Sensitivity of the PBL and Precipitation in 12-Day Simulations of Warm-Season Convection Using Different Land Surface Models and Soil Wetness Conditions

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(Manuscript received 22 June 2007, in final form 15 November 2007)

ABSTRACT

A coupled land surface–atmospheric model that permits grid-resolved deep convection is used to examine linkages between land surface conditions, the planetary boundary layer (PBL), and precipitation during a 12-day warm-season period over the central United States. The period of study (9–21 June 2002) coincided with an extensive dry soil moisture anomaly over the western United States and adjacent high plains and wetter-than-normal soil conditions over parts of the Midwest. A range of possible atmospheric responses to soil wetness is diagnosed from a set of simulations that use land surface models (LSMs) of varying sophistication and initial land surface conditions of varying resolution and specificity to the period of study.

Results suggest that the choice of LSM [Noah or the less sophisticated simple slab soil model (SLAB)] significantly influences the diurnal cycle of near-surface potential temperature and water vapor mixing ratio. The initial soil wetness also has a major impact on these thermodynamic variables, particularly during and immediately following the most intense phase of daytime surface heating. The soil wetness influences the daytime PBL evolution through both local and upstream surface evaporation and sensible heat fluxes, and through differences in the mesoscale vertical circulation that develops in response to horizontal gradients of the latter. Resulting differences in late afternoon PBL moist static energy and stability near the PBL top are associated with differences in subsequent late afternoon and evening precipitation in locations where the initial soil wetness differs among simulations. In contrast to the initial soil wetness, soil moisture evolution has negligible effects on the mean regional-scale thermodynamic conditions and precipitation during the 12-day period.

1. Introduction

Sensitivity of continental precipitation to land surface conditions, including soil moisture, has been established for a wide range of time scales. These time scales range from less than a single diurnal cycle through influences on the timing and location of deep convection initiation (e.g., Lanicci et al. 1987; Shaw et al. 1997; Ziegler et al. 1997; Trier et al. 2004; Holt et al. 2006; Sutton et al. 2006) to seasonal via impacts of persistent soil moisture anomalies on seasonal precipitation amounts (e.g., Koster et al. 2004a,b; Ruiz-Barradas and Nigam 2005).

In this paper, we use a three-dimensional atmospheric model coupled with different land surface models (LSMs) to examine relationships between the land surface, the planetary boundary layer (PBL), and precipitation. The PBL evolution is a potentially important linkage between soil moisture and precipitation because soil wetness has been observed to strongly impact the daytime moist static energy content of the PBL (Betts and Ball 1995).

We examine the sensitivity of mean PBL structure and precipitation over the southern Great Plains (SGP) of the United States for a 12-day warm-season period in 2002 during which a persistent regional soil moisture anomaly occurred. Koster et al. (2004a) note that sensitive areas for seasonal precipitation are typically transition regions between wet and dry climates, such as the SGP, where soil wetness or moisture availability (degree of saturation in the soil) is between 0.1 and 0.3. They hypothesize that below 0.1 evaporation rates are too small to significantly influence precipitation, whereas above 0.3 the soil is sufficiently moist that
evaporation, and thus precipitation, is relatively insensitive to soil moisture changes.

Although sensitivity has been demonstrated, effects of soil moisture on precipitation vary significantly in different studies and have thus been difficult to generalize. Findell and Eltahir (2003a,b) found, using a one-dimensional model, that atmospheric controls dominate convection initiation when relative humidity is either very high or low and when static stability in the 2-km-deep layer above the early morning PBL is large. However, for moderate relative humidity and more unstable lapse rates, they found strong sensitivity of convective triggering to soil moisture with drier soils preferred in drier and more unstable atmospheric conditions and wetter soils preferred under moister and more stable atmospheric regimes.

Other studies have emphasized the hydrological importance of horizontal transport of water vapor from remote regions (e.g., Trenberth 1999; Dirmeyer and Brubaker 1999). The 1993 summer regional-scale precipitation anomaly that led to widespread flooding over the upper Mississippi River basin of the central United States presents an example that has been widely investigated using both global (e.g., Trenberth and Guillemot 1996; Beljaars et al. 1996; Viterbo and Betts 1999) and regional (Paegle et al. 1996; Pan et al. 1996; Giorgi et al. 1996; Bosilovich and Sun 1999; Xue et al. 2001; Anderson et al. 2003) models. Contrasting results on the role of soil wetness were found in studies of this period.

Beljaars et al. (1996) concluded that less accurate precipitation forecasts were related to capping inversions above the PBL, which arose from strong sensible heating over anomalously dry soil located ~1 day upstream. Viterbo and Betts (1999) found that more realistic soil moisture anomalies specific to the 1993 warm season could not account for the precipitation anomaly, but, similar to Beljaars et al. (1996), they concluded that heavier precipitation was promoted by wetter soil upstream. In contrast, Paegle et al. (1996) found a negative feedback between precipitation and upstream soil wetness for July 1993. They concluded that drier soil upstream resulted in stronger PBL forcing of the nocturnal low-level jet (LLJ), which led to enhanced convergence and water vapor transport into the region of the precipitation anomaly. They added that the role of local evaporation was to enhance the precipitation anomaly downstream of the LLJ core.

Fritsch et al. (1986) noted that much of the warm-season precipitation over the central United States arises from mesoscale convective systems (MCSs), which can occur in sequences lasting up to ~1 week. A complicating factor with such organized convection is that it can travel distances of ~1000 km (e.g., Carbone et al. 2002) and be influenced by atmospheric conditions along its entire path. The atmospheric conditions may in turn be influenced by both local and remote soil conditions along the lower-tropospheric inflow to convection.

Although one-dimensional model studies and high-resolution three-dimensional simulations over small domains have advanced our understanding of local land surface effects on convection initiation, they are unable to address how land surface effects can influence precipitation over multiple life cycles of mesoscale convection. On the other hand, regional climate model simulations employed over subseasonal or greater time scales are exposed to uncertainty in the linkages among land surface effects, the PBL, and precipitation that arise from the necessary use of cumulus parameterizations, in which closure assumptions that affect convection initiation are quite varied (e.g., Pan et al. 1996; Giorgi et al. 1996).

The current study applies a different approach to analyzing sensitivity of warm-season precipitation to land surface conditions. Our simulations over a ~2-week period within a subcontinental-scale domain help us extend results from previous approaches by allowing the land surface to influence the life cycle of explicitly triggered (i.e., grid resolved) mesoscale convection over multiple convection cycles.

In the next section we describe the atmospheric model and our design of numerical simulations. In section 3 we overview the large-scale meteorological conditions and precipitation during this period by comparing a control simulation with observations. Section 4 examines the diurnal cycle of thermodynamic properties of the PBL in simulations using both different LSMs and different soil wetness conditions. Relationships between the resulting differences in PBL conditions and precipitation differences among simulations are explored in section 5.

