grass | 43 | 0.120 | 0.68 |
| | $C_4$ grass | 24 | 0.120 | 0.68 |
| Crop | Crop1 | 50 | 0.120 | 0.68 |
| | Crop2 | - | - | - |

Table 3. Surface data required for LSM2 and CLM2, their base spatial resolution, and method of aggregation to the model's grid.

| Surface Field | Resolution | Source | Aggregation Method |
|---|---|---|---|
| Percent glacier | 0.5° | Bonan et al. (2002) | Area average |
| Percent lake | 1° | LSM1 | Area average |
| Percent wetland | 1° | LSM1 | Area average |
| Percent sand, percent clay | 5-minute | IGBP dataset of 4931 soil mapping units and their sand and clay content for each soil layer | Soil mapping unit with greatest areal extent in grid cell |
| Soil color | 2.8° (T42) | Zeng et al. (2002) dataset of 8 color classes without brightened soil over the Sahara Desert and Arabian Peninsula | Soil color class with greatest areal extent in grid cell |
| PFTs (percent of vegetated land) | 0.5° | Bonan et al. (2002) | Area average, choosing 4 most abundant PFTs |
| Monthly leaf and stem area index | 0.5° | Bonan et al. (2002) | Area average |
| Canopy height (top, bottom) | 0.5° | Bonan et al. (2002) | Area average |

Table 4. Differences between LSM1 and LSM2.

| Process/Parameterization | LSM1 | LSM2 |
|---|---|---|
| Land cover | Biome approach. Biomes determine PFTs. Glaciers are a biome, but lakes and wetlands are sub-grid patches. | Sub-grid representation of glacier, lake, wetland, urban, and vegetation. Explicit representation of PFTs. |
| Vegetation structure | Leaf area index, stem area index, roughness length, displacement height, canopy top and bottom heights, and root distribution based on PFTs. | Leaf area index, stem area index, and canopy heights in surface datasets. Roughness length and displacement height depend on canopy height. Root distribution depends on PFTs. |
| Soil texture | Sand and clay constant with depth. | Sand and clay vary with depth. |
| Canopy scaling | Sunlit and shaded leaves. Sunlit leaves receive direct beam and a portion of diffuse radiation. Shaded leaves receive only diffuse radiation. | Sunlit and shaded leaves. Sunlit leaves receive all radiation. Shaded leaves dark. |
| Leaf physiology | - | Altered $V_{max25}$. |

Table 5. Conceptual similarities and differences between LSM1 and CLM2.

| Process/Parameterization | LSM1 | CLM2 |
|---|---|---|
| Land cover | 28 biomes provide 2 sub-grid PFT patches. Sub-grid lake and wetland patches. Leaf area index, stem area index, roughness length, displacement height, canopy top and bottom heights, and root distribution based on PFTs. | LSM2: Sub-grid glacier, lake, wetland, urban, and vegetation. Explicit representation of PFTs. Explicit leaf area index, stem area index, and canopy heights in surface datasets. Roughness length and displacement height depend on canopy height. Root distribution depends on PFTs using Common Land Model formulation. |
| Vegetation | 1 canopy layer for fluxes, but 2 leaves (sunlit, shaded) for canopy integration of stomatal conductance. | Common Land Model: 1 canopy layer for fluxes, but 2 leaves (sunlit, shaded) for canopy integration of stomatal conductance. |
| Snow | 1 layer mass balance. Blended with top soil layer for heat transfer. Snow covers ground in relation to snow depth and buries vegetation vertically in relation to canopy bottom height. | Common Land Model formulation of up to 5 layers depending on snow depth. LSM1 vertical burying of vegetation and Common Land Model fractional snow-covered ground. |
| Soil | 6 layers to depth of 6.3 m. First layer 10 cm thick. Thermal and hydraulic properties depend on sand and clay. Sand and clay constant with depth. | Common Land Model: 10 layers to depth of 3.43 m. First layer 1.75 cm thick. Thermal and hydraulic properties depend on sand and clay. Sand and clay vary with depth as in LSM2. |
| Lake | 6 layers to depth of 50 m. | Common Land Model: As in LSM1, but with 10 layers. |

