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Modeling Soil Moisture and Surface Flux Variability with an Untuned Land Surface Scheme: A Case Study from the Southern Great Plains 1997 Hydrology Experiment

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ABSTRACT

The Parameterization for Land–Atmosphere–Cloud Exchange (PLACE), a typical surface–vegetation–atmosphere transfer (SVAT) parameterization, was used in a case study of a 2500 km² area in southwestern Oklahoma for 9–16 July 1997. The research objective was to assess PLACE's simulation of the spatial variability and temporal evolution of soil moisture and heat fluxes without optimization for this case study. Understanding PLACE's performance under these conditions may provide perspective on results from more complex coupled land–atmosphere simulations involving similar land surface schemes in data-poor environments. Model simulations were initialized with simple initial soil moisture and temperature profiles tied to soil type and forced by standard meteorological observations. The model equations and parameters were not adjusted or tuned to improve results.

For surface soil moisture, 5- and 10-cm soil temperature, and surface fluxes, the most accurate simulation (5% error for soil moisture and 2 K for 5- and 10-cm soil temperature) occurred during the 48 h following heavy rainfall on 11 and 15 July. The spatial pattern of simulated soil moisture was controlled more strongly by soil texture than was observed soil moisture, and the error was correlated with rainfall. The simplifications of the subsurface soil moisture, soil texture, and vegetation cover initialization schemes and the uncertainty in the rainfall data (>10%) could account for differences between modeled and observed surface fluxes that are on the order of 100 W m⁻² and differences in soil moisture that are greater than 5%. It also is likely that the soil thermal conductivity scheme in PLACE damped PLACE's response to atmospheric demand after 13 July, resulting in reduced evapotranspiration and warmer but slower-drying soils. Under dry conditions, the authors expect that SVATs such as PLACE that use a similar simple initialization also would demonstrate a strong soil texture control on soil moisture and surface fluxes and limited spatial variability.

1. Introduction

Recognition of the importance of land surface processes in the prediction of weather and climate has led to efforts to incorporate improved land surface parameterizations into atmospheric models to represent land–atmosphere interaction (e.g., Avissar and Pielke 1989; Noilhan and Planton 1989; Dickinson et al. 1993; Fa-

miglietti and Wood 1994; Wetzel and Boone 1995; Bonan 1996; Sellers et al. 1996). For complex coupled land–atmosphere simulations to have predictive value, land surface parameterizations must be robust physically, that is, capable of producing reasonable results in situations in which only standard soil and vegetation parameter datasets are available. Outside of major field experiments, auxiliary datasets such as spatial patterns of surface soil moisture often are not available to support model tuning and validation. Because a number of studies have demonstrated the need for realistic simulation of soil moisture variability (e.g., Rowntree and Bolton 1983; Benjamin and Carlson 1986; Lanicci et

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al. 1987), Groves (1989), Mahfouf (1991), and Capehart and Carlson (1994) have proposed the technique of using the water and energy budget of a surface–vegetation–atmosphere transfer model (SVAT), driven by standard meteorological observations, to predict soil moisture. Similarly, four-dimensional data assimilation of surface meteorological data into land surface models also has been suggested as a method to retrieve soil moisture by simulation (McNider et al. 1994; Houser et al. 1998). Thus, it is worth exploring in detail how a currently available and typical SVAT scheme might perform “as is” (i.e., untuned) in a data-poor environment.

For this study, Wetzel and Boone’s (1995) Parameterization for Land–Atmosphere–Cloud Exchange (PLACE) model was chosen because it is a general purpose land surface parameterization designed to be coupled to atmospheric models and is broadly representative of the many schemes available in the community today (Henderson-Sellers et al. 1995). Specifically, we seek to understand how well PLACE, a typical SVAT, can simulate the spatial variability and temporal evolution of soil moisture and surface fluxes under conditions for which it has not been optimized. Accurate soil moisture and surface fluxes are essential for accurate boundary layer growth (McCumber and Pielke 1981; Pan and Mahrt 1987; Ek and Cuenca 1994; Segal et al. 1995), the generation of mesoscale circulations (Anthes 1984; Ookouchi et al. 1984; Mahfouf et al. 1987; Segal et al. 1988; Yan and Anthes 1988), and clouds and precipitation (Clark and Arritt 1995; Hong et al. 1995; Avissar and Liu 1996; Lynn et al. 1998). Although a simple simulation initialization scheme is used, advantage is taken of the wealth of remotely sensed soil moisture imagery and supporting data from the Southern Great Plains 1997 (SGP97) Hydrology Experiment (Jackson 1997) for validation of simulation results. The current case study takes place 9–16 July (192 h) and encompasses several wetting and drying cycles to observe the performance of PLACE under both conditions and its ability to make a transition from one to the other.

Several guidelines were set for conducting this research. First, since soil moisture and soil temperature are not commonly measured geophysical variables, the simulations use simple initial soil moisture and soil temperature profiles, as if PLACE were being run with limited soil moisture data. The landscape is defined using readily available soil texture and vegetation datasets and relevant literature. Second, simulations are forced with standard meteorological observations, such as temperature, pressure, and wind speed, and are not corrected or updated with observed data on soil moisture or heat fluxes. Third, tuning of model equations and parameters to optimize the simulation is avoided. By understanding the performance of PLACE under these conditions, predictions from coupled model simulations that involve similar SVAT models in limited-data environments may be put into proper perspective.

