Coupling between Large-Scale Atmospheric Processes and Mesoscale Land–Atmosphere Interactions in the U.S. Southern Great Plains during Summer. Part II: Mean Impacts of the Mesoscale

CHRISTOPHER P. WEAVER

Center for Environmental Prediction, Department of Environmental Sciences, Rutgers–The State University of New Jersey, New Brunswick, New Jersey

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ABSTRACT

This is Part II of a two-part study of mesoscale land–atmosphere interactions in the summertime U.S. Southern Great Plains. Part I focused on case studies drawn from monthlong (July 1995–97), high-resolution Regional Atmospheric Modeling System (RAMS) simulations carried out to investigate these interactions. These case studies were chosen to highlight key features of the lower-tropospheric mesoscale circulations that frequently arise in this region and season due to mesoscale heterogeneity in the surface fluxes. In this paper, Part II, the RAMS-simulated mesoscale dynamical processes described in the Part I case studies are examined from a domain-averaged perspective to assess their importance in the overall regional hydrometeorology. The spatial statistics of key simulated mesoscale variables—for example, vertical velocity and the vertical flux of water vapor—are quantified here. Composite averages of the mesoscale and large-scale-mean variables over different meteorological or dynamical regimes are also calculated. The main finding is that, during dry periods, or similarly, during periods characterized by large-scale-mean subsidence, the characteristic signature of surface-heterogeneity-forced mesoscale circulations, including enhanced vertical motion variability and enhanced mesoscale fluxes in the lowest few kilometers of the atmosphere, consistently emerges. Furthermore, the impact of these mesoscale circulations is nonnegligible compared to the large-scale dynamics at domain-averaged (200 km × 200 km) spatial scales and weekly to monthly time scales. These findings support the hypothesis that the land–atmosphere interactions associated with mesoscale surface heterogeneity can provide pathways whereby diurnal, mesoscale atmospheric processes can scale up to have more general impacts at larger spatial scales and over longer time scales.

1. Introduction

The goal of this two-part study is to investigate general questions relating to mesoscale land–atmosphere interactions in the summertime in the U.S. Southern Great Plains (SGP) region. Specifically, how do the synoptic-scale shifts in background meteorology modulate these diurnal, mesoscale land–atmosphere interaction processes, and how important might these mesoscale processes be to the overall large-scale hydrometeorology? Part I (Weaver 2004) focused on case studies drawn from monthlong, high-resolution Regional Atmospheric Modeling System (RAMS) simulations. The purpose of these case studies was to illustrate key points about the evolution of mesoscale surface flux heterogeneity and its influence on mesoscale atmospheric dynamics, convection, clouds, and precipitation.

Let us briefly revisit the findings from Part I. During the three simulated Julys (1995–97), several-day-long wet and dry periods tended to alternate: for example, large-scale organized convection would periodically enter the model domain. Under the influence of this slowly evolving background hydrometeorology, both the mean and the mesoscale spatial variability of the surface water and energy balances also evolved, creating changing patterns of heterogeneity in the surface sensible heat (SH) and latent heat (LH) fluxes. During the dry periods, this surface flux heterogeneity forced strong lower-tropospheric mesoscale circulations. These circulations had a characteristic diurnal life cycle, peaking in the midafternoon and spinning down by the evening. The background large-scale flow affected the pattern and orientation of these coherent roll structures, but did not in general compromise their intensity except (rarely) when persistently strong large-scale subsidence occurred. These circulations were often intense enough that they could be the dominant mesoscale dynamical feature present on a given day, with numerous strong updraft zones with vertical velocities on the order of 1 m s\(^{-1}\) accompanied by broader, gentler compensating downdrafts.
Depending on the vertical thermodynamic structure of the atmosphere, these updrafts could trigger shallow or even deep convection, with associated clouds and precipitation. Persistence of the signature of this convective precipitation in the soil moisture, and hence in the surface flux pattern, influenced the mesoscale atmospheric dynamics over multiple future days. Additional numerical experiments demonstrated that surface heterogeneity sufficient to drive these mesoscale circulations arose even without heterogeneity in preexisting land cover characteristics such as vegetation, that is, solely as a result of spatial variability in rainfall and other atmospheric processes.