| | | |
|---|---|---|
| | Albedo | |
| Snow | Depends on zenith angle, soot content, and grain radius. | Common Land Model: BATS snow albedo varying with snow age and zenith angle. |
| Soil | BATS color classes and soil water dependence. Blended with snow. | Common Land Model: BATS color classes and soil water dependence. Blended with snow. |
| Vegetation | Two-stream radiative transfer. PFT-dependent leaf optical properties modified by intercepted snow. Exposed leaf and stem area above snow decreases as snow accumulates above lower canopy height. | LSM1. |
| | Hydrology | |
| Interception | Maximum storage is 0.1 mm times leaf and stem area. Interception either to storage capacity or 20% of precipitation. Separate regions receiving large-scale and convective precipitation. | Common Land Model: Interception either to storage capacity (same as LSM1) or some fraction of precipitation. Fraction of precipitation intercepted increases with leaf and stem area. |
| Surface runoff | Runoff from saturated and unsaturated zones. Exponential spatial distribution of soil water determines saturated area. Separate regions receiving large-scale and convective precipitation. Precipitation has exponential distribution in each region. | Common Land Model: Saturated and unsaturated zones using TOPMODEL-like approach. Exponential decrease in saturated hydraulic conductivity determines water table depth and saturated area. |

| | | |
|---|---|---|
| Base flow (drainage) | Depends on hydraulic conductivity of bottom soil layer regardless of temperature. | Common Land Model: Ice-free soil only. Depends on hydraulic conductivity (unsaturated zone) or base rate and water table depth (saturated zone). |
| Snow | 1 layer mass balance. | Common Land Model: Up to 5 layers depending on snow depth. Compaction allowed. Ice and liquid water. Heat and water flow between layers. |
| Soil water | Darcy's law for vertical fluxes. Water removed by evaporation from top layer, transpiration from each layer in proportion to root abundance, and drainage from bottom layer. No distinction between ice and water. | Common Land Model: Darcy's law for vertical fluxes. Water removed by evaporation from top layer, transpiration from each layer in proportion to root abundance, and drainage from deep layers. Separate ice and water. |
| | Ground fluxes | |
| Turbulent fluxes | Monin-Obukhov similarity theory. | Common Land Model: Monin-Obukhov similarity theory but different flux-gradient relations. Lower aerodynamic roughness for momentum than LSM1. Separate thermal roughness for heat and water vapor. |
| Soil moisture limitation to evaporation | Surface resistance that depends on soil water. | Common Land Model: Soil relative humidity modifies ground specific humidity. |

| | | |
|---|---|---|
| Ground temperature | Newton-Raphson iteration of surface energy budget. Soil heat flux used to update soil temperatures. | Common Land Model: Coupled to snow/soil temperature algorithm. Top layer heat capacity modified by adjusting thickness to obtain ground skin temperature. |
| Soil temperature | Crank-Nicholson formulation of one-dimensional heat flow and energy conservation. Apparent heat capacity accounts for phase change. Snow blended into first soil layer. Explicit coupling with ground fluxes. | Common Land Model: Crank-Nicholson formulation of one-dimensional heat flow and energy conservation through snow/soil column. Frozen and liquid water. Phase change by setting temperature to freezing, accounting for energy change. Implicit coupling with ground fluxes. |
| | Vegetation fluxes | |
| Leaf temperature and turbulent fluxes | Sensible heat from foliage and ground. Latent heat from ground, intercepted water, and transpiration. Newton-Raphson iteration of energy budget. | Common Land Model: BATS formulation of leaf temperature and fluxes. Sensible heat from foliage and ground. Latent heat from ground, intercepted water, and transpiration. Newton-Raphson iteration of energy budget. |
| Leaf boundary layer resistance | Depends on leaf dimension and wind speed. Wind speed integrated through canopy assuming exponential decline in wind. | Common Land Model: BATS formulation. Depends on leaf dimension and friction velocity. |

| | | |
|---|---|---|
| Leaf stomatal resistance | Stomatal resistance depends on photosynthesis in relation to light, temperature, $CO_2$, vapor pressure, foliage nitrogen, and soil water. Soil water affects photosynthesis by altering $V_{max25}$. Soil water factor is a linear function of soil water scaled to 1 at some optimal value and 0 when dry. | LSM1, but Common Land Model formulation of soil water limitation. Soil water factor is non-linear function of soil water based on matric potential and root resistance. |
| Canopy resistance | Sunlit and shaded leaves used to integrate leaf resistance to canopy (Table 4). | LSM2 (Table 4). |
| Soil-to-air exchange | Within canopy aerodynamic resistance based on friction velocity and exponential profile of eddy diffusivity for heat. | Common Land Model: BATS formulation based on friction velocity times a constant transfer coefficient. |
| Above-canopy exchange | Monin-Obukhov similarity theory. | Common Land Model: Monin-Obukhov similarity theory, but different flux-gradient relations. |