2. Data and methods

a. Numerical model

PLACE is a detailed process model of the heterogeneous land surface (Wetzel and Boone 1995). It consists of individual, linked modules that parameterize key components of surface and subsurface water and energy exchange. Water and energy are transmitted through a vertical column consisting of an interception/dew reservoir, a plant-storage reservoir, five soil moisture reservoirs, and seven soil heat reservoirs and are exchanged with the atmosphere through turbulent sensible and latent heat fluxes. Plant physiology and available liquid water determine the partitioning of the fluxes into latent and sensible heat. Although PLACE has simple parameterizations for runoff and base flow, it is not intended for applications that require detailed surface and subsurface cell-to-cell horizontal water transport. PLACE emphasizes the interaction of the heterogeneous land surface with the overlying turbulent atmosphere through the vertical column.

Details on the solution of PLACE’s energy and water balance equations can be found in Wetzel and Boone (1995). Following the parameterizations of Wetzel and Chang (1988), PLACE grid cells may be either vegetated or nonvegetated (bare soil or open water), and evapotranspiration may occur at either demand-limited (potential) or supply-limited (stressed) rates depending on soil water status. PLACE has four separate evapotranspiration equations to compute the demand- and supply-limited rates for bare soil and vegetation. It uses threshold values of plant water potential (usually -200 m) and soil water potential (the calculated wilting point of the soil) to determine which parameterization will be used. The latent and sensible heat fluxes then are calculated by bulk aerodynamic formulas.

b. Study area

SGP97 was the largest airborne L-band (1.413 GHz and 21 cm) passive microwave soil moisture mapping mission to date (Jackson et al. 1999). It took place 18 June–17 July 1997 in central Oklahoma for a 10 000 km² area. Figure 1 depicts the experiment area. One of the objectives of SGP97 was to evaluate the influence of soil moisture on the local surface energy budget and the influence of mesoscale variability in the surface energy budget on the development of the atmospheric boundary layer. To accomplish experiment objectives, the SGP97 field campaign combined satellite, aircraft, and truck-mounted remote sensing; ground-based soil moisture and vegetation mapping; and boundary layer profiling at a hierarchy of scales. The key contribution was 800-m-resolution maps of surface soil moisture from passive remote sensing (described in further detail in the next section) for 16 dates during this period. Additional detail on sampling activities can be found in Famiglietti et al. (1999) and Jackson et al. (1999).

FIG. 1. The SGP97 field campaign area. Ground sampling activities took place at the three locations marked by boxes on the maps. The "lines of flight" refer to the flight paths of the NASA P-3B aircraft. The study area is a 2500 km² box around the Little Washita watershed in southwestern Oklahoma. SHAWM soil moisture profile sites are indicated with an "x," Oklahoma Mesonet is shown by a black triangle, USDA-ARS Micronet is shown by an open circle, and fields where gravimetric samples were taken are indicated by a gray square. Labeled fields are used in later figures.

The study area is a 50 km $\times$ 50 km area surrounding the 610 km² Little Washita (LW) watershed, depicted in the inset in Fig. 1. The study period covers an 8-day period (192 h) from 9 to 16 July. The Little Washita River is part of the Washita River system in southwestern Oklahoma. Summers in this region are hot and relatively dry. The average daily high temperature in July is 35°C, and the average rainfall in July is 56.4 mm (Allen and Naney 1991). Topography in LW is gently rolling, changing about 10 meters per kilometer. In general, the soils are well-drained loams and sands, 1–2 m in depth, and overlie sandstone and shale bedrock (Allen and Naney 1991). Figure 2a is a 1-km grid of basic soil types derived from an average of the upper soil layers (components) of the U.S. Geological Survey's State Soil Geographic Database (STATSGO). The climate and soils have encouraged agriculture in the region, so that most of the land cover in the 1-km grid in Fig. 2b is grass rangeland and crops. As Fig. 1 shows, LW is heavily instrumented, containing four meteorological towers of the Oklahoma Mesonet and a 5–10-km-resolution micronet (42 towers) of the U.S. De-

partment of Agriculture Agricultural Research Service (USDA-ARS). During the case study, 22 fields in LW were gravimetric soil moisture sampling sites. We chose to study the LW precisely because of its dense network of meteorological and soil moisture ground sampling sites to force simulations and evaluate model performance.

c. Initialization, forcing, and validation data

The study area was divided into 1-km² grid cells, for a total of 2500 grid cells. This size was chosen for consistency with the SGP97 remotely sensed soil moisture maps. Each grid cell was assigned a soil texture and a land cover type. In Fig. 2a, five soil types occur within the study area, with over half of the study area consisting of loam and sandy loam. Each soil type was assigned fixed percentages of sand, silt, and clay from which PLACE calculated soil hydraulic properties using the empirical relations of Cosby et al. (1984). Table 1 summarizes the soil characteristics, and Table 2 contains the initial soil moisture and temperature profiles. To

TABLE 1. Specified soil characteristics.