2. Numerical model and experimental design

The coupled Weather Research and Forecasting (WRF)–land surface modeling system is the investigative tool used in this study. The general principles underlying the coupling of the mesoscale atmospheric model with the LSM are discussed in Chen and Dudhia (2001). The 12-day simulations are from 1200 UTC 9 June to 1200 UTC 21 June 2002, which coincides with a portion of the International H2O Project (IHOP_2002; Weckwerth et al. 2004). Both episodes of long-lived organized deep convection and relatively quiescent interludes occurred over the SGP during this period (e.g.,
Wilson and Roberts (2006). Our emphasis is on the overall simulated precipitation properties and PBL conditions that influence them rather than on individual events.

### a. Atmospheric model

We integrate the Advanced Research WRF (ARW version 2) described by Skamarock et al. (2005) over the single subcontinental-scale domain shown in Fig. 1. The model contains $532 \times 385$ horizontal grid points at 4-km spacing. The vertical grid contains 31 levels, and is stretched to allow spacing of $\approx 100$ m near the lowest model grid point (at 25–30 m AGL) with $1$-km spacing at the model top near 50 hPa. The PBL scheme (Janjic 1990, 1994) predicts turbulent kinetic energy and allows vertical mixing between individual layers within the PBL. Other physical parameterizations include a bulk microphysics scheme based on Lin et al. (1983) and the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia (1989) shortwave radiation schemes.

Our choice of domain size and grid spacing reflects a compromise that allows us to assess mesoscale effects of surface conditions on the diurnal cycles of the PBL and precipitation over a large, subcontinental-scale region. The 4-km horizontal grid spacing, although not ideal for simulating isolated convective cells, is sufficient to explicitly resolve most salient mesoscale aspects of organized deep convection (e.g., Weisman et al. 1997; Bryan et al. 2003), which obviates the need for a cumulus parameterization. The initial condition and 3-h lateral boundary conditions are obtained from National Centers for Environmental Prediction (NCEP) Environmental Data Assimilation System (EDAS) analyses.

### b. Simulations

The importance of both land surface–atmosphere feedback processes and the initial land surface conditions on the daytime PBL evolution and precipitation during the 12-day period are determined from an analysis of four different simulations (Table 1). The simulations employ LSMs of varying sophistication and initial land surface conditions of differing detail and specificity during 9–21 June 2002.

The control simulation utilizes the Noah LSM (Ek et al. 2003) and employs the NCAR high-resolution land surface data assimilation system (HRLDAS) to initialize the land surface; this simulation is termed HRLDAS1. The LSM has a single vegetation canopy layer and predicts volumetric soil moisture and temperature in four soil layers. The depths of the individual soil layers are sequentially 0.1, 0.3, 0.6, and 1.0 m. The root zone is contained in the upper 1 m (top 3 layers).

HRLDAS (Chen et al. 2007) is run offline but on the same 4-km grid as the WRF simulation for an 18-month spinup period ending at the initialization time (1200 UTC 9 June 2002). This land surface initialization uses a variety of observed and analyzed conditions including 1) National Weather Service (NWS) Office of Hydrology stage 4 rainfall data on a 4-km national grid (Fulton et al. 1998); 2) 0.5° hourly downward solar radiation derived from Geostationary Operational Environmental Satellites (GOES) 8 and 9 (GOES-8 and GOES-9) as described by Pinker et al. (2002); 3) near-surface atmospheric temperature, humidity, wind, downward longwave radiation, and surface pressure from 3-hourly EDAS analyses; 4) 1-km horizontal resolution U.S. Geological Survey 24-category land-use and 1-km hori-

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<th>Land surface physics</th>
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<tr>
<td>HRLDAS1</td>
<td>Date-specific HRLDAS soil moisture</td>
<td>WRF–Noah LSM</td>
<td>Soil moisture evolves</td>
</tr>
<tr>
<td>HRLDAS2</td>
<td>Date-specific HRLDAS soil moisture</td>
<td>WRF–Noah LSM</td>
<td>Soil moisture constant</td>
</tr>
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<td>SLAB bare ground</td>
<td>Moisture availability constant</td>
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<tr>
<td>SLAB2</td>
<td>“Calibrated” moisture availability</td>
<td>SLAB bare ground</td>
<td>Moisture availability constant</td>
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zontal resolution State Soil Geographic soil texture maps; and 5) 0.15° monthly satellite-derived green vegetation fraction based on 5-yr averages (Gutman and Ignatov 1997). The resulting soil moisture field used to initialize experiment HRLDAS1 (Fig. 1) contains both localized finescale structure and a regional-scale gradient across the SGP.

HRLDAS1 is compared to a simulation that uses the less sophisticated simple slab soil model (SLAB) LSM to calculate the surface fluxes; this simulation is termed SLAB1. Comparisons with simulations that use a relatively simple LSM such as SLAB can help determine the dominant physical processes in the more complete and complicated Noah LSM. A major difference between SLAB and the Noah LSM is that SLAB does not permit evolution of soil moisture. Unlike Noah, the SLAB LSM has no explicit vegetation canopy resistance. Another difference from simulation HRLDAS1 is the lower resolution of the soil wetness field in simulation SLAB1 (Fig. 2a). In contrast to HRLDAS1, the SLAB1 surface moisture availability, $W$, depends only on the land-use type and has values broadly typical of summer. The land-use types and their distribution (Fig. 3) in SLAB1 are the same as in HRLDAS1. The differences in the specificity of the moisture availability variable are significant because of the large soil moisture anomaly for June 2002. An independent analysis (Fig. 4a) indicates much drier-than-normal soil over the western United States and SGP with a weak positive soil moisture anomaly over portions of the Midwest.

With this experimental design there are numerous possible causes for differences in the atmospheric response over the 12-day period, including 1) differences in the initial land surface condition, 2) differences in the capability for feedbacks from atmospheric inputs (including radiation and precipitation) to influence land surface conditions during the simulation, and 3) differences in the representation of physical processes within the LSMS that calculate the surface fluxes. It is not possible to unambiguously separate influences of the soil moisture anomaly from different LSM physics on the PBL and precipitation differences between the primary simulations HRLDAS1 and SLAB1, which is a drawback of the current experimental design. Fortunately, direct comparisons may be made between simulations that use the same LSMS to help assess the sensitivity to some of these differences.

The influence of not allowing feedbacks to the land surface from atmospheric inputs is addressed with experiment HRLDAS2, which is identical to HRLDAS1 except that soil moisture is held constant for the 12-day simulation. The influence of differences in the initial soil wetness is investigated using experiment SLAB2, which uses the same LSM as SLAB1 but is calibrated to produce initial surface fluxes that are similar to those of HRLDAS1. In principle, effects of the soil moisture anomaly could be alternatively explored by comparing HRLDAS1 to a simulation that uses the same Noah LSM but has climatological soil moisture. However, at the beginning of this work we were unaware of preex-
isting soil moisture climatologies that had been tested within high-resolution atmospheric models such as WRF.