In Part II, we shift our focus beyond these case studies to examine the importance of the mesoscale land–atmosphere interactions when considered from a domain-wide perspective over the full, monthlong simulations. Therefore, the emphasis will no longer be on mechanistic descriptions of individual dynamical features on a given day, but rather on spatial averages over the whole domain, time averages over many days, and composites over multiple simulations. Our interest now is in the aggregate impact of the mesoscale processes we have been examining. The hypothesis is that the coupling of the land and atmosphere provides one or more pathways by which the diurnal, mesoscale atmospheric dynamics can scale up to have an impact at larger spatial scales and longer time scales. One such pathway has already been discussed in Part I, namely, the triggering of deep convection. Pathways relating to convection are particularly important to study, since convective precipitation—as individual thunderstorms or organized groupings such as mesoscale convective systems (MCSs)—is a critical component of continental hydrometeorology. For example, Changnon (2001) showed that thunderstorms produce 72% of the total summer rainfall east of the Rocky Mountains and 48% of the annual-mean precipitation in the Mississippi River basin. The role of mesoscale land–atmosphere interactions in this convection has so far received relatively little attention.

Dynamical triggering of convection is a direct pathway by which the mesoscale land–atmosphere interactions can scale up. However, there may also be indirect pathways whereby the mesoscale processes may exert more gradual influences. For example, it has been proposed that, by affecting lower-tropospheric heat and moisture transport over large regions, many individual mesoscale circulations might in aggregate significantly affect the vertical thermodynamic structure of the atmosphere at large scales (e.g., Lynn et al. 1995b; Pielke et al. 1991, 1998; among others). Dalu et al. (2000) used an analytical model to suggest that sea breezes persisting over several days could have a significant impact on atmospheric stability, as compared to diabatic heating and planetary boundary layer (PBL) turbulence.

Following this reasoning, the vertical heat and moisture fluxes associated with the mesoscale circulation cells we examined in Part I might be generally important. These cells have strong vertical motions, penetrating up to altitudes of 3–4 km, and these motions transport a potentially significant amount of heat and moisture. Persistent mesoscale circulation activity over several days—for example, during one of the many dry periods that occurred during July 1995–97—might progressively alter the atmospheric vertical thermodynamic structure. The total rainfall squeezed out of the atmosphere by an individual convective cell embedded within a larger-scale organized system can be a strong function of factors such as the local stability profile (e.g., Chang and Wetzel 1991; Taylor et al. 1997; Taylor and Lebel 1998; among others). By modifying stability and other thermodynamic properties, the dry-period mesoscale circulation activity could conceivably influence convection and rainfall during a subsequent wet period. For example, increased mixing of PBL and free tropospheric air due to the mesoscale vertical motion might consume low-level instability and also help weaken capping inversions, thereby possibly selecting against deep, severe storms in favor of shallow cumulus [e.g., Mahrt (1977); Mahrt and Pierce (1980) for some relevant discussion].

Through these influences, the surface-forced mesoscale circulations may ultimately affect the feedbacks between land and atmosphere. Findell and Eltahir (2003a,b,c), using observations and mesoscale model simulations, found that the properties of the layer at 1–3 km above the surface in the early morning sounding controlled whether a positive or negative soil moisture–convective rainfall feedback would be favored at a given location. The mesoscale circulations we have been discussing could be expected to strongly affect this critical layer. In Part I, we saw how the memory of a previous day’s rainfall event could persist at the land surface through its signature in the soil moisture field. Similarly, the memory of diurnal, mesoscale circulations might also persist in the atmospheric thermodynamic profile, as the signature of their vertical fluxes is mixed upward into the free troposphere over the course of multiple diurnal cycles. These ideas provide background for our investigation here of the mean properties of the mesoscale features we have so far been examining at the process level. It is hoped that one outcome of this and future work will be model improvements. As noted in the introduction to Part I, most investigations of land–atmosphere interactions have so far been based on studies with either global climate models (GCMs) or relatively coarse resolution regional models and are thus limited by unresolved and/or poorly parameterized processes.

For example, in spite of the importance of convective precipitation to the terrestrial hydrologic cycle, correctly simulating the triggering of convection over land continues to be a difficult problem. It is well known that current convective parameterizations are unable to accurately capture the observed diurnal cycle of clouds.
and precipitation over land (e.g., see Dai et al. 1999; Dai and Trenberth 2004; among others). Dai and Trenberth (2004) showed that daytime convection in the Community Climate System Model Version 2 (CCSM2) typically peaks a few hours too early (in the morning rather than the afternoon), with consequences for the intensity of the simulated convection. Models today do not typically account for subgrid dynamical sources of convective triggering like surface-forced mesoscale circulations, sources that themselves have a distinct diurnal cycle. Greater understanding of the relevant mesoscale processes may thus lead to improved parameterizations and improved simulations of clouds and precipitation. In addition, we need to improve how large-scale models represent the nonlinear interactions between the subgrid convective, microphysical, and radiative processes (Randall et al. 2003). Accounting for convective-scale or mesoscale dynamical variables such as vertical velocity can provide a physical basis for prognosing microphysical and radiative process rates (see, e.g., Donner 1993; Donner et al. 2001; Pincus and Klein 2000).