	Sand	Sandy loam	Loam	Clay loam	Silt loam
Soil albedo	0.23	0.22	0.18	0.16	0.20
Fraction silt	0.00	0.30	0.45	0.55	0.65
Fraction sand	0.95	0.60	0.35	0.10	0.20
Fraction clay	0.05	0.10	0.20	0.35	0.15
Study area (%)	4	34	37	1	24

obtain the initial soil moisture profiles for each soil type, the gravimetric data on 9 July were used to determine a profile representative of fields with the same soil type. Initial soil temperature profiles also reflect field data from 9 July, but the same profile was assigned to all soil types.

The land cover grid in Fig. 2b was based on a 30-m-resolution grid compiled by Doraiswamy et al. (1999) for the SGP97 time period. Table 3 lists the important vegetation parameters used in the simulations. The 30-m-resolution land cover grid was smoothed to 1-km resolution by assigning the land cover type with the largest area within a 1-km cell to the entire 1-km cell. No distinctions were made among individual crop types, for example, corn, wheat, legumes, found in the original 30-m grid. Cells with low ($\leq 30\%$) cover were harvested by 9 July, and those with high ($\geq 70\%$) cover were not. Because no calibration curve relating the values in the normalized vegetation difference index (NDVI) imagery for SGP97 to specific values of percent cover by green vegetation was available, the assignments in Table 3 are only relative estimates of vegetation density. In Table 3, the albedos, stomatal resistances, and operating temperatures were obtained from Garratt (1992), Meeson et al. (1995), and Sellers et al. (1995). Root fractions were derived from the relations of Canadell et al. (1996) and Jackson et al. (1996). The percent green cover, leaf area index (LAI), and the constant surface roughness height ($\bar{z}_0$) were estimated from field observations, remote sensing, LW harvest activity, and a study of soil moisture in Oklahoma by DeLiberty (1994).

All simulations used a 5-min time step and were forced by shortwave radiation, longwave radiation, air temperature (at 2 m), station pressure, wind speed, specific humidity, and rainfall. To obtain values of these

FIG. 2. (a) Soil texture grid and (b) land cover grid for the study area. "Low" and "high" refer to cells with low ($\leq 30\%$) or high ($\geq 70\%$) cover by transpiring vegetation.

TABLE 2. Initial soil moisture and soil temperature profiles.

Layer	Depth* (cm)	Initial soil moisture (% of volume)					Initial soil temperature (K)	
		Sand	Sandy loam	Loam	Clay loam	Silt loam	Depth* (cm)	All (K)
1	1	0.03	0.04	0.07	0.10	0.10	2	298.2
2	5	0.03	0.04	0.10	0.10	0.10	5	298.2
3	25	0.10	0.15	0.20	0.20	0.20	10	298.2
4	100	0.15	0.20	0.25	0.25	0.25	25	300.0
5	200	0.20	0.20	0.30	0.30	0.30	100	300.0
6							150	300.0
7							200	300.0

* The bottom of the soil layer is the depth listed in columns 2 and 8, and the layer thickness is $\text{depth}_n - \text{depth}_{n-1}$.

TABLE 3. Specified vegetation characteristics. Here, ψ_c is the critical leaf water potential; RSMIn and RSMMax are the minimum and maximum stomatal resistances; root fractions 1–5 are the cumulative percentages of roots found in soil moisture reservoirs 1–5; z_0 is the constant aerodynamic surface roughness length; TempSTMin, TempSTMax, and TempSTStop are the stomatal minimum, maximum, and cessation operating temperatures, respectively. “High” and “low” refer to high or low percent transpiring cover.

	Broadleaf and needleleaf trees	Crops (high)	Crops (low)	Shrubs	Surface water	Grass (high)	Grass (low)
Albedo	0.18	0.15	0.15	0.15	0.10	0.20	0.25
% Transpiring cover	0.90	0.80	0.10	0.90	0.00	0.70	0.30
ψ_c (m)	–200	–210	–210	–200	0	–200	–200
RSMIn ($s\ m^{-1}$)	110	70	70	110	0	110	110
RSMMax ($s\ m^{-1}$)	1700	1700	1700	1700	0	1700	1700
Root fraction 1	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Root fraction 2	0.25	0.35	0.35	0.25	0.00	0.35	0.35
Root fraction 3	0.50	0.50	0.50	0.50	0.00	0.60	0.60
Root fraction 4	0.20	0.15	0.15	0.25	0.00	0.05	0.05
Root fraction 5	0.05	0.00	0.00	0.00	0.00	0.00	0.00
z_0 (m)	1.5	0.2	0.0012	0.5	0.0	0.2	0.2
TempSTMin (K)	270	278	278	283	0	283	283
TempSTMax (K)	315	325	325	323	0	328	328
TempSTStop (K)	294	305	305	315	0	313	313
LAI	4.6	5.0	0.7	4.6	0.0	2.5	1.0
Domain %	1.2	1.5	8.2	0.6	0.4	74.5	13.6