In designing the initial land-surface surface condition for SLAB2, we first define a moisture availability field that corresponds to the detailed and date specific HRLDAS1 soil moisture using

\[ W = \frac{(\Theta - \Theta_w)}{(\Theta_{\text{ref}} - \Theta_w)}. \]

In (1), \( \Theta \) is the top-layer volumetric soil moisture from HRLDAS1, \( \Theta_w \) is the volumetric soil moisture at the wilting point (no evaporation in SLAB), and \( \Theta_{\text{ref}} \) is the volumetric soil moisture at field capacity (free evaporation with no soil moisture stress in SLAB). The parameters \( \Theta_w \) and \( \Theta_{\text{ref}} \) depend on soil texture and are discussed further in Chen and Dudhia (2001).

The spatial pattern of \( W \) (Fig. 2b) derived from (1) strongly resembles that of the HRLDAS1 initial soil moisture (cf. Fig. 1) but yields significant regional biases in the surface fluxes as early as the first diurnal cycle when used unmodified to initialize SLAB2. Since the effect of the land surface on the atmosphere is communicated through the surface fluxes, it is critical to reduce the surface flux bias found at early stages. The surface flux biases and the empirical approach used to reduce them are described in the appendix.

Domainwide 12-day mean (day 1 omitted)1 biases and RMS differences from HRLDAS1 near-surface thermodynamic variables for the “calibrated” SLAB2 (hereafter simply SLAB2) and SLAB1 simulations are shown in Fig. 5. The overall similar tracks of the SLAB2 curves to their SLAB1 counterparts are indicative of significant differences in the handling of the diurnal cycle by the Noah and SLAB LSMs. However, biases (Fig. 5a) and RMS differences (Fig. 5b) from HRLDAS1 are substantially less in SLAB2. The greatest reductions, both overall and relative to SLAB1, occur from late morning to early evening (Fig. 5), when magnitudes of the SLAB2 biases and RMS differences from their HRLDAS1 counterparts are smaller than typical measurement errors (e.g., Mueller et al. 1993). The good agreement between the SLAB2 and HRLDAS1 afternoon thermodynamic variables is important since this period coincides with PBL-based convection initiation that lasts through midevening. The close afternoon agreement, despite using different

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1 Day 1 is omitted here, and in subsequent averages, to avoid possible effects of model spin up during the first diurnal cycle.
LSMs, results from calibration of the SLAB2 moisture availability to produce initial midday surface heat and moisture fluxes similar to those of HRLDAS1, and the relative importance of these fluxes to the surface energy balance at this time of day.

In SLAB2 the midday flux biases are significantly reduced (see appendix), while retaining the horizontal soil wetness gradient across the SGP (Fig. 2c) absent from SLAB1 (Fig. 2a). This is important because, as we shall see in section 4, the PBL is influenced not only by local magnitudes of fluxes but also by their spatial patterns. Notice that the SLAB2 – SLAB1 moisture availability difference field (Fig. 4b) has a pattern very similar to June 2002 soil moisture anomaly (Fig. 4a). In the forthcoming analysis we devote special attention to the two subdomains indicated in Fig. 4b where these moisture availability differences are particularly large.

3. Overview of meteorological conditions over the central United States during 9–21 June

In this section we examine basic characteristics of the atmospheric regime and precipitation using the control simulation (HRLDAS1) and observationally based analyses. The high plains of eastern Colorado and New Mexico mark the western extent of widespread, significant 12-day precipitation in both HRLDAS1 (Fig. 6a) and the stage 4 observations (Fig. 6b), which is heaviest and most widespread east of 100°W. Table 2 quantifies similar simulated and observed coverage of total precipitation exceeding 10 mm for both the west and east subdomains of interest (Fig. 6). However, for large total amounts exceeding 100 mm, the coverage of simulated precipitation in the east subdomain (19.2%) is approximately twice that observed (9.3%).

Both HRLDAS1 (Fig. 7a) and the EDAS analyses (Fig. 7b) have mean southerly surface flow over the SGP with plentiful moisture \( (q_v = 11–14 \text{ g kg}^{-1}) \) to support convection. The broad spatial distribution of total precipitation during the 12-day period (Figs. 6a,b) is consistent with the lack of a well-defined quasi-
stationary surface front or convergence zone (Fig. 7). Thus, the period differs from the prolonged periods of heavy warm-season precipitation confined to relatively focused regions that have been both simulated (Trier et al. 2006) and analyzed in climatological studies (e.g., Tuttle and Davis 2006). The July 1993 upper Mississippi River basin flood discussed in the introduction occurred when heavy rainfall from MCSs was confined to within or near a quasi-stationary frontal zone (Kunkel et al. 1994; Anderson et al. 2003).

### Table 2

Percentages of W and E subdomains depicted in Fig. 4b for which total precipitation during the 288-h simulation periods exceeds 10- and 100-mm thresholds for the HRLDAS1 simulation and the stage 4 observations.

<table>
<thead>
<tr>
<th>Precipitation threshold</th>
<th>HRLDAS1</th>
<th>Stage 4</th>
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<tr>
<td></td>
<td>W</td>
<td>E</td>
</tr>
<tr>
<td>≥10 mm</td>
<td>63.7%</td>
<td>97.3%</td>
</tr>
<tr>
<td>≥100 mm</td>
<td>1.4%</td>
<td>19.2%</td>
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Fig. 6. Total accumulated precipitation (mm) for 288 h from 1200 UTC 9 Jun to 1200 UTC 21 Jun 2002 from (a) simulation HRLDAS1, and (b) stage 4 radar observations (section 2) interpolated to the model’s 4-km grid. Insets are the W and E subdomains as in Fig. 4b.

Fig. 7. Mean 2100 UTC [1500 central standard time (CST)] horizontal winds and water vapor mixing ratio (g kg$^{-1}$) at the lowest model grid point ($-25$ m AGL) time averaged from 10–20 Jun 2002 for (a) simulation HRLDAS1 and (b) EDAS model analyses interpolated to the model’s 4-km horizontal grid. The plus symbol in (a) indicates the location of the mean simulated eastern vertical profiles discussed in section 4.

### 4. Sensitivity of PBL properties

#### a. Sensitivity of simulated near-surface conditions and comparisons with observations

The mean diurnal cycle of near-surface thermodynamic variables and the convective available potential energy (CAPE) for the most unstable 500-m-deep air parcel are compared to corresponding values from the EDAS analyses in Fig. 8. Over the west subdomain, the largest differences in potential temperature, $\theta$ (Fig. 8a), and water vapor mixing ratio, $q_v$ (Fig. 8c), are between the HRLDAS simulations and SLAB1, which use both different initial soil wetness and different LSMs. The HRLDAS simulations compare much better with the EDAS analysis than does SLAB1, which has a substan-
tial cold and moist bias. Unlike the west subdomain, the east subdomain lacks large $\theta$ differences among simulations at any time during the diurnal cycle and each simulation closely follows EDAS (Fig. 8b). However, the diurnal cycle of $q_v$ varies significantly among the simulations (Fig. 8d). Here, all simulations have 0.5–1.5 g kg$^{-1}$ smaller $q_v$ than EDAS. As in the west, SLAB1 has the largest daytime and early evening departures from EDAS (Fig. 8d).