In general, however, quantifying the sub-GCM grid-scale distributions and statistics of relevant variables, and their dependence on GCM-resolved variables, is challenging. Examining high-resolution simulations of convective and mesoscale processes, such as those carried out for this study, is one potentially useful approach.

2. Method

The RAMS simulations are described in detail in Part I and are only summarized here. Briefly, the results shown are from monthlong simulations for July 1995, 1996, and 1997, referred to as CTL95, CTL96, and CTL97, respectively. For each, the simulation domain consists of three nested grids centered on the Atmospheric Radiation Measurement (ARM) Program SGP site: grid 1 covers 2200 km $\times$ 2200 km with 40-km horizontal grid spacing, grid 2 covers 600 km $\times$ 600 km with 10-km grid spacing, and grid 3 covers 200 km $\times$ 200 km with 2.5-km grid spacing. Convection was parameterized with the Kain–Fritsch (1992) scheme, on grids 1 and 2 but not on grid 3. The purpose of these two coarse grids is to downscale the large-scale flow as provided by the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis boundary conditions, imposed in a lateral boundary buffer zone on grid 1. Individual, 1 month spinup runs for the previous June were carried out in order to provide land (and atmospheric) initial conditions, particularly soil moisture. All grids used 37 stretched vertical levels from the surface to 21-km altitude, with more than half the levels in the lowest 3 km. RAMS contains a state-of-the-art, fully interactive soil–vegetation model, in which 11 soil layers down to 2.5-m depth were used. The high-resolution grid 3 is the domain of interest for this study. Throughout the rest of this paper, all means and statistics shown are for the entire 200 km $\times$ 200 km domain covered by grid 3. Hourly model output was used for all results.

In addition to the CTL runs, three additional simulations were carried out, again one each for July 1995–97. These simulations, NOHET95, NOHET96, and NOHET97, differ from the CTL runs in that the vegetation was set to be constant ("tall grass") on all model grids, all finescale heterogeneity in topography was smoothed out, and all other surface features, such as lakes, were also eliminated. The NOHET runs also differed from the CTL runs in that they used horizontally homogeneous soil moisture/temperature initial conditions. Therefore, while the CTL simulations had heterogeneity in both static land cover and soil initial conditions, the NOHET simulations had homogeneous land cover combined with homogeneous soil initial conditions. Thus, they allow us to isolate the development of surface flux heterogeneity resulting solely from spatially and temporally varying meteorological processes such as rainfall events.

3. Results

We begin our examination of the statistics of key mesoscale variables with the time periods surrounding the Part I case studies. We then generalize these results by moving to an examination of the full monthlong periods of the various simulations.

a. Vertical motion and mesoscale fluxes during the case study periods

In this paper we focus most heavily on two quantities: the mesoscale vertical motion field and the mesoscale vertical flux of water vapor. By quantifying their evolution over the course of the simulations, we hope to gain insight into the larger importance of the mesoscale processes.

Beginning with vertical motion, Fig. 1 shows the horizontal mean and horizontal standard deviation of vertical velocity ($w$) in the grid 3 domain for the days surrounding the case 1, case 2, and case 3 time periods, drawn from the CTL95, CTL97, and CTL96 runs, respectively. For case 1, in the mean (Fig. 1a), there is a clear difference between the first part of the period shown, when relatively weak mean subsidence and ascent prevail, and the second part, after 16 July, when the signature of the large-scale-organized deep convective systems is apparent. The standard deviation of $w$ ($\sigma(w)$; Fig. 1d) shows the influence both of this deep convection, and of the landscape-forced mesoscale circulations during 11–15 July as described in Part I. During these days, $\sigma(w)$ reflects the characteristic life cycle and vertical structure of these circulations, with a mid-to-late afternoon peak at around 2-km altitude. It is this enhanced low-level $w$ variability that has the potential to trigger convection, as it did on 13 and 14 July (recall Part I). A key point is that the peak $\sigma(w)$ values are of
the same order of magnitude during both the surface-forced mesoscale circulation period and the deep convection period, clearly illustrating the dynamical significance of the mesoscale flows that arise from the surface flux heterogeneity.