variables at each grid cell, the continuous fields of short-wave radiation, air temperature, and relative humidity were interpolated linearly (Cressman weighting) from the 46 mesonet and micronet sites. Pressure and wind speed also were interpolated but from just the four mesonet sites. Specific humidity and longwave radiation were calculated from relative humidity, pressure, and air temperature from the blackbody radiation equation and the empirical relationships of Bolton (1980) and Croley (1989). Because rainfall is not continuous spatially, the radar-derived accumulated rainfall (stage III) product available from the National Weather Service Arkansas–Red River Forecast Center for the study time period was used. Figure 3 is the accumulated rainfall from a squall line passage on 10–11 July. In addition to this major event, two briefer events took place on 9 and 15 July over the western and southwestern study area. Because the stage-III product has 4-km and 1-h resolution, smoothing and temporal interpolation were required. Rainfall amounts from the 4-km stage-III cells were assigned to all of the (1 km) model cells within them. The percentage of rain gauge rainfall that fell at each time step with respect to the hourly total was calculated, and this percentage was applied to the nearby stage-III cells for the same time step. If there were no rain gauges within 10 km, or rain gauge data were missing, the stage-III hourly amount was divided by 12 and applied to each time step within the hour.

During the SGP97 field campaign, near-daily overflights by the National Aeronautics and Space Administration (NASA) P-3B aircraft fitted with the Electronically Scanned Thinned Array Radiometer (ESTAR), an L-band passive microwave sensor, occurred. Figure 1 shows the lines of flight flown by the P-3B missions and the approximate area of coverage. The P-3B overflew LW at approximately 1600 UTC (1000 CST). The footprint of the raw brightness temperature data is 400

m, but the raw data were resampled to 800 m to derive soil moisture maps. Further details on the ESTAR instrument and the inversion of ESTAR brightness temperatures to volumetric soil moisture can be found in LeVine et al. (1994), Jackson et al. (1995), and Jackson and LeVine (1996). ESTAR-derived soil moisture estimates are within 3% of estimates of volumetric soil moisture from SGP97 ground samples (Jackson et al. 1999). The gray squares in Fig. 1 are the 22 ground sampling fields in LW during the study. The mean gravimetric soil moisture content of these fields was estimated every day of the study period, and the gravimetric soil moisture was converted to volumetric soil moisture using field-observed mean bulk densities (Jackson 1997; Famiglietti et al. 1999; Jackson et al. 1999).

Some of the fields (Fig. 1) in LW contained soil heat and water measurement (SHAWM) stations. SHAWM-equipped fields provided vertical soil water profiles for comparison with simulation results. SHAWM stations have a nest of soil heat dissipation sensors, Campbell Scientific Model 229-L, with three at 5 cm, and one each at 10, 15, 20, 25, and 60 cm. Reece (1996) found that the 229-L gave reliable estimates of soil water matric potential between -10 and $-1200\ J\ kg^{-1}$ (-0.1 to -12 bar), inclusive. The 229-L consists of a small, hollow, stainless steel tube inserted into a porous ceramic matrix. The stainless steel tube contains a miniature heater and thermocouple. An initial soil temperature is taken with the sensor; the sensor is heated for 21 s, and then a final soil temperature reading is made. The difference ΔT between the initial and ending soil temperature readings is inversely related to matric potential. The ΔT –matric potential relationship is developed from sensor calibration (Starks et al. 1999). Conversion of matric potential to volumetric soil moisture required field-specific soil water retention curves (not listed here)

FIG. 3. Accumulated rainfall (mm) for the major rain event that took place from approximately 2000 UTC on 10 Jul to 0500 UTC on 11 Jul. Two smaller rain events on 9 and 15 Jul are not depicted.

derived by the USDA-ARS Grazinglands Research Laboratory.

3. Results

In keeping with the study goal, the initial conditions described in the previous section (Tables 1, 2 and 3) were not adjusted to improve model results. In this section, surface and subsurface soil moisture, soil temperature, and latent and sensible heat fluxes will be compared with available field and remote sensing data.

a. Surface soil moisture

Figure 4 compares the ESTAR and model-derived soil moisture maps just after the 10–11 July storm and after several days of dry-down. The model-derived soil moisture map is the weighted average of PLACE's top two soil moisture layers. The "no-data" areas in the model-derived map were defined as surface water. In the maps in Fig. 4 and for other days (12, 13, and 16 July) not shown here, the modeled soils are drier than the ESTAR-observed soils in the west and southwest and wetter everywhere else, particularly in the northeast and south-

east. Except for the area surrounding the reservoir on the western edge of the study area, the errors in the difference grid for 11 July are negative and randomly distributed, with no significant correlation to rainfall (Pearson correlation coefficient $R = 0.29$). In contrast, the errors in the 14 July difference grid have a high correlation to rainfall ($R = 0.70$), with high positive errors in the wettest areas and high negative errors in the driest areas. In Fig. 5, the soil textures are plotted over the soil moisture maps for 14 July. These overlays reveal a noticeable control of soil texture over soil moisture in both the model-derived and ESTAR maps, although the soil texture control is stronger in the model-derived map. Jackson et al. (1995), Hollenbeck et al. (1996), and Mattikalli et al. (1998) have previously observed a similar control of soil texture over soil moisture in remotely sensed soil moisture images.

