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ABSTRACT—Simulation models of soil-related biological processes usually require soil temperature data. Frequently these soil temperatures are simulated, and the soil temperature algorithms cannot be more complicated than the original process model. This situation has led to the use of semi-empirical-type relationships in these process models. The objective of this study was to evaluate a hybrid soil temperature model, which combines empirical and mechanistic approaches, in an agroecosystem and a tallgrass prairie in the Great Plains. The original hybrid soil temperature model was developed and verified for a temperate forest system. This model simulated soil temperatures on a daily basis from meteorological inputs (maximum and minimum air temperatures) and soil and plant properties. This model was modified using different extinction coefficients for the plant canopy and ground litter. The agroecosystem consisted of a no-till rotation system of corn (*Zea mays* L.) and soybeans (*Glycine max* [L.] Merr.). Soil temperatures were measured at different depths in multiple years (three years and two-and-a-half years in the agroecosystem and tallgrass prairie, respectively). In the agroecosystem, the root mean square error of the modified model simulation varied from 1.41º to 2.05ºC for the four depths (0.1, 0.2, 0.3, and 0.5 m). The mean absolute error varied from 1.06º to 1.53ºC. The root mean square error and mean absolute error of the modified model were about 0.1º–0.3ºC less than the original model at the 0.2–0.5 m depths. For the tallgrass prairie, the mean absolute errors of the simulated soil temperatures were slightly greater than the agroecosystem, varying from 1.48º to 1.7ºC for all years and from 1.09º to 1.37ºC during the active growing seasons for all years.

Key Words: corn-soybean rotation, litter, no-tillage, simulation, soil temperature, tallgrass prairie

INTRODUCTION

In crop and land surface models, the accurate simulation of soil temperature is important because it is a critical factor that regulates chemical, physical, and biological processes in the soil. Variation in soil temperature influences seedling growth and survival (King et al. 1997), root growth and turnover (Forbes et al. 1997; Hendrick and Pregitzer 1997), mycorrhiza colonization and development (Gavito et al. 2003), soil organic matter decomposition (Parton et al. 1987; Kirschbaum 1995), nutrient dynamics (Parton et al. 1988; Mellander et al. 2004), and soil respiration (Boone et al. 1998; Raich and Schlesinger 1992).

Different approaches to simulate soil temperature, with different degrees of complexity, are currently implemented in different crop and land surface models. These approaches can be grouped into three categories. The first approach, empirical models, are relatively easy to implement and are based on statistical relationships between soil temperature and surface observations and soil characteristics (e.g., Kemp et al. 1992). These models are site specific and usually need location-specific calibrations. The second approach, mechanistic models, are characterized by quantification of the energy balance processes that determine soil temperature such as the radiative, sensible, latent, and soil heat fluxes (e.g., Campbell 1985; Smirnova et al. 1997). These models have general applicability but can require input data, which may not be readily available. These numerical models may be more complex than the initial application (e.g., the crop simulation model) that requires soil temperature data and may provide more accurate answers than the application may require. The third approach, the hybrid or empirical-mechanistic models, combines empirical and mechanistic approaches and attempts to overcome weaknesses inherent in each of these approaches. Hybrid models (e.g., Wierenga and deWit 1970; Hanks et al. 1971) quantify the main processes such as soil heat conduction and land surface cover that govern soil temperature in a mechanistic way, while other processes such as latent and sensible heat fluxes are calculated with statistical algorithms describing these processes.

Kang et al. (2000) developed and evaluated a hybrid model for simulating soil temperature at 0.1 m depth in forest soils in Korea. The model combined the empirical relationship between air temperature and soil temperature (Zheng et al. 1993) and a solution to the Fourier heat transfer equation (Campbell 1985) and incorporating the effects of topography, canopy cover, and surface litter. This model was successfully applied in a simulation study of soil CO_2 concentration, where soil CO_2 concentration was a function of soil temperature, in a forested catchment in Virginia (Welsch and Hornberger 2004), suggesting a general applicability of the model. Soil temperature models of this nature are important because they can easily be incorporated into existing crop simulation models to expand their scope or be combined with other algorithms to form a new simulation model to address applied problems.

Agricultural and deciduous forest systems share some common factors. Canopy cover is not constant during a year, and soil moisture and temperature changes depending on soil texture, leaf area index (LAI), and water input. These two systems differ with respect to litter, which can be modified in an agricultural system depending on tillage and the amount of residues left by the preceding crop. Litter can have great impact on soil water and temperature dynamics (e.g., Grant et al. 1995).

The objectives of this study were to evaluate, and if necessary modify, the hybrid soil temperature model developed by Kang et al. (2000) at several depths in an agroecosystem consisting of a no-till corn (*Zea mays* L.) and soybean (*Glycine max* [L.] Merr.) rotation, and in a tallgrass prairie in the Great Plains.

MATERIALS AND METHODS

Model Description

There is a coupling between soil and air temperature variations that can be used to develop a hybrid soil temperature model. Zheng et al. (1993) used this coupling concept to develop a model that simulated mean daily values of soil temperature (T_j) at a depth of 0.1 m on day j from the 11-day average of mean daily air temperatures ($A11_j$) for a vegetated surface. In particular, $A11_j$ is the average of daily air temperature from day $j - 10$ to day j. When $A11_j > T_{j-1}$, then

$$T_j = T_{j-1} + [A11_j - T_{j-1}] \cdot M \cdot \exp(-K \cdot LAI), \quad (1)$$

where T_j and T_{j-1} are the mean soil temperatures on the current and previous day, respectively, M is an empirical coefficient, K is the extinction coefficient, and LAI is the leaf area index. When $A11_j < T_{j-1}$, then

$$T_j = T_{j-1} + [A11_j - T_{j-1}] \cdot M. \quad (2)$$

Zheng et al. (1993) provided no explanations for why these two equations were conditioned upon the values

of $A11_j$ and T_{j-1}. The following reasons are proposed to explain the need for these two equations. When $A11_j > T_{j-1}$ (equation 1), it was assumed that heat flow into the soil was positive. The temperature gradient from air to the soil, based upon heating from the sun, will be a function of canopy *LAI*. In equation 2, ($A11_j < T_{j-1}$), the soil releases heat (the heat flow into the soil is negative), and the leaf area index does not influence this negative heat flow.