The results for cases 2 and 3 are consistent with those for case 1. For example, in case 2, we see the transition from dry conditions, and landscape-forced mesoscale circulations, to deep convection on 20–21 July and then back again to clear skies and mesoscale circulations shortly after. In case 3, we see the signature of the mesoscale circulations on 2 July as well as the suppression of this mesoscale dynamics during 3–4 July due to the strong mean subsidence.

This vertical motion transports heat and moisture. For example, Fig. 2 shows the mean and standard deviation of water vapor \( q \) and potential temperature \( \theta \) during the case 1 time period. Particularly for \( q \), the upward movement of moist, near-surface air in either mesoscale circulation cells or deep convection is apparent in the mean and especially the standard deviation. For example, during 13–14 July, a band of enhanced \( \sigma(q) \) is clearly visible at about 3–4-km altitude. In contrast, during the strong convection on 16–20 July, the peak is at higher altitudes. Note that relative to the mean, \( \sigma(q) \) is more than an order of magnitude larger than \( \sigma(\theta) \). Now let us quantify the vertical fluxes implied by Figs. 1 and 2.

The role of landscape-forced mesoscale circulations in enhancing vertical atmospheric fluxes has been discussed in previous studies (e.g., Avissar and Chen 1993; Chen and Avissar 1994; Lynn et al. 1995a; Wang et al. 1998; Weaver and Avissar 2001; among others). Here, we calculate the so-called mesoscale fluxes of \( \theta \) and \( q \) from the RAMS output, as follows (see Avissar and Chen 1993):

\[
F_\theta = \rho C_p \left( \epsilon - \frac{\partial w'}{\partial z} \right) = \rho \left( w - \overline{w} \right) (\theta - \overline{\theta}),
\]
\[
F_q = \rho L \epsilon q' = \rho \left( w - \overline{w} \right) (q - \overline{q}).
\] (1)

The overbars represent averaging over the two-dimensional spatial domain (200 km × 200 km) at each time and vertical level, and the primes represent the individual, grid-cell deviations from those means. Here \( \rho \) is air density, \( C_p \) is the specific heat at constant pressure of air, and \( L \) is the latent heat of condensation. As defined in Eq. (1), the mesoscale flux is an eddy flux and occurs because of the resolved motions at the 2.5-km RAMS horizontal grid spacing. It is important to emphasize that, so defined, the mesoscale flux includes the contributions of all resolved processes, for example, both the surface-heterogeneity-forced mesoscale circulations that occur during the dry periods and the deep convection that occurs during the wet periods.

Quantifying these mesoscale fluxes helps improve our understanding of the importance of mesoscale dynamical processes in the summertime SGP. One way to assess their overall contribution to the time- and domain-mean hydrometeorology is to compare them to the fluxes associated with the large-scale-mean flow, for example, the large-scale subsidence that plays a major role in

Fig. 1. Mean and std dev of \( w \) (m s\(^{-1}\)) in the RAMS 200 km × 200 km grid 3 domain, shown separately for the days surrounding the case 1, 2, and 3 time periods from CTL95, CTL97, and CTL96, respectively. Std dev is indicated as \( \sigma(w) \). The time axis shows LST.
regional adiabatic warming and drying. These large-scale-mean fluxes are defined as follows:

\[
M_q = \rho C_p \overline{w \theta}, \quad M_u = \rho L \overline{q}. \tag{2}
\]

Again, the overbars indicate horizontal averaging over the 200 km × 200 km grid 3 domain.

In Fig. 3, \(M_q, F_q, M_u, \) and \(F_u\) are shown again for the days surrounding case 1. (Note that the grayshade bars show different ranges.) For example, for \(q\), the large-scale-mean flux (Fig. 3a) largely follows \(\overline{w}\) (Fig. 1a), while also accounting for the sharp vertical gradient in \(q\) from the surface into the free troposphere. Mesoscale \(q\) flux \(F_q\) (Fig. 3c) resembles the standard deviation of \(w\) (Fig. 1b), but with the peak flux shifted upward somewhat following the peak \(q\) variability (Fig. 2c); \(F_q\) reflects the strong vertical movement of near-surface and lower-tropospheric water vapor, either by surface-forced mesoscale circulations or by deep convection. More or less similar patterns hold for \(M_u\) and \(F_u\), though peak \(F_u\) from the mesoscale circulations occurs at a lower altitude than peak \(F_q\).