Figure 6 summarizes the model results versus ESTAR and field-observed 5-cm soil moisture. The modeled mean soil moisture is wetter than the ESTAR mean is, particularly at the end of the dry-down (13–14 July), and has a smaller, slower rate of decrease. Conversely, the modeled mean is drier than the field data from 11 through 13 July. The difference between the model and

FIG. 4. 5-cm soil moisture maps of ESTAR and model for 11 and 14 Jul plus maps of the difference 100 (ESTAR - model). The legend plotted in the 11 Jul model map applies to the other soil moisture maps. The legend plotted in the 11 Jul difference map applies to the 14 Jul difference map.

FIG. 5. Soil texture plotted over the ESTAR and modeled 5-cm soil moisture maps for 14 Jul.

FIG. 6. Time series of the mean modeled 5-cm soil moisture (a) vs the mean of ESTAR and (b) vs the mean of field data. (c) Time series of the rmse. (d) Rmse by soil type with comparison points listed above each bar.

FIG. 7. Time series of model-derived vs SHAWM soil moisture at 25 cm for (a) representative loam and (b) sandy loam stations. The fields used are labeled in Fig. 1. In (c), the rmse at 25 and 60 cm by soil type with comparison points listed above 60-cm bars is shown and, in (d), time series of rmse for sandy loam and loam sites are shown.

both the ESTAR and field means narrows after rewetting by rainfall on 15 July. In Fig. 6c the root-mean-square error (rmse) versus ESTAR increases with time after rainfall but decreases versus field data. With respect to soil type, the lowest rmse were for loam and sandy loam (70% of the study area), and the highest were for clay loam and silt loam. A small area of clay loam borders the reservoir on the western edge of the study area. The ESTAR-versus-model comparison in this particular area is questionable because some of the ESTAR soil moisture values are greater than 50% even on 14 July, several days after the heavy rain on 11 July (Jackson et al. 1999).

b. Subsurface soil moisture

To initialize a simulation only one soil moisture profile was assigned to all of the grid cells of a particular soil type. Figures 7a,b compare the behavior at four SHAWM sites¹ for the model versus the soil heat dis-

sipation sensor at 25 cm. The temporal evolution of the modeled loams compares better to the SHAWM-observed loams than the modeled sandy loams compare to the SHAWM-observed sandy loams. Infiltration into the modeled loams from the 10–11 July storm occurs approximately at the same time as observed, and the modeled loams dry down only slightly faster than do the observed loams from 11–13 July. In contrast, the modeled sandy loams become much wetter from the rain events on 10–11 and 15 July than do the observed sandy loams. Observed infiltration at LW18 from the 10–11 July storm takes place 24 h after model infiltration. After 11 July, the observed 25-cm soil moisture values change very little for both LW07 and LW18, whereas the model values decrease noticeably (~20%). Figure 7c confirms that, for all the SHAWM sites, the loam results are better than those for the sandy loam, and the 60-cm results are better than the 25-cm results. Rmse at individual fields ranges from 2% to 9% at 25 cm, and from 0.5% to 8% at 60 cm. Figure 7d reveals a correlation between time and error. The 25-cm error for both loam and sandy loam rises dramatically after the 10–11 July storm, and it rises for the sandy loams after the 15 July storm. If the 24 h between 11 and 12 July are dropped, the 25-

¹ Two loam (LW02 and LW15) and two sandy loam (LW07 and LW18) sites are plotted; locations are in Fig. 1.

FIG. 8. Time series for 5-, 10-, 15-, and 30-cm depths of the mean soil temperature of the 46 meteorological towers vs the average of the model grid cells corresponding to those locations.

cm rmse improves by 2%. By comparison, if the silt loam field is ignored, the 25-cm rmse for series “all” in Fig. 7c improves by 0.5%.

c. Soil temperature

Soil temperatures are recorded at 15-min intervals for the 5-, 10-, 15-, and 30-cm levels, respectively, at all of the mesonet and micronet towers (46 total). Figure 8 compares time series of the means of the observed soil temperatures and the means of the modeled soil temperatures corresponding to the same points. The initial modeled soil temperatures (Table 1b) are much cooler than those observed, and approximately 12 h are required for model adjustment. Throughout the study period, the modeled mean soil temperatures in the 5- and 10-cm layers have a stronger diurnal cycle than the observed means, peaking several degrees higher in the middle of the day and dropping several degrees lower in the middle of the night. This discrepancy increases as the dry-down progresses and is not changed by the rewetting on 15 July. Diurnal fluctuations dampen quickly with depth, such that the modeled mean has a

weaker diurnal signal than does the observed mean in the 30-cm layer.

The rmse for the study period is not related strongly to rainfall ($R < 0.20$) or soil type. In Fig. 9a for all soil types, the rmse decreases by over 50% in nearly equal increments from the 5- to 30-cm depths but does not vary more than 1 K among soil types at the same depth. Small differences (no greater than 2 K) among soil types at 5 cm appear in the time series in Fig. 9b. Error caused by the simple initialization scheme decreases more rapidly at the silt loam towers and remains lower than for the other soil types through the 10–11 July storm. Silt loam had the wettest initial soil moisture profile (Table 2), and four of the seven silt loam towers had the highest rainfall totals for all of the rain events during the study period (noted on Fig. 9b). The loam towers had the lowest rainfall totals. The error at the loam towers is slightly higher through the first 48 h of dry-down, although late in the dry-down, all of the soil types have errors of comparable magnitude. The error for the silt loam and sandy loam towers decreases at the end of the study period as the soil dries down from rainfall on 15 July at some of these towers.