Table 6. Comparison of the blended snow/soil formulation of LSM1 with multi-layered heat transfer in snow for 20 cm and 50 cm snowpacks. Both methods use the soil temperature numerical algorithms of LSM1. The blended approach uses 6 soil layers, blending the thermal properties of snow with the first soil layer. The layered method adds two 10 cm snow layers (20 cm) or five 10 cm snow layers (50 cm) on top of the 6 soil layers, using the heat capacity and thermal conductivity of snow. Results show the temperature of the first model layer after a 100 day simulation with surface temperature of -10°C and initial soil temperature profile of 2°C.

| | Temperature (°C) | |
|---|---|---|
| Snow Depth | Blended | Layered |
| 20 cm | -7.0 | -8.6 |
| 50 cm | -5.5 | -9.1 |

Table 7. Optical properties for each plant functional type. See Table 2 for a definition of the plant types. Reflectance and transmittance are for visible (VIS) and near-infrared (NIR) wavebands.

| Plant Functional Type | Leaf Angle | Leaf Reflectance | | Stem Reflectance | | Leaf Transmittance | | Stem Transmittance | |
|---|---|---|---|---|---|---|---|---|---|
| | | VIS | NIR | VIS | NIR | VIS | NIR | VIS | NIR |
| NET temperate | 0.01 | 0.07 | 0.35 | 0.16 | 0.39 | 0.05 | 0.10 | 0.001 | 0.001 |
| NET boreal | 0.01 | 0.07 | 0.35 | 0.16 | 0.39 | 0.05 | 0.10 | 0.001 | 0.001 |
| NDT boreal | 0.01 | 0.07 | 0.35 | 0.16 | 0.39 | 0.05 | 0.10 | 0.001 | 0.001 |
| BET tropical | 0.10 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BET temperate | 0.10 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BDT tropical | 0.01 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BDT temperate | 0.25 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BDT boreal | 0.25 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BES temperate | 0.01 | 0.07 | 0.35 | 0.16 | 0.39 | 0.05 | 0.10 | 0.001 | 0.001 |
| BDS temperate | 0.25 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| BDS boreal | 0.25 | 0.10 | 0.45 | 0.16 | 0.39 | 0.05 | 0.25 | 0.001 | 0.001 |
| $C_3$ grass arctic | -0.30 | 0.11 | 0.58 | 0.36 | 0.58 | 0.07 | 0.25 | 0.220 | 0.380 |
| $C_3$ grass | -0.30 | 0.11 | 0.58 | 0.36 | 0.58 | 0.07 | 0.25 | 0.220 | 0.380 |
| $C_4$ grass | -0.30 | 0.11 | 0.58 | 0.36 | 0.58 | 0.07 | 0.25 | 0.220 | 0.380 |
| Crop1 | -0.30 | 0.11 | 0.58 | 0.36 | 0.58 | 0.07 | 0.25 | 0.220 | 0.380 |
| Crop2 | - | - | - | - | - | - | - | - | - |

Table 8. Morphology for each plant functional type. See Table 2 for a definition of the plant types. Roughness length and displacement height are as a fraction of canopy top height. Root distribution at depth z (m) is $f(z) = 1 - 0.5[\exp(-a\,z) + \exp(-b\,z)]$.