What we are interested in here are the relative properties of the mesoscale and large-scale-mean fluxes. For example, it is clear that the mesoscale \(\theta\) flux is probably not very important compared to the large-scale-mean flux: the peak \(M_\theta\) values are at least an order of magnitude greater than the peak \(F_\theta\) values, and often even larger. The situation is different for \(q\), though, as the peak \(M_q\) and \(F_q\) values are closer in magnitude. For example, during the 12–15 July dry period dominated by the surface-forced mesoscale circulations, peak \(F_q\) is perhaps only a factor of 3–5 smaller than peak \(M_q\).

Equally important, it has a different vertical structure, so at certain altitudes it may even be greater than \(M_q\) (and/or of opposite sign). These conclusions hold for the case 2 and 3 time periods as well (Fig. 4; only the \(q\) fluxes are shown).

From Figs. 1–4, we begin to develop an understanding of the characteristic vertical profiles associated with different processes. For example, “signature profiles” of quantities such as the mesoscale fluxes and the vertical motion statistics seem to consistently emerge during the time periods when we know that these mesoscale circulations were active. Such characteristic vertical structure can be considered to be a footprint of the surface-heterogeneity-forced mesoscale circulations. This footprint can be defined more explicitly, as follows.

Recall that in Part I, we used the NOHET simulations to establish that significant spatial heterogeneity in the surface fluxes would develop even in a simulation devoid of any preexisting variability in initial soil moisture/temperature states and static land cover features (e.g., vegetation, topography, and water bodies) simply as a result of processes like spatially variable precipitation falling on the domain and moistening the soil in a nonuniform way. For example, results from NOHET95 for the same time period as case 1 in CTL95 showed that significant heterogeneity did develop, and this heterogeneity was able to create strong, surface-forced mesoscale circulations like those in the CTL run. As a follow-on to these Part I results, Fig. 5 shows \(\sigma(w)\) (Fig. 5a) and \(F_q\) (Fig. 5c) from NOHET95 for 11–15 July. Comparing with the previous figures, the typical vertical
structure of the mesoscale circulations is again apparent, underscoring the Part I findings.

More relevant for our discussion here, though, is the contrast between these NOHET95 results shown in Figs. 3a and 3c with the additional sets of profiles shown in Figs. 5a and 5c with the additional sets of profiles shown in Figs. 5b and 5d. These come from the four homogeneous simulations spawned from NOHET95, as described in Part I, section 3b. To review briefly, each of these used

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**Fig. 3.** Domain-averaged (a) large-scale-mean $q$ flux ($M_q$), (b) large-scale-mean $\theta$ flux ($M_\theta$), (c) mesoscale $q$ flux ($F_q$), and (d) mesoscale $\theta$ flux ($F_\theta$) for the days surrounding the case 1 time period from CTL95. All fluxes are in W m$^{-2}$ and are as defined in Eqs. (1) and (2). Averages are over the grid 3 domain. Note that the gray shade bars show different ranges of values.

**Fig. 4.** Domain-averaged large-scale-mean $q$ flux ($M_q$) and mesoscale $q$ flux ($F_q$) for the days surrounding the case 2 and 3 time periods from CTL97 and CTL96, respectively. All fluxes are in W m$^{-2}$ and are as defined in Eqs. (1) and (2). Averages are over the grid 3 domain.
D ECEMBER 2004 1253

WEAVER

FIG. 5. The $\sigma(w)$ and mean $F_q$ over the grid 3 domain for the case I time period in both NOHET95 and the four homogeneous, 24-h simulations spawned from NOHET95.

atmospheric initial and boundary conditions taken, in turn, from 0600 LST on 11, 12, 13, and 14 July of the NOHET95 run. In addition, each was initialized with domain-average, and thus spatially homogeneous, NOHET95 soil moisture and soil temperature from 12 July, and then run for 24 h. This setup effectively smoothed out all surface heterogeneity at the start of the 24-h period of each run. Part I showed that, as a result of this change, the strong mesoscale dynamical features apparent in NOHET95 on these days never arose. The results shown in Figs. 5b and 5d are consistent with this. Removing the surface heterogeneity, and hence the mesoscale circulations, has the effect of eliminating the enhanced $\sigma(w)$ (Fig. 5b) and the enhanced $F_q$ (Fig. 5d). (Note that, to facilitate display, these four simulations have been artificially strung together in Fig. 5 into a single, 96-h period.)