FIG. 9. In (a) the rmse for modeled vs observed soil temperature for 5-, 10-, 15-, and 30-cm depths by soil type with comparison points listed above each series is shown. (b) Time series of rmse at 5 cm for silt loam, sandy loam, and loam sites. Rain events are marked lowercase for brief events covering only part of the study area and uppercase for the major event covering the entire study area.

d. Surface fluxes

Surface flux data in LW during the study period are limited to observations at fields LW02, a loam field, and LW07, a sandy loam field (see Fig. 1 for locations). Figure 10 compares the observed and modeled time series of latent and sensible heat fluxes at both of these fields. On 9, 10, and 15 July, periods of very light rain occurred and are apparent as the spikes in Figs. 10b and 10d. Light rain tends not to infiltrate PLACE soils but evaporates almost immediately, briefly forcing very high latent heat fluxes. For the first 48 h following the 10–11 July storm, the magnitude and temporal evolution of the modeled and observed latent heat fluxes are comparable. For dry conditions prior to the 10–11 July storm and after 13 July, the peak modeled latent heat flux is as much as 100 W m^{-2} lower at LW02 and 250 W m^{-2} lower at LW07 than the observed. The peak modeled latent heat flux declines by 25% at LW02 and 50% for LW07 from 12 to 15 July, although there is no decrease in the observed peak latent heat flux at LW07 and only a 10% decrease at LW02. Late in the dry-down, there are also fewer hours with latent heat fluxes greater than

200 W m^{-2} for the model fields. Prior to the 10–11 July storm, the magnitude of the peak modeled sensible heat fluxes is 200 W m^{-2} higher than observed. After 13 July, the difference is 50 W m^{-2} at LW07 and 100 W m^{-2} at LW02. More noticeable is the 1–2-h time lag between the modeled and observed peaks. After peaking at midday, modeled net radiation and latent heat flux decline, but ground heat flux and sensible heat flux continue to rise briefly to maintain energy balance, creating a lag between observed and modeled sensible heat flux.

Because maps of observed surface fluxes are not available, and soil moisture has a strong relationship to surface fluxes, the modeled surface fluxes will be compared with ESTAR soil moisture and accumulated rainfall (mm) to assess how well they reflect environmental conditions. The R between the ESTAR soil moisture and the corresponding latent heat flux at each grid cell for 2000 UTC and also R between the accumulated rainfall and latent heat flux were calculated. Between 11 and 14 July, the total from the storms on 9 and 10–11 July was used, and on 15 and 16 July the total from the storms on 15 July was added. Time series of the correlations to rainfall and ESTAR soil moisture are in Fig. 11. Because intercepted rain during light rainfall causes a spike in PLACE's calculation of evapotranspiration rates, the dip in the time series on 15 July reflects that rain was occurring in some places at 2000 UTC, and small rainfall amounts did not correlate well with high latent heat fluxes. Otherwise, the correlations ($>50\%$) in Fig. 11 imply that the model latent heat fluxes did reflect the rainfall gradient for much of the study area for most of the study period. The correlation with rainfall rises to 72% after the 10–11 July storm and remains above 70% through the dry-down. The correlation with ESTAR soil moisture is 5%–10% lower than the correlation with rainfall during the same time period. Because ESTAR soil moisture reflects the rainfall that actually infiltrated the soils in the study area, the difference between the time series in Fig. 11 was attributed to differences between the stage-III rainfall forcing the simulation and the observed rainfall. In the next section, the uncertainty in the rainfall forcing data and its effect on the results is quantified.

4. Discussion

In this section, how much the simple initialization scheme, forcing data, and model structure affect the results is considered. First the period when the simulation was most accurate was identified. With respect to surface soil moisture, 5- and 10-cm soil temperature, and surface fluxes, the most accurate simulation occurred during the 48 h following heavy rainfall. A standard of 5% error (vs ESTAR) for soil moisture and 2 K for soil temperature,² was met by this simulation for

² Because of the limited observed surface flux data, a similar assessment was not made for model surface fluxes.

FIG. 10. Time series of modeled and observed surface fluxes for two different sites, LW02 and LW07 (locations in Fig. 1). To emphasize the daytime fluxes, the scales are truncated at the bottom.

FIG. 11. Time series of the correlation between modeled latent heat flux and rainfall and ESTAR soil moisture. Correlations were calculated at 2000 UTC. On 15 Jul light rainfall was occurring in some parts of the western and southwestern study area, causing the dip in the "vs-rainfall" time series.

surface soil moisture on 11–13 and 16 July and for 5- and 10-cm soil temperature on 10–13 and 16 July. At LW02 and LW07, the model surface fluxes most-closely tracked the magnitude and temporal evolution of observed surface fluxes on 11–13 July. For subsurface soil moisture, each depth and soil type had its own best time period. Only the 60-cm loams were consistently at or below 5%. The 25-cm loams dropped and stayed below 5% late on 12 July after 30 h of drying. Except for a brief period after 11 July at 60 cm, the sandy loams were consistently above 5%.