| Plant Functional Type | Leaf Dimension (m) | Roughness Length | Displacement Height | Root Distribution a | Root Distribution b |
|---|---|---|---|---|---|
| NET temperate | 0.04 | 0.055 | 0.67 | 7.0 | 2.0 |
| NET boreal | 0.04 | 0.055 | 0.67 | 7.0 | 2.0 |
| NDT boreal | 0.04 | 0.055 | 0.67 | 7.0 | 2.0 |
| BET tropical | 0.04 | 0.075 | 0.67 | 7.0 | 1.0 |
| BET temperate | 0.04 | 0.075 | 0.67 | 7.0 | 1.0 |
| BDT tropical | 0.04 | 0.055 | 0.67 | 6.0 | 2.0 |
| BDT temperate | 0.04 | 0.055 | 0.67 | 6.0 | 2.0 |
| BDT boreal | 0.04 | 0.055 | 0.67 | 6.0 | 2.0 |
| BES temperate | 0.04 | 0.120 | 0.68 | 7.0 | 1.5 |
| BDS temperate | 0.04 | 0.120 | 0.68 | 7.0 | 1.5 |
| BDS boreal | 0.04 | 0.120 | 0.68 | 7.0 | 1.5 |
| $C_3$ grass arctic | 0.04 | 0.120 | 0.68 | 11.0 | 2.0 |
| $C_3$ grass | 0.04 | 0.120 | 0.68 | 11.0 | 2.0 |
| $C_4$ grass | 0.04 | 0.120 | 0.68 | 11.0 | 2.0 |
| Crop1 | 0.04 | 0.120 | 0.68 | 6.0 | 3.0 |
| Crop2 | - | - | - | - | - |

Table 9. Photosynthetic parameters for each plant functional type. See Table 2 for a definition of the plant types. Path, photosynthetic pathway; $V_{max25}$, maximum carboxylation at 25°C ($\mu$mol $CO_2$ $m^{-2}$ $s^{-1}$); $\alpha$, quantum efficiency ($\mu$mol $CO_2$ $\mu$mol $photon^{-1}$); m, slope of conductance-photosynthesis relationship.

| Plant Functional Type | Path | $V_{max25}$ | $\alpha$ | m |
|---|---|---|---|---|
| NET temperate | $C_3$ | 51 | 0.06 | 6 |
| NET boreal | $C_3$ | 43 | 0.06 | 6 |
| NDT boreal | $C_3$ | 43 | 0.06 | 6 |
| BET tropical | $C_3$ | 75 | 0.06 | 9 |
| BET temperate | $C_3$ | 69 | 0.06 | 9 |
| BDT tropical | $C_3$ | 40 | 0.06 | 9 |
| BDT temperate | $C_3$ | 51 | 0.06 | 9 |
| BDT boreal | $C_3$ | 51 | 0.06 | 9 |
| BES temperate | $C_3$ | 17 | 0.06 | 9 |
| BDS temperate | $C_3$ | 17 | 0.06 | 9 |
| BDS boreal | $C_3$ | 33 | 0.06 | 9 |
| $C_3$ grass arctic | $C_3$ | 43 | 0.06 | 9 |
| $C_3$ grass | $C_3$ | 43 | 0.06 | 9 |
| $C_4$ grass | $C_4$ | 24 | 0.04 | 5 |
| Crop1 | $C_3$ | 50 | 0.06 | 9 |
| Crop2 | - | - | - | - |

Table 10. Effect of IGBP soil texture on the simulated climate of the Sahara Desert. Simulations are with LSM2 forced with atmospheric data for the period 1979-1995 (see Bonan et al. 2002 for details) and using the IGBP and LSM1 soil texture datasets. Data are spatial averages of the difference (IGBP minus LSM1) for land grid cells in the region 20°-30°N and 20°W-32°E over the period 1984-1995 for the December-February (DJF) and June-August (JJA) seasons.

| Surface Variable | DJF | JJA | Clay (%) | Sand (%) |
|---|---|---|---|---|
| Soil Water ($mm^3\ mm^{-3}$) | | | | |
| 5 cm | -0.06 | -0.06 | -7 | 28 |
| 20 cm | -0.04 | -0.04 | -6 | 26 |
| 50 cm | -0.03 | -0.03 | -5 | 24 |
| 110 cm | -0.03 | -0.03 | -5 | 21 |
| 230 cm | -0.04 | -0.04 | -8 | 24 |
| 470 cm | -0.04 | -0.04 | -8 | 24 |
| Ground Temperature (°C) | -0.3 | -0.4 | - | - |
| Absorbed Solar Radiation ($W\ m^{-2}$) | -4.4 | -7.9 | - | - |

Table 11. Surface energy balance (W $m^{-2}$) for northern Eurasia (50°-70°N, 5°-130°E) and North America (40°-60°N, 130°-60°W) during the December-February season. Data are the LSM1 and CLM2 simulations.