Figure 6 shows the same Fig. 5 results for NOHET95 and its homogeneous counterparts, but now collapsed into time averages over the whole 96-h period. Also included in Fig. 6, for completeness, are the averages from CTL95 for the same time period. The qualitative difference between the homogeneous profiles on the one hand, and the NOHET95/CTL95 profiles on the other, is immediately apparent. For $\sigma(w)$ (Fig. 6a), there is a strong peak just below 2 km in both the CTL95 and NOHET95 averages. Similarly, CTL95 and NOHET95 $F_q$ (Fig. 6b) peaks at around 3-km altitude with a positive magnitude of $\sim 25$–30 W m$^{-2}$. Also, a thin layer of smaller-magnitude negative flux occurs near the surface. These features are absent or extremely weak in the homogeneous profiles. For example, the average $F_q$ computed from the homogeneous runs has a peak magnitude of $<5$ W m$^{-2}$.

From Fig. 6, it is clear that the presence of mesoscale surface heterogeneity imposes a characteristic signature on the vertical profiles of $\sigma(w)$ and $F_q$. The average CTL95 or NOHET95 vertical profiles, as compared to those from the homogeneous ensemble, can be taken as a first-order estimate of this signature.

Now we attempt to generalize the contribution of the surface-heterogeneity-forced mesoscale dynamics by examining the full simulations, not just the case study periods. The question then becomes, does this footprint emerge in longer-term averages, and if so, in which large-scale meteorological or dynamical regimes?

b. Composite analysis

This section focuses on composite averages, that is, averages over specific times in the simulations chosen to satisfy particular criteria. The goal of this approach is to identify the processes that dominate under different large-scale meteorological and/or dynamical conditions.

For example, as discussed above and in Part I, one way to think about the meteorological variability during the simulated Julys is to divide them into alternating wet and dry periods. Then, “wet” and “dry” composites are produced by averaging the variables of interest over all the respective periods. Table 1 shows such av-
Fig. 6. The $\sigma(w)$ and mean $F_q$ over the grid 3 domain and averaged over the 96-h period from 0600 LST on 11 Jul 1995 to 0600 LST on 15 Jul 1995 (i.e., the period displayed in Fig. 5). (a), (b) Results from CTL95 (solid), NOHET95 (long dashed), and the aggregated homogeneous simulations (short dash).

eraging periods for each of the simulations. The division
into broad blocks of wet/dry days is somewhat subjective,
and it is arrived at by inspection of the domain-
averaged time series of rainfall (as shown in Figs. 2–4
of Part I) and other fields. In general, the resulting composites
are not particularly sensitive to the exact averaging period chosen. Note that, except for July 1997,
both the CTL and NOHET simulations had the same averaging periods for a given month, again reflecting
the importance of the NCEP–NCAR reanalysis boundary conditions in controlling the general behavior of the
model solution.

Figures 7–9 show the wet and dry composites for
July 1995–97, respectively. We focus on the main variables
under discussion: $w$ and the vertical fluxes (both
mesoscale and large-scale mean) of $q$. For convenience,
both CTL and NOHET results are shown on the same
plot, though the discussion centers on the CTL profiles. As noted in Part I, the CTL and NOHET runs for a
given month should not necessarily show very similar
results, and there are in fact significant differences be-
tween their respective sets of composites, but in general
the similarities outweigh the differences.

Beginning with the wet composites for July 1995
(Figs. 7a,c), mean $w$ is everywhere slightly positive (up-
ward), and $\sigma(w)$ resembles the results shown in Fig. 1
for the periods of deep convection, with a broad layer
of elevated $\sigma(w)$ from about 1-km altitude up into the
midtroposphere and a peak at around 8 km. For $M_q$,
we see upward flux due to the large-scale-mean flow peak-
ing at around 2-km altitude. This low-level peak reflects
the rapid decrease in mean $q$ with height. Similar to
$\sigma(w)$, the vertical structure of the $F_q$ profile is again
consistent with deep convection. Above 6 km, $M_q$ and
$F_q$ are roughly equal in magnitude. Moving to the dry composites (Figs. 7b,d), we see
that mean $w$ is slightly negative (downward), whereas
for $\sigma(w)$ the major feature is a distinct peak in the lower
troposphere. As far as the fluxes, $M_q$ is still predomin-
antly upward, but it is strongest right at the surface
and decreases rapidly with height; $F_q$ peaks just below
2-km altitude, and in the 1–3-km layer it is roughly
equal in magnitude to $M_q$. It is obvious that these dry $\sigma(w)$ and $F_q$ profiles reproduce very closely the profiles
shown in Fig. 6. From this we can conclude that, on
average, the surface-heterogeneity-forced circulations
were the dominant mesoscale dynamical process during
the dry days in July 1995 and thus contributed the most
to the composite averages.
FIG. 7. "Wet" and "dry" composite averages for CTL95 and NOHET95 of \(w\), the light gray curves in (a) and (b), and \(\sigma(w)\), the dark gray curves in (a) and (b), along with \(M_q\), the light gray curves in (c) and (d), and \(F_q\), the dark gray curves in (c) and (d). Here \(w\) and \(\sigma(w)\) are in m s\(^{-1}\), and \(M_q\) and \(F_q\) are in W m\(^{-2}\). The averaging periods for the composite profiles are given in Table 1. The solid curves are for CTL95, and the dashed curves are for NOHET95.