The initialization scheme has a noticeable signal in the time series of error in the simulation of subsurface soil moisture and soil temperature (Figs. 7, Figs. 8, and Figs. 9b). The model 5–15-cm soil temperatures are lower than observed until PLACE is able to adjust to daytime heating. In Fig. 7d prior to the 10–11 July storm, the rmse for subsurface soil moisture by soil type and depth ranges from 3% to 9%. At the beginning of the study period, there was more variability in subsurface soil moisture than in surface soil moisture. On 9 July, the 5-cm mean soil moisture for the gravimetric observation sites ranged from 2% to 12% of volume,

TABLE 4. Summary of results of sensitivity tests at LW07. The 5-cm soil moisture on 16 Jul was 10% for sand and 30% for clay, and intermediate soil types (sandy loam, loam, etc.) were between 10% and 30%. The other tests are interpreted similarly.

Test	5-cm soil moisture on 16 Jul	Maximum latent heat flux on 14 Jul
Vary soil type sand to clay	10%–30%	400–100 W m^{-2}
Vary cover 10%–90%	10%–15%	100–400 W m^{-2}
Vary initial soil moisture layers 1–2 (5%–30%)	No difference	No difference
Vary initial soil moisture layers 3–5 (10%–40%)	10%–15%	300–400 W m^{-2}

as compared with 5%–18% at 25 cm and 11%–25% at 60 cm for the SHAWM sites. Because spatial variability in surface and subsurface soil moisture on 9 July did not have a strong relationship to soil type, it was not reflected in the initialization scheme. Only after the 10–11 July storm was the effect of the initialization scheme eliminated for the 5–15-cm soil temperature, 5-cm soil moisture, and surface fluxes. The modeled 30-cm soil temperatures (Fig. 8d) and soil moisture below 60 cm (not shown), however, hardly changed during the study period. Hence, the initialization scheme largely determined the condition of the lowest soil layers, even during a time period with considerable daytime heating and rainfall.

To understand the sensitivity to the initialization scheme for the study time period, a series of sensitivity tests were conducted on initial soil moisture, soil type, and cover by transpiring vegetation. To save computation time a single location, LW07, representative of much of the study area (sandy loam soil and grass cover), was used. The initial soil moisture profile for sandy loam (Table 2) was used to fix the grass cover at 70% and the soil types³ were allowed to vary, and then the soil type was fixed at sandy loam and the amount of transpiring vegetation was varied from 10% to 90% of total area. With sandy loam soil and 70% grass cover, the initial soil moisture was varied from 5% to 30% for the top two layers with the bottom three layers fixed, and then the initial soil moisture was varied from 10%–40% for the bottom three layers with the top two layers fixed. Table 4 summarizes the results. As in Fig. 5, the specification of soil type had the greatest impact on surface soil moisture and heat fluxes. The specification of transpiring cover and soil moisture in the lower reservoirs but not in the upper reservoirs had a significant impact, $O(100 \text{ W m}^{-2})$, on surface fluxes, particularly late in the dry-down. For a summertime simulation, even one with heavy rainfall such as this case study, the accuracy of the simulation of surface soil moisture, soil temperature, and heat fluxes late in a dry-down would

FIG. 12. (a) Histograms of rainfall totals after the 10–11 Jul storm for the rain gauges vs the stage-III cells corresponding to the same sites. (b) Results of sensitivity tests of rainfall on maximum latent heat flux on 14 Jul at LW07, in which rainfall was added or subtracted in 10% increments.

depend heavily on the availability of water and the energy balance in these lower soil reservoirs and thus the accuracy of the initialization of these reservoirs.

The forcing data for the simulation of the watershed were the stage-III rainfall product (temporally interpolated from mesonet and micronet rain gauge data). In Fig. 12a are histograms of rainfall totals after the 10–11 July storm for the rain gauges versus the stage-III cells corresponding to the same sites. In Fig. 12a there are more rain gauges with both low (<30 mm) and high (>120 mm) rainfall totals, although there were more stage-III cells with totals between 80 and 120 mm. Consistent with the 14 July 5-cm soil moisture difference grid in Fig. 4, the stage-III rainfall totals tended to be higher than the driest sites in LW and significantly lower than the wettest sites. The rmse of the stage-III totals versus the rain gauges is 20 mm, a 10% error for the wettest rain gauges in LW and a 35% error for the average rain gauge total. Note also that there were seven rain gauges below 20 mm. Sensitivity tests of forcing data at LW07 were conducted because it contains a me-

³ Clay, silty clay, clay loam, silt loam, loam, and sand were tested in order.

sonet tower (Apache, Oklahoma), so that the sensitivity to atmospheric variables could be isolated directly without having to consider an interpolation error. In Fig. 11 the correlation of modeled latent heat fluxes with the rainfall forcing data is higher than is the correlation with ESTAR soil moisture. Figure 12b depicts the results of increasing and decreasing rainfall at LW07 on maximum latent heat flux on 14 July. The latent heat flux increases (decreases) almost linearly with increasing (decreasing) rainfall. The value for the stage-III cell covering LW07 is 19% higher than the rain gauge data, and thus the maximum latent heat flux on 14 July is almost 100 W m^{-2} higher in Fig. 10d than in Fig. 12b for the zero-change simulation. For an error in the rainfall of 35%, the change in latent heat flux from Fig. 12b would be about 150 W m^{-2} .