Having constant soil moisture (HRLDAS2) exerts little effect on the mean diurnal cycle of thermodynamic variables in either subdomain (cf. HRLDAS1 and HRLDAS2 curves in Figs. 8a–d). This is consistent with the small daily differences and lack of a persistent trend in the subdomain averages of HRLDAS1 soil moisture (Figs. 9a,b). In contrast, the SLAB2 simulation reveals substantial deviations in the mean diurnal cycle of thermodynamic variables from SLAB1 (Figs. 8a–d) indicating the importance of the initial land surface condition. Recall that SLAB2 uses the same LSM as SLAB1 but was calibrated to produce midday surface sensible (SH) and latent (LH) heat fluxes similar to those of HRLDAS1 during the first diurnal cycle. These fluxes remain similar to HRLDAS1 during the next 11 days (Figs. 9c–f), and during the afternoon and early evening the mean SLAB2 thermodynamic variables much more closely approximate those of the HRLDAS simulations than SLAB1 (Figs. 8a–d).
The afternoon and early evening CAPE differences among simulations (Figs. 8e,f) are dominated by the $q_v$ differences (Figs. 8c,d). This is not surprising since daytime CAPE differences are typically dominated by near-surface moist static energy differences, and $\Delta q_v = 1$ g kg$^{-1}$ produces a moist static energy change comparable to $\Delta T = 2.5$ K (Crook 1996). As for $q_v$, afternoon and early evening CAPE in HRLDAS1, HRLDAS2, and SLAB2 compares more favorably with the EDAS CAPE than does CAPE from SLAB1 (Figs. 8e,f). The close correspondence of the mean midafternoon CAPE among all simulations but SLAB1 (Figs. 8e,f) is a robust feature evident in a majority of individual days (Figs. 9g,h). However, typical CAPE differences between SLAB1 and the other simulations of a few hundred joules per kilogram are dominated by daily variability associated with synoptic-scale horizontal moisture advections, particularly in the east subdomain (Fig. 9h).

b. Sensitivity of mean daytime PBL evolution

The foregoing analysis of the diurnal cycle indicates the greatest differences in near-surface thermodynamic conditions among simulations occur in mid-to-late afternoon, with SLAB1 being the outlying simulation. In the following we illustrate the association of these differences with differences in the preceding surface fluxes. Both HRLDAS1 and SLAB1 mean 1800 UTC surface fluxes show general west–east SH decreases (Figs. 10a and 10b, respectively) and west–east LH increases (Figs. 10c and 10d, respectively). However, in HRLDAS1, regional gradients over the SGP (102°–94°W) are much stronger than in SLAB1.
Both local and regional-scale circulations influence the mean daytime PBL evolution across the SH gradient in HRLDAS1, which we examine at locations indicated by the + symbols in Fig. 10a (HRLDAS1) and Fig. 10b (SLAB1). Over southwest Kansas differences in PBL evolution between HRLDAS1 and SLAB1 are consistent with local surface flux differences. Here, the HRLDAS1 lower troposphere, which is 0–1 g kg\(^{-1}\) drier than SLAB1 at 1500 UTC (Fig. 11c), becomes even drier in the lowest 1 km AGL during the afternoon, while becoming moister than SLAB1 near and above the PBL top (Fig. 11c). This evolution is consistent with weaker LH (cf. Figs. 10c,d) and deeper vertical mixing in HRLDAS1 (Figs. 11a,c) driven by locally stronger SH (cf. Figs. 10a,b). The greater PBL drying from deeper vertical mixing is consistent with Betts and Ball (1995) who attributed decreased afternoon PBL \(\theta_e\) relative to that over moister soils to this effect. Similarly, we find decreased HRLDAS1 afternoon CAPE relative to SLAB1 over this general region (Fig. 8e).

Unlike for southwest Kansas, daytime differences in PBL evolution over southeast Kansas cannot be explained solely by local heating and vertical mixing differences between the simulations. Here, HRLDAS1, which is drier than SLAB1 between 1.5 and 2.5 km AGL at 1500 UTC (as in the west), remains drier than SLAB1 within this layer throughout the afternoon, while significantly moistening relative to SLAB1 within the PBL below (Fig. 11d).

This location where HRLDAS1 experiences a daytime PBL \(q_v\) (Fig. 11d) increase is situated within a relatively narrow longitudinal corridor that behaves similarly (Fig. 12a). Note that this region is well west of the zone of maximum LH (cf. Fig. 10c), suggesting the importance of nonlocal effects such as horizontal moisture transport into the region (Fig. 7a).
The mean vertical profiles over southeast Kansas are within a >2 g kg\(^{-1}\) local maximum of 2100 UTC HRLDAS1 – SLAB1 \(q_v\) differences (Fig. 12b). Horizontal moisture advection differences during the preceding 6 h account for 0.4 g kg\(^{-1}\) or \(\sim \frac{1}{3}\) of this \(q_v\) difference (Fig. 12b) and are consistent with the earlier upstream surface evaporation differences over eastern Oklahoma (Fig. 12c). Local LH differences and static stability differences above the PBL (to be discussed shortly) between simulations may also contribute to the surface \(q_v\) difference.

Warming of \(\sim 1\) K above the PBL between 1500 and 2100 UTC in HRLDAS1 over southeast Kansas (Fig. 11b), which is twice that found for SLAB1, is suggestive of mean daytime subsidence in HRLDAS1. To investigate this possibility, 11-day-averaged vertical motion using hourly model output from 16 to 20 UTC is calculated along vertical cross section W–E in Fig. 10. Prior to time averaging, the scale-selective filter used by Barnes et al. (1996),

\[
F(x, y) = \left[ \frac{\sin(\pi x/D)}{\pi x/D} \right] \left[ \frac{\sin(\pi y/D)}{\pi y/D} \right].
\]

is applied to the vertical velocity field at each grid point on constant pressure surfaces, with \(D = 60\) km selected to emphasize large mesoscale wavelengths. The resulting spatially filtered time-averaged vertical velocity is then averaged for 100 km across transect W–E to produce the vertical velocity, \(\bar{w}\), displayed in Fig. 13.