Kang et al. (2000) developed a model to simulate soil temperature in a forest system based upon the model of Zheng et al. (1993). Instead of using the 11-day average of mean daily air temperatures ($A11_j$), they used the actual mean daily air temperature, A_j, and they added the leaf area index equivalent of the leaf litter that covers the soil surface, LitterAI, and a dampening ratio term to replace the empirical coefficient *M* in the Zheng et al. (1993) model. The dampening ratio term (*DR*) on day *j* was defined as

$$DR_j = \exp[-z \, (\pi \, / \, \kappa_{sj}\, p)^{0.5}], \qquad (3)$$

where $p = 365 \times 24 \times 60 \times 60$ was the period of annual temperature variation (s), κ_{sj} was the thermal diffusivity (m^2/s) of the soil at depth (*z*) on day *j*, and π has its usual meaning. Kang et al. (2000) used a constant κ_s for the forest soils in their study. In this study, κ_{sj} was computed on a daily basis following the procedures given in Campbell (1985) and Bristow (1998):

$$\kappa_{sj} = \lambda_j \, / \, p_{cj}, \qquad (4)$$

where λ_j is the thermal conductivity and p_{cj} is the volumetric heat capacity for day *j*. Volumetric heat capacity was estimated following Campbell (1985):

$$p_{cj} = (p_c)_m(1-P)+(p_c)_w \, \theta_{vj}, \qquad (5)$$

where *m* indicates the average soil mineral fraction and *P* was the total porosity, θ_{vj} is the volumetric water content for day *j*, $(p_c)_m$ and $(p_c)_w$ are the volumetric heat capacity of minerals and water, respectively.

Thermal conductivity λ_j was calculated as (Campbell 1985):

$$\lambda_j = A + B\theta_{vj} - (A-D)exp[-(C\theta_{vj})^E], \qquad (6)$$

where *A*, *B*, *C*, *D*, and *E* are soil dependent coefficients related to soil properties (Campbell 1985):

$$A=(0.57+1.73\phi_q+0.93\phi_{rm})/(1-0.74\phi_q-0.49\phi_{rm})-2.8\phi_s(1-\phi_s)$$

$$B = 2.8\phi_s$$

$$C = 1+2.6/ \, (m_c^{0.5})$$

$$D = 0.03+0.7\phi_s^2$$

$$E = 4 \, ,$$

where ϕ is volume fraction and m_c is the clay fraction. The subscripts *q*, *rm*, and *s* indicate quartz, minerals other than quartz, and total solids, respectively. We assumed that for a given field site the soil texture and soil mineralogy will be uniform for each soil layer, so the thermal conductivity, volumetric heat capacity, thermal diffusivity, and the dampening ratio are determined only by the variations in soil water content, θ_{vj}.

The mean daily soil temperature at depth *z* (m), $T_j(z)$, was estimated by the following relationships (Kang et al. 2000). As in Zheng et al. (1993), when $A_j > T_{j-1}(z)$, then

$$T_j(z) = T_{j-1}(z)+[A_j - T_{j-1}(z)] \cdot DR_j \cdot \exp[-K \, (LAI_j+LitterAI_j \,)]. \quad (7)$$

And when $A_j < T_{j-1}(z)$,

$$T_j(z) = T_{j-1}(z)+[A_j - T_{j-1}(z)] \cdot DR_j \cdot \exp[-K \, (LitterAI_j \,)]. \qquad (8)$$

Though Kang et al. (2000) used the model to simulate the soil temperature at the 0.1 m depth, the inclusion of the dampening ratio (*DR*) made it possible to simulate the soil temperature at various depths. In Zheng et al. (1993), simulating soil temperatures at different depths required changing the value of the empirical coefficient *M* for each depth. Following Zheng et al. (1993), Kang et al. (2000) assumed that the soil temperature would not fall below 0ºC, that is, when $T_j(z)$ was less than 0ºC, it was set to 0ºC. In the forest sites of their studies, Kang et al. (2000) found that soil temperatures at 0.1 m depth rarely went below 0ºC.

If this soil temperature model is used as part of a plant simulation model, the 0ºC assumption is not a limitation for spring-planted crops or for annual plants that start to grow in the spring and become dormant in the fall. It is also not a limitation for fall-planted crops, as the crops will be dormant during the time of the year when the soil temperatures are less than zero. In the application of Welsch and Hornberger (2004), where the objective was to simulate the spatial and temporal soil CO_2 concentration, this assumption was not a limitation.

Model Modification

Two contrasting vegetative sites were used in this study: a tallgrass prairie and an agroecosystem (a corn-soybean rotation). More details of these sites are given in the "Field Observations and Assumptions" section. For the tallgrass prairie site, the extinction coefficient was assumed to be the same as the litter, thus equations 7 and 8 could be used. However, for the agroecosystem, it was assumed that the extinction coefficients of LAI (K_{lai}) and LitterAI (K_{litter}) were different. Equations 7 and 8 were thus modified in the following way:

When $A_j > T_{j-1}(z)$, then

$$T_j(z) = T_{j-1}(z) + [A_j - T_{j-1}(z)] \cdot DR_j \cdot \exp[-K_{lai} LAI_j - K_{litter} LitterAI_j]. \quad (9)$$

And when $A_j < T_{j-1}(z)$, then

$$T_j(z) = T_{j-1}(z) + [A_j - T_{j-1}(z)] \cdot DR_j \cdot \exp[-K_{litter} LitterAI_j]. \quad (10)$$

This study will evaluate the modified hybrid soil temperature model in an agroecosystem and a tallgrass prairie. For the agroecosystem, the modified and the original models will also be compared and evaluated. Soil temperatures were simulated on a daily time step. The required input data to run the model (equations 7 and 8 or equations 9 and 10) were daily average air temperature, leaf area index (LAI), and litter-equivalent leaf area index (LitterAI). The initial value of the soil temperature $T_0(z)$ was the observed air temperature one day prior to the starting date. To evaluate the sensitivity of the model to variations in soil thermal diffusivity and water content the following data were also required: the soil bulk density, quartz fraction, clay fraction, and soil water content at each depth. These soil parameters are used to compute the soil thermal conductivity and volumetric heat capacity (equations 5 and 6), which are then used to compute the soil thermal diffusivity and the dampening ratio (equations 3 and 4).