Similar composites for July 1996 and 1997 are shown in Figs. 8 and 9. Many of the main features are consistent with the July 1995 results, though there are some differences. The largest difference is in the vertical structure of \(M_q\). For example, in both 1996 and 1997, the dry composite shows mostly downward flux in the lower troposphere, rather than upward or near-zero flux as for 1995. In addition, the wet \(M_q\) composite for July 1997 (at least for CTL97) does not show uniformly upward \(M_q\) as in July 1995 and 1996, but instead shows downward flux between 2 and 6 km.

The \(F_q\) and \(w\) composites are more similar among the 3 months. The signatures of deep convection and surface-forced mesoscale circulations are again generally apparent in the wet and dry composites, respectively, though this is least true for the July 1996 dry profiles. There, the \(\sigma(w)\) and \(F_q\) profiles show a weaker signal. This is consistent with the simulated July 1996 being generally wetter, with correspondingly weaker surface heating, and July 1997 being generally drier, with correspondingly stronger surface heating (recall Table 1 and Figs. 2–4 from Part I), and hence there being less
favorable conditions in 1996 for strong surface-forced mesoscale circulations.

These results are summarized in Fig. 10, which shows the average of the various composite curves from all three CTL simulations. Once again, the dry composites clearly show the characteristic footprint of the surface-forced mesoscale circulations. Another important point is that, on average, $F_q$ is nonnegligible with respect to $M_q$ during both the wet and dry periods. In addition, $F_q$ is in the same direction (upward) as $M_q$ during the wet periods, at all altitudes, whereas during the dry periods, they generally oppose each other in the lower troposphere. In other words, the surface-forced mesoscale circulations that arise under dry conditions partially counteract the tendency of the large-scale flow toward downward moisture flux during these same times.

We get similar results compositing over large-scale-mean vertical motion instead of wet versus dry conditions. For example, Fig. 11 shows composite profiles based on upward and downward large-scale-mean $w$ above the PBL (at 3-km altitude). It is clear that there is significant overlap between wet and ascent periods and between dry and subsidence periods. Once again, we can easily pick out the characteristic signature of the surface-forced mesoscale circulations in the $\sigma(w)$ and $F_q$ profiles (Figs. 11b,d). The main difference compared to Fig. 10 is that, since the compositing is now over $w$, there is no possibility of sometimes mixing upward and downward motions as there was for the wet/dry compositing, thereby reducing mean $w$ and $M_q$. Therefore, the relative contribution of $F_q$ compared to $M_q$ is smaller in Fig. 11 compared to Fig. 10. Nevertheless, on average during the 3 months, $F_q$ is still significant compared to $M_q$. For example, in the layer from 1- to 3-km altitude $F_q$ during the subsidence periods—that is, $F_q$ associated with the surface-forced mesoscale circulations—averages ~15% of $M_q$.

Figure 12 shows the moisture divergence associated with the dry and subsidence composite $F_q$ profiles (i.e., $\frac{\partial F_q}{\partial z}$) from Figs. 10 and 11, respectively. From near the surface to about 2 km, $F_q$ dries the column, while above that, to about 4–5 km, $F_q$ moistens the column. This illustrates the role of the mesoscale vertical motion in transporting moisture from near the surface upward into the top of the PBL and the lower free troposphere. The peak magnitude of this moistening/drying is around 0.5–1.0 g kg$^{-1}$ day$^{-1}$, depending on the atmospheric level. This suggests that daily reoccurring mesoscale circulations over a period of several days, as happened on multiple occasions during the three simulated Julys, could significantly alter the vertical atmospheric water vapor distribution, with implications for stability and convective potential. This effect falls into the category...
of indirect pathways, as discussed in the introduction, for the surface-heterogeneity-forced mesoscale circulations to influence regional hydrometeorology over time. For example, in Part I, we saw how the convection triggered by the mesoscale circulations during case 1 was a function of the changing mean atmospheric thermodynamic profile over the 3-day time period (recall Figs. 11–13 in section 3a). While other processes such as the large-scale dynamics were undoubtedly important in setting this profile, it is likely that the circulations themselves contributed to the modification of the vertical atmospheric thermodynamic structure over these days. The relative roles of the large-scale and mesoscale atmospheric processes in fixing prestorm conditions needs further study.