| | Northern Eurasia | | North America | |
|---|---|---|---|---|
| | LSM1 | CLM2 | LSM1 | CLM2 |
| Incoming solar radiation, S↓ | 21.3 | 19.1 | 51.2 | 49.4 |
| Reflected solar radiation, S↑ | 11.3 | 8.7 | 15.1 | 16.5 |
| Absorbed solar radiation, S↓-S↑ | 10.0 | 10.4 | 36.1 | 32.8 |
| Net longwave radiation (L↑-L↓) | 23.6 | 19.4 | 41.1 | 35.2 |
| Net radiation (S↓-S↑+L↓-L↑) | -13.6 | -9.0 | -5.0 | -2.4 |
| Latent heat | 4.2 | 5.1 | 8.2 | 8.7 |
| Sensible heat | -6.9 | -6.7 | -1.4 | -3.6 |
| Soil heat | -11.7 | -8.2 | -13.9 | -9.5 |
| Snow melt | 0.7 | 0.8 | 2.1 | 2.1 |

Table 12. As in Table 10, but for northern Asia (45°-70°N, 75°-140°E).

| Soil Depth (cm) | Clay (%) | Sand (%) | DJF Soil Water ($mm^3\ mm^{-3}$) | DJF Temperature (°C) |
|---|---|---|---|---|
| 5 | 8 | -20 | 0.05 | 0.8 |
| 20 | 9 | -20 | 0.06 | 1.2 |
| 50 | 10 | -20 | 0.06 | 1.1 |
| 110 | 9 | -20 | 0.06 | 0.9 |
| 230 | 8 | -19 | 0.05 | 0.6 |
| 470 | 8 | -19 | 0.05 | 0.4 |

Table 13. Annual hydrologic cycle. P, precipitation. I, interception. T, transpiration. E, ground evaporation. R, total runoff. Interception is as a percent of precipitation. Transpiration, evaporation, and runoff are as a percent of water reaching the ground (P-I). Regions are defined in Figure 8 to Figure 11.

| | P (mm) | | I (% P) | | T (% P-I) | | E (% P-I) | | R (% P-I) | |
|---|---|---|---|---|---|---|---|---|---|---|
| Region | LSM1 | CLM2 | LSM1 | CLM2 | LSM1 | CLM2 | LSM1 | CLM2 | LSM1 | CLM2 |
| Arctic and boreal | | | | | | | | | | |
| Alaska and NW Canada | 798 | 755 | 7 | 15 | 6 | 3 | 26 | 20 | 66 | 75 |
| Northern Europe | 759 | 700 | 9 | 22 | 9 | 9 | 28 | 22 | 62 | 68 |
| West Siberia | 673 | 600 | 6 | 19 | 9 | 10 | 41 | 42 | 48 | 48 |
| East Siberia | 566 | 541 | 8 | 24 | 8 | 11 | 39 | 26 | 51 | 62 |
| Middle latitudes | | | | | | | | | | |
| Western U.S. | 682 | 597 | 8 | 16 | 14 | 10 | 36 | 21 | 50 | 68 |
| Central U.S. | 688 | 570 | 9 | 28 | 25 | 20 | 67 | 55 | 8 | 24 |
| Eastern U.S. | 952 | 848 | 8 | 27 | 24 | 22 | 50 | 43 | 25 | 34 |
| Central Europe | 815 | 687 | 6 | 22 | 18 | 17 | 51 | 34 | 28 | 49 |
| Tropics | | | | | | | | | | |
| Central America | 1192 | 1080 | 12 | 29 | 37 | 16 | 35 | 22 | 27 | 61 |
| India | 1395 | 1105 | 6 | 12 | 13 | 8 | 50 | 40 | 36 | 51 |
| Amazon | 2292 | 2058 | 13 | 39 | 44 | 15 | 15 | 15 | 39 | 68 |
| Congo | 2244 | 2128 | 12 | 35 | 34 | 14 | 17 | 17 | 47 | 68 |
| Indochina | 1100 | 992 | 8 | 22 | 26 | 14 | 67 | 48 | 6 | 37 |
| Arid | | | | | | | | | | |
| Sahara Desert | 465 | 303 | 6 | 12 | 18 | 4 | 68 | 63 | 14 | 33 |
| S. South Amer. | 845 | 723 | 5 | 25 | 11 | 16 | 80 | 62 | 9 | 21 |
| South Africa | 869 | 817 | 8 | 24 | 22 | 13 | 61 | 38 | 16 | 48 |
| Australia | 569 | 374 | 9 | 22 | 22 | 16 | 69 | 67 | 9 | 17 |