Field Observations and Assumptions

For the agroecosystem, data used to develop empirical relationships and model verification came from a three-year study (2002–2004) in a corn and soybean rotation under no tillage with one site irrigated (41º9'53.5"N, 96º28'12.3"W) and the other site rainfed (41º10'46.8"N, 96º26'22.7"W) located at Mead, NE. These data were collected as part of the Carbon Sequestration Program at the

University of Nebraska (http://csp.unl.edu/public/). The soils at both sites were deep silty clay loams consisting of four soil series: Yutan (fine-silty, mixed, mesic Mollic Hapludalfs), Tomek (fine, smectitic, mesic Pachic Argial-bolls), Filbert (fine, smectitic, mesic Vertic Argialbolls), and Filmore (fine, smectitic, mesic Vertic Argialbolls). The sites were disk plowed in 2001 to incorporate residues from previous crops and fertilizers, and to homogenize the first 0.1 m of the soil. Since 2001 all sites have been under this no-till rotation, and crop management is based on the standard best-management practices prescribed for production-scale corn systems (http://csp.unl.edu/public/). Corn crops were sown on the second week of May in 2001 and 2003, and harvested in late October. Soybean crops were sown in mid-May in 2002 and 2004 and harvested in early October.

Soil temperature was measured hourly with YSI 44004 precision thermistors (http://www.omega.com/pptst/TJ36-44004.html) at 0.1, 0.2, 0.3, and 0.5 m depths. The soil temperatures at 02, 0.3, and 0.5 m depths were measured between the rows, whereas the soil temperature at 0.1 m depth was measured both in the row and between the rows. The soil temperature at the 0.1 m depth used in this study was the arithmetic mean of these two values. The daily average soil temperature was calculated using the 24 hourly values to compare with the simulated data. Hourly soil water content was measured in the row at 0.1, 0.25, and 0.5 m depths using ThetaProbe ML2x (www.dynamax.com). The average of soil water content was used for the calculation of thermal diffusivity (Campbell 1985) at each depth, and the soil water content at 0.25 m was used for calculating soil temperatures at 0.2 and 0.3 m depths.

Daily maximum and minimum air temperatures were recorded directly at the irrigated site at 2 m height. For the rainfed site, the daily maximum and minimum air temperatures recorded at the Mead Turf Farm weather station (41º10'N, 96º28'W) were used. This station was located less than 2 km distance from rainfed site. Daily mean air temperature was calculated as the arithmetic mean of the maximum and minimum air temperatures.

Leaf area index (LAI) was measured every 10 to 14 days during the growing season and plotted against day of year for each site. Polynomial equations were developed to interpolate the daily leaf area index values. LitterAI was not measured in the field. It was assumed that soybeans made no long-term contribution to the LitterAI because these leaves decayed relatively quickly. The contribution from cornstalks to the LitterAI was estimated as 0.4 of the maximum corn leaf area index when the corn was grown.

It was further assumed, based on visual observations, that LitterAI was accumulated after the corn harvest and remained constant until the next corn harvest. Although LitterAI increased with time, it was further assumed that the effects of LitterAI on soil temperatures remained constant for LitterAI values above 2.0; this value was denoted as the maximum LitterAI (LitterAI$_{max}$). Quantifying the relationships between surface litter and temperature, water, carbon, and nitrogen cycling is not a trivial task (Guerif et al. 2001; Findeling et al. 2007).

The tallgrass prairie site and associated measurements have been described in Suyker and Verma (2001), Suyker et al. (2003), and Hanan et al. (2005). The site was located near Shidler, OK (36º56'N, 96º41'W). Most of the grasses at this site were warm-season C$_4$ grasses dominated by little bluestem (*Schizachyrium scoparium* [Michx.] Nash), blue grama (*Bouteloua graciliis* [H.B.K.] Lag. ex Steud.), big bluestem (*Andropogon gerardii* Vitman), and indiangrass (*Sorghastrum nutans* [L.] Nash), which reached peak activity and biomass in the mid- to late summer. The soil was a silty clay loam of the Wolco-Dwight complex (thermic Pachic Argiustolls and mesic Typic Natrustolls) with 1 to 2 m layers of dense clay below 0.6 m. Soil temperatures were measured at 0.1, 0.2, 0.3, and 0.45 m below the soil surface during the period from September 1997 to March 2000. Soil temperature data at the 0.2 and 0.45 m depths contained too many questionable values, and data from these depths were not included in the data analysis. Quality control of questionable temperature values at the 0.1 and 0.3 m depths were done by manual inspection of the data. The total leaf area index was measured weekly or biweekly and daily values were obtained from polynomial relationships. The prairie was burned each spring (either late March or early April) and the LitterAI was small, estimated to be 0.28 based on the information provided by Stubbendieck (2008 personal communication). For the purposes of this simulation study, the active growing period was from May 1 through September 30, while the entire year consisted of the period from May 1 to the end of February.

For the agroecosystem, the extinction coefficients for leaf area index were 0.6 and 0.65 for soybeans and corn, respectively (Flenet et al. 1996). The value of the extinction coefficient for the LitterAI was fixed at 0.86 (Sauer et al. 1997). This value, higher than that considered for the extinction coefficients associated with leaf area index, takes into account the irregular distribution of litter on the soil surface, which affects soil temperature attenuation. For the tallgrass prairie site, the extinction coefficients for LAI and LitterAI were taken as 0.44

based on values given in Suyker and Verma (2001) and Kiniry et al. (2007).