A well-defined regional-scale vertical motion couplet within the mean vertical cross section in HRLDAS1 (Fig. 13a) is absent for SLAB1 (Fig. 13b). In HRLDAS1, the ascending branch of the vertical circulation occurs over the Rocky Mountain foothills and adjacent high plains with descent farther east. Using \(-\partial\theta/\partial z = -5\) K km\(^{-1}\) (Fig. 11b) and \(\bar{w} = -1\) cm s\(^{-1}\) at \(x = 820\) km (Fig. 13a), the diagnosed subsidence explains the 6-h HRLDAS1 1–2-km potential temperature change, \(\Delta\theta = -\bar{w}(\partial\theta/\partial z)\Delta t = -1\) K (Fig. 11b). Greater subsidence associated with the regional-scale vertical circulation in

Fig. 11. Mean 10–20 Jun midmorning-to-midafternoon (1500–2100 UTC) PBL evolution for HRLDAS1 and SLAB1 simulations. (a) Potential temperature and (c) water vapor mixing ratio at the westernmost location, marked by the plus symbols in Fig. 10, are shown. (b) Potential temperature and (d) water vapor mixing ratio at the easternmost location, marked by the plus symbols in Fig. 10, are shown. The inset in (b) is the HRLDAS1 – SLAB1 difference of potential temperature for the 1–2-km MSL layer discussed in the text.
HRLDAS1 results in greater static stability through a ~1-km-deep layer above the PBL (Fig. 11b, inset). The greater static stability at the PBL top in HRLDAS1 may augment the HRLDAS1 – SLAB1 PBL (and CAPE) differences arising from enhanced HRLDAS1 evaporation and horizontal moisture advection (Figs. 12b,c) by limiting the depth of vertical mixing, but, on the other hand, may impede convective triggering. Differences in PBL depth over the SGP contribute to lower-tropospheric horizontal density gradients that

Fig. 12. (a) Mean 10–20 Jun 6-h evolution of water vapor mixing ratio (g kg$^{-1}$) from 1500 to 2100 UTC for simulation HRLDAS1 at the lowest model grid point. The shaded region indicates moisture increases (with 0.5 g kg$^{-1}$ intervals). Dashed regions indicate moisture decreases with 0.5 g kg$^{-1}$ contour intervals. The plus symbols locate the western and eastern Kansas vertical profile locations from Fig. 11 and the inset is the east subdomain depicted in (b) and (c). (b) Mean 10–20 Jun 2100 UTC HRLDAS1 – SLAB1 water vapor mixing ratio difference (g kg$^{-1}$) with positive values shaded in 0.5 g kg$^{-1}$ intervals starting at 0.5 g kg$^{-1}$ and negative values contoured in −0.5 g kg$^{-1}$ intervals starting at −0.5 g kg$^{-1}$. Bold solid contours are the 10–20 Jun mean 6-h-averaged HRLDAS1 – SLAB1 horizontal moisture advection difference from 1500 to 2100 UTC with values of 0 and 2 × 10$^{-5}$ g kg$^{-1}$ s$^{-1}$. (c) Mean 10–20 Jun 1800 UTC HRLDAS1 – SLAB1 surface latent heat flux difference with positive values shaded in 50 W m$^{-2}$ intervals starting at 50 W m$^{-2}$ and negative values contoured in −50 W m$^{-2}$ intervals starting at −50 W m$^{-2}$; the winds represent the 10–20 Jun mean 6-h-averaged HRLDAS1 horizontal flow from 1500 to 2100 UTC.
drive the time-mean vertical circulation as hypothesized in earlier studies (e.g., Ogura and Chen 1977; Sun and Ogura 1979). Across the center of the mesoscale vertical circulation the PBL depth (AGL) rises 452 m in HRLDAS1 (Fig. 13a) over a 200-km E–W distance (x = 400–600 km), but only 202 m in SLAB1 (Fig. 13b). These differences in the slope of the PBL top are, in turn, associated with differences in the mesoscale patterns of soil wetness (cf. Figs. 1, 2a) through effects on horizontal gradients of SH (cf. Figs. 10a,b).

The SLAB2 mesoscale pattern of soil wetness (Fig. 2c), slope of the PBL depth (368 m over 200 km), and associated mesoscale vertical circulation (Fig. 13c) differ substantially from those of SLAB1 (Figs. 2a and 13b, respectively) but are quite similar to their HRLDAS1 counterparts (Figs. 1 and 13a, respectively). This result points to the importance of differences in the pattern of soil wetness in the initial condition in influencing both differences in the regional-scale vertical circulation and the daytime evolution of near-surface thermodynamic variables.

The E–W horizontal gradient of soil wetness was also found by Bosilovich and Sun (1999) to influence circulations in the 1993 flood case. However, in their simulations its primary role was to impact the meridional LLJ strength. In the current case we find little difference among simulations at the southern boundary of the east subdomain (Fig. 8h) and only small differences of up to 2 m s$^{-1}$ at the southern boundary of the west subdomain (Fig. 8g). Unlike in Paegle et al. (1996), the magnitude of the diurnal oscillation in meridional wind strength is similar among simulations (Fig. 8g), which may partly reflect the less extreme upstream soil moisture differences among simulations in our study.

5. Sensitivity of simulated precipitation

a. Simulated mean diurnal cycle of precipitation and comparison with observations

The mean diurnal cycle of precipitation for observations and each of the four simulations is inferred from

Fig. 13. Vertical cross sections of mean 10–20 Jun 1600–2000 UTC time-averaged and spatially filtered vertical velocity (see text for details) along transect W–E of Fig. 10 for simulations (a) HRLDAS1, (b) SLAB1, and (c) SLAB2. Contour intervals are 0.5 cm s$^{-1}$ with positive values (solid) starting 0.25 at cm s$^{-1}$ and negative values (dashed) starting at $-0.25$ cm s$^{-1}$. Shaded regions indicate upward motions, and the western and eastern plus symbols indicate the locations of the mean vertical profiles from Figs. 11a,c and Figs. 11b,d, respectively. The thick bold line in each panel indicates the 11-day mean 1600–2000 UTC PBL height, which is obtained using the same temporal averaging and spatial filtering as the vertical velocity.
frequency diagrams of area-averaged precipitation rates. Because of the non-Gaussian distribution of precipitation amounts, frequencies of rainfall exceeding a given threshold more reliably capture different modes of the diurnal cycle (e.g., Englehart and Douglas 1985; Dai et al. 1999; Knievel et al. 2004). One threshold (0.04 mm h$^{-1}$) captures all events in which significant hourly precipitation falls somewhere within the subdomains. The other (0.80 mm h$^{-1}$) indicates widespread heavy precipitation. These hourly frequencies are averaged within bins centered on 0000, 0300, 0600, 0900, 1200, 1500, 1800, and 2100 UTC to mitigate noise from small differences in daily timing of precipitation in the relatively short sample.

The west subdomain (Fig. 14a) has a pronounced diurnal cycle for all events (top set of curves) that is similar to observations for this period (dotted curve) in all simulations. Maximum frequencies occur from late afternoon through early evening with minimum frequencies from midmorning through local noon. This diurnal cycle is broadly similar to the diurnal cycle depicted over the high plains in climatological studies of warm-season precipitation (e.g., Carbone et al. 2002). The paucity of major precipitation events (lower set of curves) over this subdomain is consistent with this region having few days with large CAPE (cf. Figs. 9d,e) and it being a generation zone for precipitation systems that tend to intensify and mature farther east.