4. Conclusions

This is Part II of a two-part study of mesoscale land–atmosphere interactions in the summertime U.S. SGP region. The point of view from which these two papers were written is that heterogeneous landscapes, combined with diurnal, mesoscale atmospheric flows and convection, and embedded within a more slowly varying, large-scale hydrometeorological environment, make up a fundamentally coupled system. This work is an attempt to gain insight into how mesoscale land–atmosphere interactions link with the regional-scale dynamics, hydrologic cycle, and climate. Part I (Weaver 2004) focused on case studies drawn from high-resolution RAMS simulations for three 1-month periods, July 1995–97. These case studies highlighted the development of lower-tropospheric mesoscale circulations in response to heterogeneity in the surface fluxes under different background meteorological conditions. This paper, Part II, places these mesoscale dynamical processes in a larger perspective, in order to assess their importance in the overall regional hydrometeorology.

To accomplish this, domainwide spatial statistics (for the 200 km × 200 km high-resolution model domain) of key mesoscale variables such as $w$, $q$, and vertical fluxes were quantified for all the simulations. Examination of these statistics revealed consistent patterns associated with different processes such as deep convection and the surface-forced mesoscale circulations. These processes displayed characteristic vertical profiles of quantities like mesoscale eddy fluxes and the horizontal spatial variability of the vertical motion field. Composite averages of the mesoscale and large-scale mean variables over different meteorological and dynamical regimes demonstrated that, during dry periods in the simulations, or, similarly, during periods of large-scale-mean subsidence, surface-heterogeneity-forced mesoscale circulations tended to dominate, whereas deep convection dominated during wet/ascent periods. Furthermore, at these domainwide spatial scales and over weekly to monthly time scales, the moisture flux associated with the mesoscale processes was nonnegligible compared to the large-scale-mean flux, with a different vertical structure. For example, during dry/subsidence periods, the mesoscale moisture flux due to the surface-forced mesoscale circulations partially compensated for the tendency of the large-scale-mean downward motion to dry the upper PBL and lower free troposphere.

One premise underlying this study is that land–atmosphere coupling is nonlinear, not only because of nonlinearities inherent in the individual components of the system (e.g., evapotranspiration from a vegetated surface, cloud and rain formation, etc.), or because of feedbacks between these components at a given scale, but also because of feedbacks between scales. Implicit in the idea of feedbacks between scales is the existence of pathways that allow the larger-scale processes to influence the smaller-scale ones, and vice versa. The science questions here relate to such scale interactions, that is, between the mesoscale dynamical processes and the large-scale hydrometeorology. The “downscaling” branch of these pathways, that is, the influence of the large scale on the mesoscale dynamics and mesoscale land–atmosphere interactions, includes factors such as the mesoscale variability of the convection embedded in the large-scale-organized rainfall systems that sweep through the domain during wet periods. The nonuniform wetting of the surface resulting from this variability drives time-varying mesoscale heterogeneity in the surface fluxes that in turn drives mesoscale circulations.

The “upsampling” branch, by which these mesoscale processes might be able to influence the large scale, is more difficult to examine. This study identifies two candidate contributing processes: the direct triggering of deep convection by surface-heterogeneity-forced me-
oscale vertical motion and the alteration of the large-scale-mean atmospheric thermodynamic profile due to vertical moisture transport in the mesoscale circulations. The results shown in Parts I and II suggest that both may be important to the overall hydrometeorological evolution of the coupled land–atmosphere system in this region and season.

Clearly, however, more work is needed. Fundamental questions remain to be answered. While some of the impacts of the mesoscale processes have been quantified, their full role in the larger system has not yet been comprehensively evaluated. For example, what is the relative importance of the large-scale-mean vertical atmospheric thermodynamic structure, the mesoscale and local perturbations to this mean profile, the large-scale dynamical forcing, and the mesoscale dynamical forcing to the frequency of occurrence, organization, and intensity of convection? Are significant long-term feedbacks indeed associated with the potential upscaling pathways described above? Answering these and related questions will require additional numerical experiments and analyses. Ultimately, the mesoscale processes discussed here will likely need to be better represented in GCMs.

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REFERENCES