Statistical Analysis

The root mean square error (RMSE), mean absolute error (MAE), and mean bias error (MBE) were used to test the goodness of fit of the hybrid soil temperature model and observed data (Willmott 1982):

$$RMSE = \left[N^{-1} \sum_{i=1}^{N} \left(S_i - O_i \right)^2 \right]^{0.5} \qquad (11)$$

$$MAE = N^{-1} \sum_{i=1}^{N} \left| S_i - O_i \right| \qquad (12)$$

$$MBE = N^{-1} \sum_{i=1}^{N} \left(S_i - O_i \right), \qquad (13)$$

where N is the number of observations, S is the simulated soil temperature, and O is the observed soil temperature.

The robustness of the simulation was tested by decomposing the mean square error (MSE) into two components: the systematic (MSE$_s$) and unsystematic (MSE$_u$) errors (Willmott 1982):

$$MSE_s = N^{-1} \sum_{i=1}^{N} \left(\widehat{S}_i - O_i \right)^2 \qquad (14)$$

$$MSE_u = N^{-1} \sum_{i=1}^{N} \left(S_i - \widehat{S}_i \right)^2, \qquad (15)$$

where $\widehat{S}$ was estimated by the linear regression between simulated (S) and observed (O) soil temperatures. In a robust simulation model, systematic mean square error would be very small, while unsystematic mean square error would approach the mean square error.

RESULTS

Figure 1 compares the simulated and observed daily soil temperature for the four depths for the independent data set (rainfed site) over the three years in the agroecosystem. Both the magnitude and the temporal variations of the simulated soil temperature follow the observation values, especially for the 0.1, 0.2, and 0.3 m depths. Simulations for 0.2 and 0.3 m depths were better than those for the 0.1 and 0.5 m depths based on root mean square error values (1.41º and 1.49ºC compared to 1.61º and 2.05ºC,
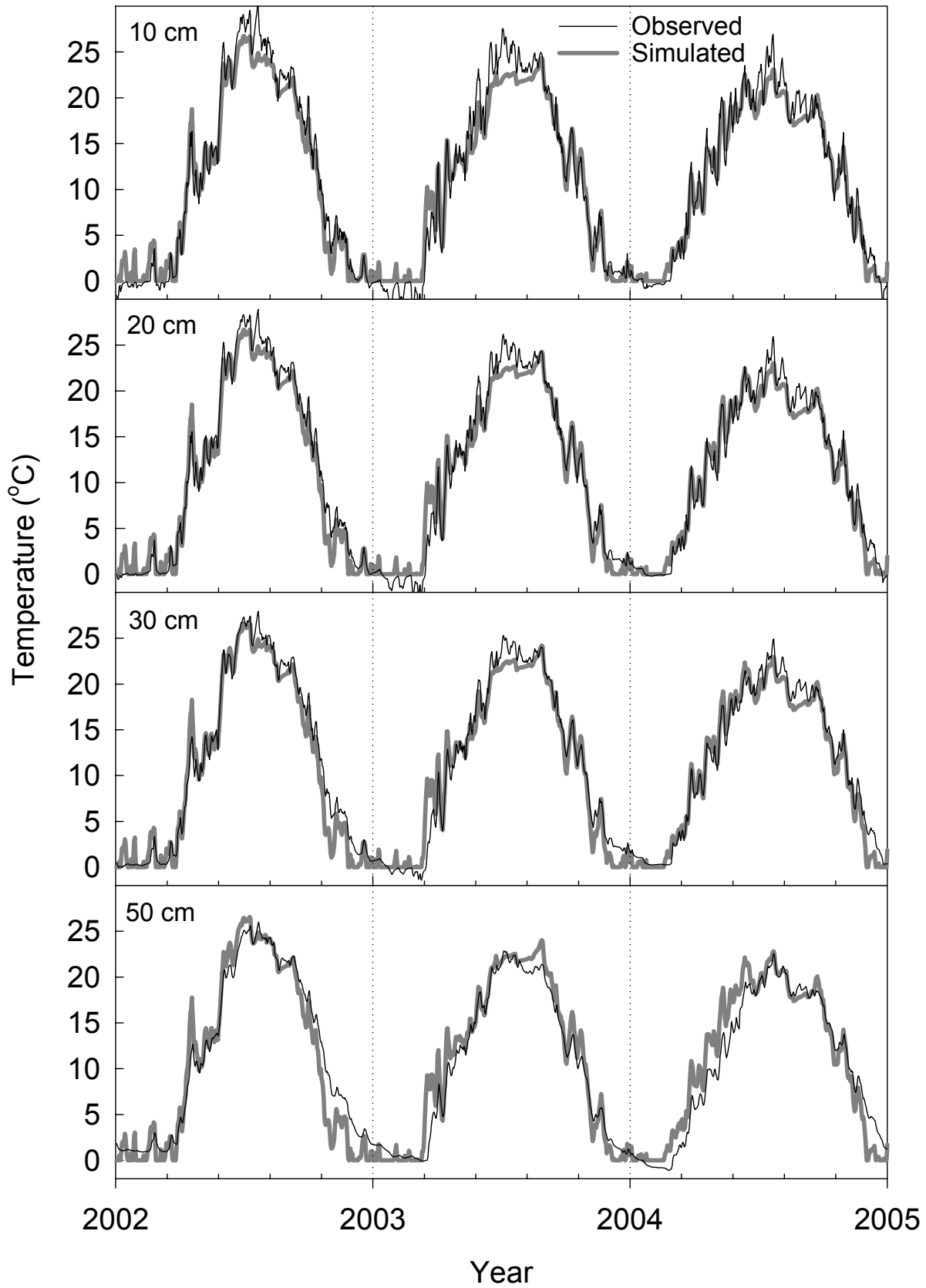
Figure 1. Observed and simulated daily soil temperatures (°C) for the independent data set (rainfed site) at Mead, NE.