When all precipitation events are considered, the east subdomain (Fig. 14b) has a less pronounced diurnal cycle than the west subdomain, and possesses a late night–early morning minimum and a broad midday-to-early evening maximum in both simulations and observations. The ~6-h later timing of the diurnal maximum of major events in the observations (Fig. 14b) in the east subdomain is the biggest discrepancy between the observations and the simulations. The reasons for this are unclear and not necessarily related to land surface–atmosphere interactions in the model. For example, differences in the timing of synoptic forcing in a few cases could significantly bias the relatively small sample of events. Since the heavier precipitation events typically have the longest lifetimes and often originate well upstream (not shown), we speculate that model errors in the explicit representation of convection could also contribute to these simulated rainfall phase errors by influencing MCS propagation speeds.

Systematic differences in regional precipitation frequencies are found among the simulations. However, no single simulation most accurately reproduces the observed diurnal cycle of precipitation at all times over both regions (Fig. 14). Typically, the biggest differences are found between SLAB1 and HRLDAS1. SLAB1 has greater overall frequencies of evening precipitation in the west subdomain (Fig. 14a, upper curves) and lesser frequencies of heavy midafternoon and evening precipitation in the east subdomain (Fig. 14b, lower curves). In this regard, SLAB2 much more closely approximates HRLDAS1 than SLAB1. In contrast, SLAB2 has overall daytime precipitation frequencies in the east subdomain that are intermediate between HRLDAS1 and the greater SLAB1 frequencies (Fig. 14b, upper curves).

b. Relationship between precipitation and PBL thermodynamic differences

In the west subdomain HRLDAS1 is always warmer (Fig. 15a) and drier (Fig. 15c) than SLAB1. From midafternoon through late evening HRLDAS1 has less CAPE than SLAB1 for at least 75% of cases (Fig. 15e).
Fig. 15. Distributions of HRLDAS1 – SLAB1 area-averaged differences over the 600 × 600 km² (left) W and (right) E subdomains depicted in Fig. 16: (a), (b) surface $\theta$; (c), (d) surface $q_v$; (e), (f) CAPE; and (g), (h) hourly rain rate. Each area average is indicated by a plus symbol, with the 25th and 75th percentiles of the distributions indicated by dashed and solid lines, respectively.
This contrasts with the east subdomain where HRLDAS1 CAPE exceeds that of SLAB1 (Fig. 15f) and is clearly dominated by $q_v$ differences (Fig. 15d).

The precipitation difference distributions are complicated by some days having little domain-averaged precipitation, which results in negligible precipitation differences on those days. From late afternoon through overnight (2100–0900 UTC), area-averaged precipitation differences $\pm 0.1$ mm h$^{-1}$ typically occur in less than half of the 11 cases (Figs. 15g,h). However, when significant evening precipitation differences do occur they have the same sign as the more general CAPE differences, with the onset of the rainfall differences lagging the onset of the CAPE differences by $\sim 3$ h in the west and $\sim 6$ h in the east. Such lags are expected when local thermodynamics affect precipitation generation, since it often takes several hours for precipitation systems to reach maturity and produce their largest area-averaged rainfall.

The spatial structure of difference fields of thermodynamic variables and precipitation frequencies is illustrated in Fig. 16. The mean midafternoon CAPE differences comprise a dipole structure with extrema located within the previously analyzed west and east subdomains (Fig. 16a). These differences are associated with earlier LH differences (Fig. 16b) through their local and remote (section 4b) influences on midafternoon $q_v$ differences (Fig. 17a).

The mean SLAB2 – SLAB1 2100 UTC $q_v$ differences (Fig. 16c) on the other hand have a similar pattern with amplitudes
only slightly less than HRLDAS1 – SLAB1 (Fig. 17a). Over the SGP, \( q_v \) differences are correlated with soil wetness differences (cf. Figs. 4b, 17b). This correlation reflects the importance of the soil wetness in the mean \( q_v \) differences among SLAB simulations and suggests its importance in the similar mean HRLDAS1 – SLAB1 \( q_v \) differences. In contrast, the mean HRLDAS1 – HRLDAS2 2100 UTC \( q_v \) differences are negligibly small (Fig. 17c) except over highly localized regions, indicating the relative unimportance of the soil moisture evolution. Here, the 12-day length of the simulations is likely too short for the initial soil condition to be “forgotten” and significant coupling–feedback effects to occur.

Overall, the HRLDAS1 – SLAB1 precipitation frequency difference fields for different 2100–0300 UTC precipitation amount thresholds (Figs. 16c,d) exhibit less spatial coherence than the mean CAPE (Fig. 16a) and \( q_v \) (Fig. 16a) difference fields at the start of the 6-h period. However, for the low 6-h precipitation threshold of 0.25 mm (\( \sim 0.01 \) in) coherent mesoscale regions with large frequency differences exist in addition to more random smaller-scale precipitation frequency differences (Fig. 16c). Moreover, several of these mesoscale regions are spatially correlated with the preceding CAPE differences (Fig. 16a). One such region comprises much of the northern part of the west subdomain (Fig. 16c). As noted earlier, the west subdomain is a generation region for precipitation systems and CAPE here is generally small, rarely exceeding 1500 J kg\(^{-1}\) (Fig. 9e). As a result, small-to-moderate CAPE differences in this region may exert an important local influence on whether precipitation becomes widespread.

The region of southeast Kansas examined in section 4b also has precipitation frequency differences. While HRLDAS1 has lesser overall frequencies here than SLAB1 (Fig. 16c), it has 1–2 more 6-h precipitation events >25 mm (\( \sim 1 \) in) during the 11-day period (Fig. 16d). Here, the fewer but heavier precipitation events for HRLDAS1 are supported by the local differences in the thermodynamic conditions. Greater midafternoon stability near the PBL top (Fig. 11b) associated with earlier subsidence (Fig. 13a) could either delay or lessen the total number of events in HRLDAS1. However, significantly greater mean CAPE (Fig. 16a) and \( q_v \) (Figs. 17a) could positively influence precipitation amounts. While the increase in frequency of large rain events is small, it is important because the total rainfall for the period over the east subdomain in all simulations is dominated by relatively few large events (Table 3).
c. Sensitivity to model domain

Sensitivity tests with alternative model domain configurations indicate that at locations well removed from the southern boundary the overall spatial pattern of the HRLDAS1 – SLAB1 differences is robust. In these test simulations (Fig. 18a) an inner domain (termed D02) with 4-km horizontal grid spacing and explicit deep convection is situated within an outer domain (D01) with 12-km horizontal grid spacing that uses the Kain and Fritsch (1993) cumulus parameterization.