The maximum observed latent flux at LW07 on 14 July is 430 W m^{-2} , greater even than the simulation driven by +50% rainfall shown in Fig. 12b. The soil thermal conductivity parameterization in PLACE is based on the work of McCumber and Pielke (1981). In a study comparing the effect of common soil thermal conductivity schemes on model heat fluxes and temperatures, Peters-Lidard et al. (1998) conclude that the McCumber and Pielke (1981) scheme overestimates the thermal conductivity during wet periods and underestimates it during dry periods. The version of PLACE used in this study did not account for soil water vapor flux, reducing the potential evapotranspiration. Hence, late in a dry-down, one can expect higher-than-observed daytime skin temperatures, surface soil temperatures, and higher sensible heat flux, with magnitudes strongly controlled by soil type (Peters-Lidard et al. 1998). As seen in Figs. 4, 5, and 6a, the modeled soils at 5 cm tended to be wetter than the ESTAR soils except in the wettest portion of the study area, had less spatial variability than those of ESTAR, were strongly controlled by soil type, and had errors correlated to rainfall. Because the model evapotranspiration was not as high as it could have been based on atmospheric demand (high) late in the dry-down, because the moisture content of subsurface layers was initialized by soil type, and because stage-III rainfall totals tended to be higher except in the wettest portions of the study area, the drying of soil layers, both surface and subsurface, was suppressed, thereby reinforcing the control of soil type and reducing spatial variability.

5. Conclusions

In this study, an SVAT driven by standard meteorological observations was used to conduct a case study of a 2500 km^2 area around the Little Washita watershed in southwestern Oklahoma for 9–16 July 1997, during the Southern Great Plains 1997 Hydrology Experiment. We chose the Parameterization for Land–Atmosphere–Cloud Exchange, a land surface parameterization designed for use in mesoscale and global atmospheric

models. The objective was to evaluate how well this typical SVAT simulated the spatial variability and temporal evolution of soil moisture and heat fluxes without optimization for the case study. Tuning of model equations or parameters was avoided, and simple initial soil moisture and temperature profiles assigned by soil type were used. The experimental design assumed a data-poor environment, typical of much of the world, in which only general soil and vegetation parameter datasets are available and limited-to-no auxiliary data (such as soil moisture) are available. The rich SGP97 dataset was taken advantage of, however, to evaluate the results to assess the quality of the simulation.

The initialization scheme did not account for the spatial variability of subsurface soil moisture (recall Fig. 7d), nor, because of the lack of a calibration curve for NDVI imagery, was spatial variability in vegetation cover able to be accounted for. Moreover, 30-m-resolution soil texture data covered only LW but not the rest of the study area, so a soil texture grid derived from an average of the upper soil layers (components) of the STATSGO database was relied upon. All of these factors, as well as significant uncertainty in the rainfall data ($>10\%$), could account for differences between modeled and observed surface fluxes that were on the order of 100 W m^{-2} and differences in soil moisture of greater than 5%. It also is likely that the soil thermal conductivity scheme in PLACE (McCumber and Pielke 1981) damped PLACE's response to atmospheric demand after 13 July (Peters-Lidard et al. 1998), resulting in reduced evapotranspiration and warmer but slower-drying soils. Even with the simplifications of the initialization, the simulation results were satisfactory, meeting the performance criteria, for at least the first 48 h following the rainfall on 10–11 July and again after the rainfall on 15 July. The rainfall on 10–11 July essentially reset the simulation by eliminating the effect of the initialization scheme on at least the first 25 cm of soil. PLACE simulated the general pattern of the soil moisture and soil temperature in the study area reasonably well for moist conditions, drifting from observation under dry conditions when spatial variability in soil moisture was highest. It is expected that SVATs such as PLACE (Henderson-Sellers et al. 1995) also would demonstrate a strong soil texture control on soil moisture and surface fluxes and a similarly limited spatial variability.

In an environment where only $1^\circ \times 1^\circ$ soil texture and vegetation data are available, initial soil moisture and soil temperatures may be linked de facto to soil type as in this initialization scheme. Critically important subsurface soil moisture and soil temperature data are even less likely to be available than surface data are. The implication of this case study is that, in a data-poor environment, uncertainties associated with a simple initialization and forcing data based on model reanalysis or remote sensing may be greater than the magnitude of the forcing variables used. These uncertainties mi-

grate to simulated variables such as latent heat flux and are exacerbated by dry conditions. Since a slow-growing, moist boundary layer favors deep convection, the modeled soil moisture and surface fluxes influence the opportunity for deep convective clouds and precipitation in a coupled (land–atmosphere) simulation, creating a feedback not present in the offline simulations. Further investigation is warranted to quantify the scope of this influence.

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