TABLE 1
ROBUSTNESS OF THE SIMULATED SOIL TEMPERATURES FOR THE INDEPENDENT DATA SET
(RAINFED SITE) AT MEAD, NE, USING THE MODIFIED MODEL (EQUATIONS 9 AND 10)

Depth (m)	Root mean square error (RMSE)	Systematic mean square error (MSE_s)	Unsystematic mean square error (MSE_u)	Mean absolute error (MAE)	Mean bias error (MBE)
0.1	1.61 (1.76)	1.06 (1.64)	1.54 (1.46)	1.24 (1.37)	-0.35 (-0.94)
0.2	1.41 (1.39)	0.47 (0.79)	1.50 (1.14)	1.06 (1.09)	-0.37 (-0.69)
0.3	1.49 (1.31)	0.27 (0.41)	1.94 (1.30)	1.11 (1.02)	-0.44 (-0.51)
0.5	2.05 (1.94)	0.09 (0.64)	4.11 (3.13)	1.53 (1.44)	0.22 (0.70)

Note: Numbers in parentheses are values for the growing season April 1 to October 31.

respectively; Table 1). The model tended to under-predict soil temperature for the first three depths, ranging from -0.35º to -0.44ºC and to over-predict at the 0.5 m depth, 0.22ºC, as indicated by the mean bias error. Similar results were obtained for the growing season data (April 1 to October 31). The systematic mean square error values were less than unsystematic mean square error values, especially at the 0.2, 0.3, and 0.5 m depths (0.47, 0.27, and 0.09, respectively), indicating the robustness of the model in simulating soil temperatures.

A series of sensitive analyses were carried out on the dependent data set (irrigated site) and the tallgrass prairie site to identify the model performance with the change of selected input parameters. The thermal diffusivity and soil water content were evaluated, using a wide range of values, because these inputs are site-specific parameters.

For the agroecosystem, the modified hybrid model (equations 9, 10) was not sensitive to the influence of thermal diffusivity on soil temperature based on RMSE values. The soil thermal diffusivity was varied from 2.0 to 10×10^{-7} m^2/s, which resulted in changes of RMSE that were less than 0.1ºC for 0.1 to 0.3 m depth and 0.2ºC changes for the 0.5 m depth.

The modified model was also not sensitive to the variations of soil water content. Figure 2 illustrates the influence of soil water content on soil thermal conductivity, volumetric heat capacity and the soil thermal diffusivity using the soil water content at 0.1 m depth at rainfed site (independent data set) during 2002-2004. The soil water content varied from 0.1 to 0.35 m^3/m^3. The soil thermal conductivity varied from 0.4 to 1.3 W m^{-1} ºC^{-1}, and the volumetric heat capacity of the soil at the 0.1 m depth varied from 1.6 to 2.6×10^6 J m^{-3} ºC^{-1}. The soil thermal diffusivity, calculated by the algorithms in Campbell

(1985; see also equation 4), varied from 3.0 to 5.5×10^{-7} m^2/s based on the changing values of the volumetric water content. However, these large variations in the soil water content, the soil thermal conductivity, volumetric heat capacity, and the soil thermal diffusivity resulted in very small changes in the dampening ratio. As shown in Figure 2, variations of dampening ratio ranged from 0.94 to 0.96, which explains the insensitivity of the model simulations to the variations of soil water content.

Surface litter can have great impact on soil water and temperature dynamics (e.g., Grant et al. 1995; Guerif et al. 2001). Kang et al. (2000) also found that the hybrid soil model is very sensitive to the variations of LitterAI. In this study, quantification of the effects of the LitterAI on soil temperature was not carried out because published data on LitterAI were used; no field measurements of LitterAI were available at either site.