Consistent with the findings of Seth and Giorgi (1998), the mean PBL thermodynamic parameters and precipitation from the smaller (single domain) simulations are more accurate than those from the larger domain simulations (not shown). However, the sensitivities among difference fields are broadly similar. For example, the mean 2100 UTC HRLDAS1 – SLAB1 CAPE difference fields from the single-domain runs (Fig. 16a) and the inner domain D02 of the larger-area nested simulation (Fig. 18b) both evince the large-scale dipole structure in the interior with only moderate differences in intensity and location.

6. Summary and discussion

This study examines soil wetness effects on simulated PBL properties and precipitation over the U.S. southern Great Plains (SGP) region. A novel aspect of the current work is the analysis of land surface effects on the PBL and precipitation over a period that is less than seasonal but includes the passage of multiple synoptic-scale weather systems. Much of the precipitation that occurs during a given warm season over the SGP can result from a few such periods, which signifies their importance. In this study we use coupled atmospheric–land surface models that explicitly simulate deep convection for an extended period, which constitutes an additional unique aspect of the work.

The simulations comprise 9–21 June 2002, during which a widespread soil moisture anomaly occurred. Significant differences are found in the atmospheric response during the 12-day period in a set of simulations that used LSMs of varying sophistication and differing detail and date specificity in the initial land surface conditions. Our analysis focuses on the central U.S. region where soil moisture anomalies, and hence differences in the land–atmosphere exchange among the simulations are greatest.

The range of atmospheric responses is illustrated by
comparing two primary simulations in which both the LSM and the initial soil wetness distribution are different. The control simulation (HRLDAS1) uses a version of the Noah LSM and a high-resolution date-specific initial land surface condition (obtained using an offline land surface data assimilation system “HRLDAS”). In contrast, simulation SLAB1 uses a generic soil wetness that is more representative of climatology and a less sophisticated LSM that provides no capacity for soil moisture to evolve.

For a 12-h period that follows the onset of the daytime surface heating, near-surface potential temperature, water vapor mixing ratio, and CAPE from HRLDAS1 were much closer to the EDAS observationally based model analysis values than were their counterparts from SLAB1. Overnight, these accuracy differences between the simulations are less significant and more regionally dependent.

Precipitation differences among simulations are less robust than the PBL thermodynamic differences. This is partly due to the relatively small number of precipitation events during the 12-day simulations, and prevents us from drawing firm statistically based conclusions on the precipitation. Nevertheless, the current results are suggestive of a connection between soil wetness and its horizontal gradients, the afternoon PBL, and subsequent evening precipitation over specific mesoscale regions.

Contributions from local and regional-scale influences of the land surface condition on precipitation vary among locations. Over precipitation generation regions such as the western high plains, where CAPE is frequently marginal and soil wetness is low-to-moderate, precipitation differences among simulations appear to be dominated by local differences in sensible heating and evaporation. Farther east, the relationship between soil wetness, PBL evolution, and subsequent precipitation is more complicated and may also be influenced by regional-scale effects including mesoscale vertical circulations forced by horizontal gradients in soil wetness and large-scale moisture advections influenced by upstream evaporation. Moreover, a significant fraction of the precipitation in this region is associated with propagating convection, which can also be influenced by factors controlling its earlier development in remote locations.

Near-surface thermodynamic differences and associated precipitation differences between the primary simulations HRLDAS1 and SLAB1 can result from 1) differences in the initial land surface condition, 2) differences in the capability for feedbacks from atmospheric inputs (including radiation and precipitation) to influence land surface conditions during the simulation, and 3) differences in the representation of physical processes within the LSMs that calculate the surface fluxes.

Two additional experiments were devised to evaluate the importance of some these differences. A simulation (HRLDAS2) that used the same LSM and initial soil moisture as HRLDAS1 demonstrated that evolution of soil moisture during the 12-day simulation had little effect on regional-scale mean thermodynamic variables and precipitation. In contrast, a simulation (SLAB2) that used the same LSM as SLAB1 but an initial soil wetness condition based on that used in HRLDAS1 indicated substantial differences compared with SLAB1.

The SLAB2 simulation also demonstrates that given a calibrated initial condition with a regional gradient in soil wetness many of the salient features of HRLDAS simulations can be reproduced using a less sophisticated LSM. These features include midday surface fluxes and an associated mesoscale vertical circulation that differ substantially from those of SLAB1. These attributes of SLAB2 contribute to greater accuracy in afternoon near-surface thermodynamic variables and CAPE, which more closely resemble those of the HRLDAS simulations than SLAB1. The similarity in these aspects of SLAB2 to the HRLDAS simulations suggest the primary importance of the initial land surface condition in determining the thermodynamic variability between HRLDAS1 and SLAB1 during and immediately following the strong surface heating phase of the diurnal cycle. The importance of these thermodynamic differences lies in their apparent influence on subsequent evening precipitation during the 12-day period.

The importance of the initial soil wetness compared to its evolution is strongly influenced by the relatively short length of the simulations in the current study. It remains to be investigated what governs the time scale at which soil wetness evolution and coupling–feedback mechanisms could become dominant. This is anticipated to be a function of many factors including the strength of the initial soil moisture anomaly, the large-scale flow pattern and its degree of stationarity, and the precipitation characteristics including organization (isolated or mesoscale) and movement (propagating or stationary), which are in turn influenced by the large-scale flow.

A limitation of the current study is that the experimental design does not allow the roles of the soil moisture anomaly and LSM differences in the different atmospheric responses within the primary simulations (HRLDAS1 and SLAB1) to be completely separated. The recent NCEP hydrology reanalysis dataset at 1° grid spacing would allow a more direct and rigorous test of the role of the soil moisture anomaly on atmospheric
conditions during the 12-day period by enabling a derived climatological land surface condition that could be used in a simulation with the Noah LSM. Such a simulation could be compared to the HRLDAS simulations without potentially complicating effects of different LSM physics. Other important avenues of ongoing and future research include sensitivity studies that examine influences of different PBL parameterizations and more detailed and date-specific green vegetation fraction information in the initial condition on land surface–atmospheric interactions and precipitation.

**Acknowledgments.** This work has benefited from discussions with Rit Carbone (NCAR), Roy Rasmussen (NCAR), David Gochis (NCAR), Thomas Hamill (NOAA), and the comments of two anonymous reviewers. The authors thank David Ahijevych (NCAR) for assistance in performing the linear regression calculations. The image in Fig. 4a is courtesy of NOAA/OAR/ERL PSD, Boulder, Colorado and was obtained from their web site at http://www.cdc.noaa.gov. This work is supported by the U.S. Weather Research Program (USWRP) Grant NSF 01, the NCAR TIMES Water Cycle Program, and the NASA-THP (Dr. Jared Entin; NNG06GH17G) and NASA-GWEC (NNG05GB41G) programs.