For the tallgrass prairie, as in the agroecosystem, variations in thermal diffusivity and soil moisture did not greatly influence the simulation of soil temperatures. Figure 3 compares the simulated and observed daily soil temperatures at 0.1 and 0.3 m depths for tallgrass prairie from September 1997 to March 2000. Both the magnitude and the temporal variations of the simulated soil temperatures closely follow the observed values. The root mean square error values (Table 2) were slightly greater than for the agroecosystem, varying between 1.95º and 2.26ºC for the entire season, while the values for the active growing season, 1.44º and 1.81ºC, were similar to those in Table 2. The mean absolute error values over the entire season were slightly greater than those of the agroecosystem (1.70ºC at 0.3 m [Table 2] vs. 1.32ºC at 0.5 m [Table 1]), while the values over the warm season fell within the values found in Table 1. The mean bias error values were
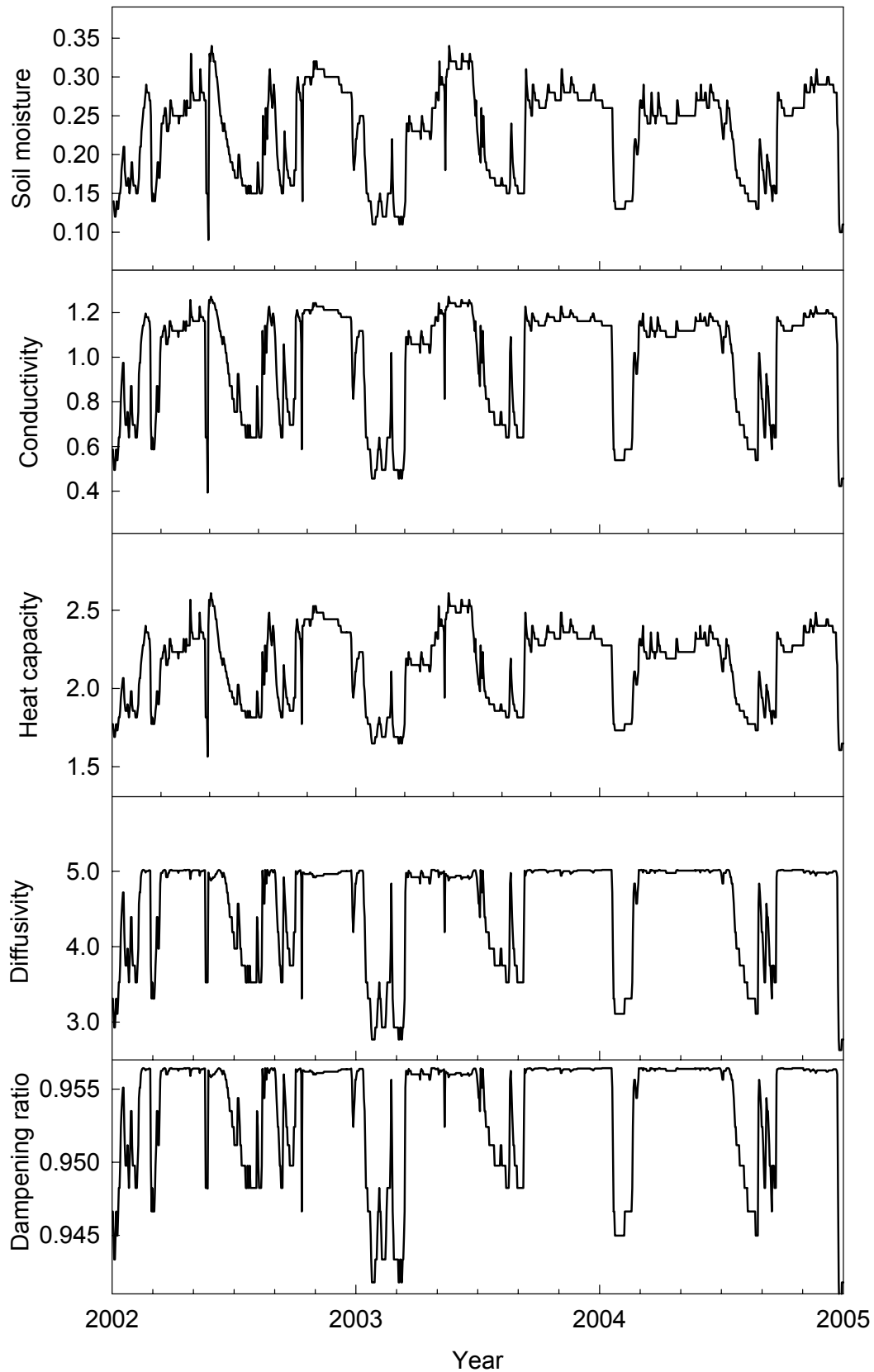
Figure 2. Variations of the soil water content, soil conductivity (W m^{-1} °C^{-1}), volumetric heat capacity ($\times$ 10^6 J m^{-3} °C^{-1}), soil thermal diffusivity ($\times$ 10^{-7} m^2 s^{-1}), and dampening ratio at the 0.1 m depth for the independent data set (rainfed site) at Mead, NE.
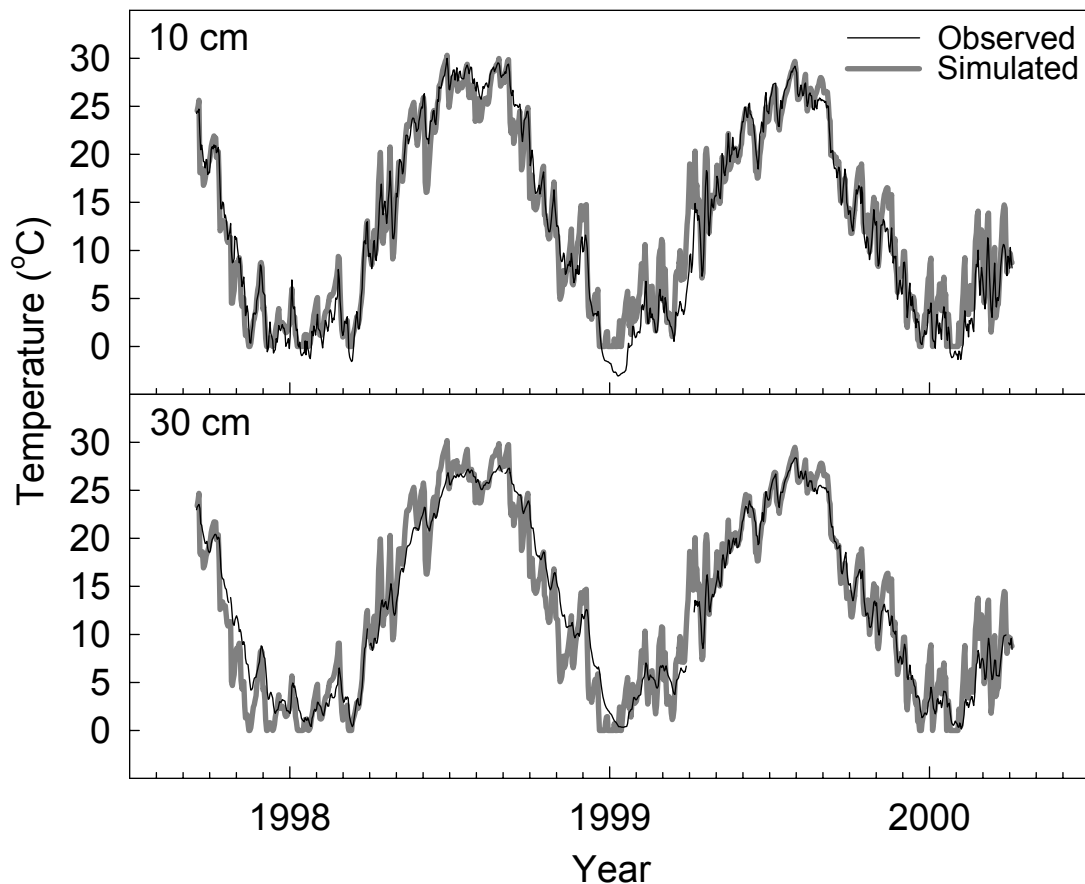
Figure 3. Observed and simulated daily soil temperature (°C) for the tallgrass prairie.

similar for both ecosystems, as were the mean square error values. These latter values indicated that the systematic errors were small with respect to the random errors in hybrid soil temperature model.

DISCUSSION

In evaluating their results for forest soils, Kang et al. (2000) found that the mean absolute error varied from 0.96° to 1.48°C at the 0.1 m depth. For the calendar year, the mean absolute error in Table 1 varied from 1.06° to 1.53°C, while the mean absolute error for the growing season varied from 1.02° to 1.44°C. Thus, in terms of this statistic, simulation results between these two studies were similar.

One of the major differences between this modified model and the original model by Kang et al. (2000) is that the modified model uses a different extinction coefficient for ground litter, based on the observations made by Sauer et al. (1997). The soil temperatures simulated by

the original model (equations 7 and 8) are summarized in Table 3. Comparing the data in Tables 1 and 3 indicates that the modified model better simulates soil temperature variations, especially at the 0.2–0.5 m depths. At these depths, the root mean square error and the mean absolute error of the modified model are about 0.1°–0.3°C smaller than the original model.

Compared to the agroecosystem, a possible reason for the slightly larger root mean square error values (mean absolute error values) in tallgrass prairie determined over the entire growing season may be that this site contains several grass species growing simultaneously during the entire year in a non-uniform pattern. In contrast, the agroecosystem has a relatively uniform pattern of vegetation, and after harvest the residue on the soil surface forms another relatively uniform pattern.

A sensitivity analysis indicated that this model was not sensitive to the variations of soil water content and thermal diffusivity in the agroecosystem and the tallgrass prairie. The data in Figure 2 suggests that even though

TABLE 2
ROBUSTNESS OF THE SIMULATED SOIL TEMPERATURES
FOR THE OKLAHOMA TALLGRASS PRAIRIE SITE

Depth (m)	Root mean square error (RMSE)	Systematic mean square error (MSE_s)	Unsystematic mean square error (MSE_u)	Mean absolute error (MAE)	Mean bias error (MBE)
0.1	1.95 (1.44)	1.00 (0.36)	2.80 (1.71)	1.48 (1.07)	0.24 (-0.59)
0.3	2.26 (1.81)	0.03 (0.17)	5.05 (3.11)	1.70 (1.35)	-0.08 (0.40)

Note: Numbers in parentheses are values for the active growing season May 1 to September 30.

TABLE 3
ROBUSTNESS OF THE SIMULATED SOIL TEMPERATURES FOR THE INDEPENDENT DATA SET
(RAINFED SITE) AT MEAD, NE, USING THE ORIGINAL MODEL
(KANG ET AL. 2000) (EQUATIONS 7 AND 8)

Depth (m)	Root mean square error (RMSE)	Systematic mean square error (MSE_s)	Unsystematic mean square error (MSE_u)	Mean absolute error (MAE)	Mean bias error (MBE)
0.1	1.68 (1.75)	1.00 (1.41)	1.81 (1.65)	1.28 (1.38)	-0.33 (-0.95)
0.2	1.62 (1.56)	0.45 (0.69)	2.18 (1.74)	1.24 (1.25)	-0.34 (-0.70)
0.3	1.82 (1.66)	0.26 (0.37)	3.03 (2.38)	1.36 (1.29)	-0.41 (-0.52)
0.5	2.44 (2.39)	0.08 (0.67)	5.87 (5.03)	1.82 (1.77)	0.25 (0.70)

Note: Numbers in parentheses are values for the growing season April 1 to October 31.

the soil water content varied about 70% and the thermal diffusivity varied about 40% between wet and dry periods, their impact on dampening ratio was less than 2%. Therefore, the impact of soil moisture and thermal diffusivity on soil temperature variations was very small when using this hybrid model. For this application, setting the thermal diffusivity at a constant value, as was done in Kang et al. (2000), was valid. These results are consistent with those of Lin et al. (2003), who used a different soil temperature model. Hu and Feng (2005) analyzed observations at 0.8 m below surface at more than 240 stations over Eurasia; they found that the soil temperature at this depth was not sensitive to the change of precipitation or the soil water content.

The above results indicate a weak influence of soil moisture on soil temperature. However, some modeling and observational studies (e.g., Luo et al. 1992; Kane et al. 2001) have shown that soil moisture can strongly influence the surface energy balance and soil temperature. These differences are likely caused by different temporal resolutions of soil temperature. Luo et al. (1992) and Kane et al. (2001) analyzed the influence of rainfall and soil moisture on hourly soil temperature variations, which was dominated by a diurnal temperature signal. Changes in soil moisture (water uptake by plants and soil evaporation or drainage after rain) can greatly alter heat storage in the soil column and hence influence hourly soil temperature variations. In this study and the studies of Lin et al. (2003) and Hu and Feng (2005), daily or monthly soil temperature variations were analyzed, which were dominated by an annual temperature signal. An observational study (Smerdon et al. 2003) suggested that the annual soil temperature signal was dominated by conductive heat transport in the soil which was not sensitive to rainfall changes (a surrogate of soil moisture).

Advantages of this model are (1) its relative simplicity using easily available data, (2) it can be incorporated into a larger model where soil temperatures are required, and (3) the impact of surface litter on soil temperatures can be incorporated into existing models (assuming there is a separate model for litter dynamics). Given the applied nature of this hybrid model, it would be well suited to simulate soil temperatures in the first 50 cm of soil over a vegetated surface for processes related to soil respiration, soil organic matter decomposition, and soil-borne pests.

ACKNOWLEDGMENTS

REFERENCES

Boone, R.D., K.J. Nadelhoffer, J.D. Canary, and J.P. Kaye. 1998. Roots exert a strong influence on the temperature sensitivity of soil respiration. *Nature* 396:570–72.

Bristow, K.L., 1998. Measurement of thermal properties and water content of unsaturated sandy soil using dual-probe heat-pulse probes. *Agricultural and Forest Meteorology* 89:75–84.

Campbell, G.S. 1985. Soil physics with BASIC: Transport models for soil-plant systems. Elsevier, New York.

Findeling, A., P. Garnier, F. Coppens, F. Lafolie, and S. Recous. 2007. Modelling water, carbon and nitrogen dynamics in soil covered with decomposing mulch. *European Journal of Soil Science* 58:196–206.

Flenet, F., J.R. Kiniry, J.E. Board, M.E. Westgate, and D.C. Reicosky. 1996. Row spacing effects on light extinction coefficients of corn, sorghum, soybean, and sunflower. *Agronomy Journal* 88:185–90.