**APPENDIX**

**Calibration of Surface Moisture Availability to Reduce Surface Flux Biases in SLAB2**

Large regional biases in the SLAB2 surface fluxes from those of HRLDAS1 occur during the first diurnal cycle when Eq. (1) of section 2 is used to convert soil moisture to the moisture availability variable W used in the SLAB LSM. These biases for the “uncalibrated SLAB2” are listed in Table A1 (middle row). Fortunately, the close relationship between moisture availability and surface fluxes may be exploited to calibrate the moisture availability in such a way that reduces these biases.

The logarithmic ratios of W for simulations SLAB1 and the uncalibrated version of SLAB2 are highly correlated with the logarithmic ratios of latent heat flux (LH; Fig. A1a) and sensible heat flux (SH; Fig. A1b). The points in these example scatterplots are for the shrubland land-use category and include model grid points located ≥200 km from the lateral boundaries of the model domain. They are obtained from an 11-day

![Fig. A1](image_url)

**TABLE A1.** Surface flux biases relative to the HRLDAS1 simulation at 1800 UTC 9 Jun 2002 (t = 6 h) for the entire 2128 × 1540 km² model domain and the 600 × 600 km² W and E regional subdomains shown in Fig. 4b. Absolute differences are in units of W m⁻² and percentage differences from HRLDAS1 are in parentheses.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Entire domain</th>
<th>W subdomain</th>
<th>E subdomain</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ΔSH</td>
<td>ΔLH</td>
<td>ΔSH</td>
</tr>
<tr>
<td>SLAB1</td>
<td>61 (-24%)</td>
<td>65 (+28%)</td>
<td>120 (+37%)</td>
</tr>
<tr>
<td>SLAB2 (uncalibrated)</td>
<td>76 (-30%)</td>
<td>66 (+28%)</td>
<td>107 (-33%)</td>
</tr>
<tr>
<td>SLAB2 (calibrated)</td>
<td>19 (-7%)</td>
<td>1 (+0%)</td>
<td>-37 (-12%)</td>
</tr>
</tbody>
</table>

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Table A2. Regression parameters $a$, $b$, and $c$ [see Eq. (A1) of appendix] used to derive the moisture availability scaling parameter $W_1/W_2$ that converts SLAB2 moisture availability from Figs. 2b–c, and percent coverage of individual land-use categories in Fig. 3 over model domain.

<table>
<thead>
<tr>
<th>Land-use category</th>
<th>$a$</th>
<th>$b$</th>
<th>$c$</th>
<th>$W_1/W_2$</th>
<th>% coverage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Urban</td>
<td>1.0123</td>
<td>−1.002</td>
<td>1.065</td>
<td>0.27</td>
<td>0.53</td>
</tr>
<tr>
<td>Dry land crop/pasture</td>
<td>1.0542</td>
<td>−0.803</td>
<td>1.152</td>
<td>0.78</td>
<td>17.98</td>
</tr>
<tr>
<td>Irrigated crop/pasture</td>
<td>1.1401</td>
<td>−0.817</td>
<td>1.139</td>
<td>0.54</td>
<td>1.10</td>
</tr>
<tr>
<td>Crop/grassland</td>
<td>1.0394</td>
<td>−0.923</td>
<td>1.080</td>
<td>0.59</td>
<td>8.51</td>
</tr>
<tr>
<td>Crop/woodland</td>
<td>1.0634</td>
<td>−0.894</td>
<td>1.145</td>
<td>0.93</td>
<td>6.03</td>
</tr>
<tr>
<td>Grassland</td>
<td>1.0457</td>
<td>−1.095</td>
<td>0.984</td>
<td>0.48</td>
<td>24.63</td>
</tr>
<tr>
<td>Shrubland</td>
<td>0.9968</td>
<td>−0.868</td>
<td>1.067</td>
<td>0.17</td>
<td>17.63</td>
</tr>
<tr>
<td>Savanna</td>
<td>1.0078</td>
<td>−0.967</td>
<td>1.096</td>
<td>0.67</td>
<td>4.55</td>
</tr>
<tr>
<td>Broadleaf forest</td>
<td>1.0337</td>
<td>−0.973</td>
<td>1.117</td>
<td>0.99</td>
<td>6.50</td>
</tr>
<tr>
<td>Needleleaf forest</td>
<td>1.0129</td>
<td>−0.907</td>
<td>1.113</td>
<td>0.67</td>
<td>10.20</td>
</tr>
<tr>
<td>Mixed forest</td>
<td>1.0128</td>
<td>−0.955</td>
<td>1.093</td>
<td>0.71</td>
<td>0.76</td>
</tr>
<tr>
<td>Barren/sparse vegetation</td>
<td>1.0850</td>
<td>−0.976</td>
<td>0.955</td>
<td>0.06</td>
<td>0.13</td>
</tr>
<tr>
<td>Water bodies</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
<td>1.42</td>
</tr>
</tbody>
</table>

(10–20 June) time average at 1800 UTC, the approximate diurnal maximum of the surface fluxes.

Multiple linear regression (e.g., Wilks 1995) is performed for each land-use category in which the logarithmic ratios of moisture availability and surface fluxes for the time-averaged model output are well correlated. Using the regression results, power-law relations of the form

$$
W_1/W_2 = a \left( \frac{SH_1}{SH_2} \right)^b \left( \frac{LH_1}{LH_2} \right)^c \tag{A1}
$$

are derived for each land-use category. The coefficient $a$, exponents $b$ and $c$, and the percentage of the model domain covered by each land-use category are provided in Table A2. Together these land-use categories comprise 99.97% of the model domain.

To calibrate the unmodified $W$ for SLAB2 (Fig. 2b, section 2) to produce fluxes more consistent with those from HRLDAS1 during its first diurnal cycle, $SH_2$ and $LH_2$ from the unmodified SLAB2 and $SH_1$ and $LH_1$ from HRLDAS1 are spatially averaged for each land-use category at 1800 UTC 9 June ($t = 6$ h). The values of $a$, $b$, and $c$ from Table A2 (derived using the 11-day time-averaged model data) are used with these area-averaged fluxes in Eq. (A1) to obtain a moisture availability scaling factor $W_1/W_2$ for the different land-use categories. The appropriate scaling factor is then applied to the unmodified moisture availability $W_2$ (Fig. 2b, section 2) at each horizontal grid point to produce the calibrated moisture availability $W_1$ used in SLAB2 (Fig. 2c, section 2). A scaling factor of 1 was applied for both the 0.03% of the grid that contained land-use categories too limited in coverage for reliable power-law relations to be derived, and for water bodies where $W = 1$.

We emphasize that this approach to modify $W$ is purely empirical and makes no effort to consider differences in the handling of physical processes within the different LSMs. Nonetheless it results in a substantial reduction of the flux biases during the first diurnal cycle over both the entire domain on average and over the smaller regional subdomains introduced in section 2. The improvements in regional flux differences from HRLDAS1 at $t = 6$ h for the calibrated SLAB2 over both the uncalibrated SLAB2 and SLAB1 are quantified in Table A1.

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