Forbes, P.J., K.E. Black, and J.E. Hooker. 1997. Temperature-induced alteration to root longevity in *Lolium perenne*. *Plant and Soil* 190:87–90.

Gavito, M.E., P. Schweiger, and I. Jakobsen. 2003. P uptake by arbuscular mycorrhizal hyphae: Effect of soil temperature and atmospheric CO_2 enrichment. *Global Change Biology* 9:106–16.

Grant, R.F., R.C. Izaurralde, and D.S. Chanasyk. 1995. Soil temperature under different surface managements: Testing a simulation model. *Agricultural and Forest Meteorology* 73:89–113.

Guerif, J., G. Richard, C. Durr, J.M. Machet, S. Recous, and J. Roger-Estrade. 2001. A review of tillage effects on crop residue management, seedbed conditions, and seedling establishment. *Soil and Tillage Research* 61:13–32.

Hanan, N.P., J.A. Berry, S.B. Verma, E.A. Walter-Shea, A.E. Suyker, G.G. Burba, and A.S. Denning. 2005. Testing a model of CO_2, water and energy exchange in Great Plains tallgrass prairie and wheat ecosystems. *Agricultural and Forest Meteorology* 131:163–79.

Hanks, R.J., D.D. Austin, and W.T. Ondrechen. 1971. Soil temperature estimation by a numerical method. *Soil Science Society of America Journal* 35:665–67.

Hendrick, R.L., and K.S. Pregitzer. 1997. The relationship between fine-root demography and the soil environment in northern hardwood forests. *Ecoscience* 4:99–105.

Hu, Q., and S. Feng. 2005. How have soil temperatures been affected by the surface temperature and precipitation in the Eurasian continent? *Geophysical Research Letter* 32:L14711, doi:10.1029/2005GL023469.

Kane D.L., K.M. Hinkel, D.J. Goering, L.D. Hinzman, and S.I. Outcalt. 2001. Non-conductive heat transfer associated with frozen soils. *Global and Planetary Change* 29:275–92.

Kang, S., S. Kim, S. Oh, and D. Lee. 2000. Predicting spatial and temporal patterns of soil temperature based on topography, surface cover, and air temperature. *Forest Ecology and Management* 136:173–84.

Kemp P.R., J.M. Cornelius, and J.F. Reynolds. 1992. A simple model for predicting soil temperatures in desert ecosystems. *Soil Science* 153:280–87.

King, J.S., R.S. Thomas, and B.R. Strain. 1997. Morphology and tissue quality of seedling root systems of *Pinus teada* and *Pinus ponderosa* as affected by varying CO_2, temperature, and nitrogen. *Plant and Soil* 195:107–19.

Kiniry, J.R., B.L. Burson, G.W. Evers, J.R. Williams, H. Sanchez, C. Wade, J.W. Featherston, and J. Greenwade. 2007. Coastal bermudagrass, bahiagrass, and native range simulation at diverse sites in Texas. *Agronomy Journal* 99:450–61.

Kirschbaum, M.U.F., 1995. The temperature dependence of soil organic matter decomposition, and the effect of global warming on soil organic C storage. *Soil Biology and Biochemistry* 27:753–60.

Lin, X., J.E. Smerdon, A.W. England, and H.N. Pollack. 2003. A model study of the effects of climatic precipitation changes on ground temperatures. *Journal of Geophysical Research* 108(D7):4230, doi:10.1029/2002JD002878.

Luo, Y., R.S. Loomis, and T.C. Hsiao. 1992. Simulation of soil temperature in crops. *Agricultural and Forest Meteorology* 61:23–28.

Mellander, P.E., K. Bishop, and T. Lundmark. 2004. The influence of soil temperature on transpiration: A plot scale manipulation on a young Scots pine stand. *Forest Ecology and Management* 195:15–28.

Parton, W.J., D.S. Schimel, C.V. Cole, and D.S. Ojima. 1987. Analysis of factors controlling soil organic matter levels in Great Plains grasslands. *Soil Science Society of America Journal* 51:1173–79.

Parton, W.J., J.W.B. Stewart, and C.V. Cole. 1988. Dynamics of C, N, P and S in grassland soils: A model. *Biogeochemistry* 5:109–31.

Raich, J.W., and W.H. Schlesinger. 1992. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus* 44B:81–99.

Sauer, T.J., J.L. Hatfield, and J.H. Prueger. 1997. Overwinter changes in radiant energy exchange of a corn residue-covered surface. *Agricultural and Forest Meteorology* 85:279–87.

Smerdon J.E., H.N. Pollack, J.W. Enz, and M.J. Lewis. 2003. Conduction-dominated heat transport of the annual temperature signal in soil. *Journal of Geophysical Research* 108:2431, doi:10.1029/2002JB002351.

Smirnova, T.G., J.M. Brown, and S.G. Benjamin. 1997. Performance of different soil model configurations in simulating ground surface temperature and surface fluxes. *Monthly Weather Review* 125:1870–84.

Suyker, A.E., and S.B. Verma. 2001. Year-round observations of the net ecosystem exchange of carbon dioxide in a native tallgrass prairie. *Global Change Biology* 7:279–89.

Suyker, A.E., S.B. Verma, and G.G. Burba. 2003. Interannual variability in net CO_2 exchange of a native tallgrass prairie. *Global Change Biology* 9:255–65.

Welsch, D.L., and G.M. Hornberger. 2004. Spatial and temporal simulation of soil CO_2 concentrations in a small forested catchment in Virginia. *Biogeochemistry* 71:415–36.

Wierenga, P.J., and C.T. deWit. 1970. Simulation of heat flow in soils. *Soil Science Society of America Proceedings* 34:845–48.

Willmott, C.J. 1982. Some comments on the evaluation of model performance. *Bulletin of the American Meteorological Society* 63:1309–13.

Zheng D., E.R. Hunt, Jr., and S.W. Running. 1993. A daily soil temperature model based on air temperature and precipitation for continental applications. *Climate Research* 2:183–91.