Dependence of Simulated Precipitation on Surface Evaporation during the 1993 United States Summer Floods

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ABSTRACT

Regional summertime atmospheric conditions of 1993 are analyzed with the University of Utah Local Area Model (ULAM) by nudging boundary values and large internal scales of the local model toward values produced by the Nested Grid Model (NCEP/NOAA) initial analyses and forecasts archived at 6-h intervals. The approach allows the local ULAM to develop finer-scale structures in the precipitation and circulation forecasts than those resolved by the NGM. The study focuses on the influence of surface evaporation upon rainfall and low-level flow in regional simulations. Much of the rainfall simulated in the control experiment occurred from the late afternoon to early morning hours, with a pronounced midday minimum over the flood region. The moisture flux from the south due to the low-level jet (LLJ) provides much of the moisture source for the precipitation, and it is shown that the net moisture influx is significantly larger than the rainfall rate over the flood region. As a consequence, modifications of surface evaporation apparently are relatively more important in changing the buoyancy and resulting LLJ strength than they are in providing additional moisture to the already plentiful moisture influx from the Gulf of Mexico. This suggests that accurate surface evaporation in the Great Plains is necessary for accurate simulation of dynamic support for rainfall.

The LLJ and especially its diurnal oscillation increase for drier surface conditions in the vicinity of the jet core, providing more effective convergence patterns to support rainfall in these cases than in cases of stronger surface evaporation. This appears to be a more important mechanism for rainfall release over the Mississippi River basin than moistening through local evapotranspiration, although the latter also contributes to more rainfall when this moistening occurs downwind of the jet core.

1. Introduction

The 1993 summer flooding in the Mississippi watershed was the worst on record (Kunkel et al. 1994). The flood conditions were accompanied by persistent lower than normal heights over the western United States, while higher than normal heights dominated over the eastern United States (Bell and Janowiak 1995). Mo et al. (1995) studied the physical basis of the 1993 summer circulation and emphasized large-scale dynamical foundations for the flood. They suggested that the pattern may be explained by strong synoptic-scale eddy momentum transport that accelerated a westerly current over the Rocky Mountains, producing a topographic response that was approximately similar to the observed pattern. They also demonstrated that the low-level jet (LLJ) was strong in this and other periods when anomalously low heights persisted over the United States. The present study is a continuation of an earlier investigation by Mo et al. (1995) concerning the dynamical and physical basis of the Mississippi River basin flood of 1993.

Mo et al. (1995) inferred that the heavy rainfall may be at least partly explained by the strong transport of moisture from the Gulf of Mexico by the LLJ and strong moisture convergence downwind of the jet core above the Mississippi River basin. The linkages between the frequency of the occurrence and the strength of the nocturnal low-level jet and heavy rainfall in the continental United States have also been found in the GEOS-1 simulation by Helfand and Schubert (1995) at the Goddard Space Flight Center.
Other types of surface evaporation–precipitation links have also been discussed. Shukla et al. (1990) suggest that deforestation of the Amazon Basin reduces surface evaporation and rainfall.

Cook (1994) uses versions of the general circulation model (GCM) developed at the Geophysical Fluid Dynamcis Laboratory to assess how surface drying perturbs tropical precipitation. She concludes that surface drying tends to increase precipitation rates in the interior of tropical continents due to forcing by low-level convergence caused by surface warming.

McCormick (1986) studied the influence of soil moisture upon model predictions of the LLJ and boundary layer convergence patterns for a case of pronounced Great Plains convection that occurred on Memorial Day weekend of 1984. He found that the LLJ and low-level convergence are supported in models that retain both dry and moist soils. However, the dry soil case produces stronger diurnal oscillations of the low-level flow, and it provides more effective convergence patterns to support the observed nocturnal precipitation pattern. That model only predicted the boundary layer state and could not resolve the effect of atmospheric moistening upon rainfall.

Figure 1a displays observed July 1993 rainfall over the United States obtained from gridded analysis of station data. The gridded data were produced at the Climate Analysis Center/National Centers for Environmental Prediction (NCEP)/National Weather Service (NWS)/National Oceanic and Atmospheric Administration (NOAA) with a resolution of 2° latitude and 2.5° longitude. Over most of the central and northern Great Plains, as well as the Midwest, monthly totals exceed 0.15 m, while over most of Texas and much of Oklahoma monthly totals are less than 0.05 m. The latter, southern Great Plains area, is the entrance region of the southerly moisture inflow (Fig. 1b), as shown by the July mean vertically integrated meridional water vapor transport provided by the Global Data Assimilation System (GDAS) of the NCEP/NWS/NOAA. The small rainfall is consistent with relatively low soil moisture in the southern Great Plains. Consequently, low-level convergence downwind of the jet core may have promoted rainfall as in McCormick’s 1986 study, but land surface moistening along low-level flow trajectories would have been relatively unimportant upwind of the heaviest rain.

This dynamical interpretation of the flood events is at least superficially at variance with studies that show increased precipitation with increased soil moisture and it is more closely aligned with Cook’s (1994) findings. The goal of the present study is to provide a regional high-resolution simulation of the atmospheric state over the flood region using the mesoscale Utah Limited Area Model (ULAM) developed at the University of Utah (Paegle and McLawhorn 1983). The ULAM was used as a high-resolution interpolator, and it was able to simulate the observed heavy rainfall events in the central United States. Various experiments incorporating different surface evaporation conditions were made with the ULAM to examine the relationship between the LLJ and rain in the central United States.

The ULAM is a mesoscale model, which allows specification of outer model states, both through the boundaries using Davies nudging (Davies 1976) and internally by nudging in spectral space. This approach produces realistic large-scale conditions over the area of interest. The model is outlined in section 2. Section 3 presents precipitation forecasts, and section 4 describes experiments in which the surface evaporation is systematically modified over the model domain. These modifications produce stronger rainfall in the flood region for the case of relatively drier southern Great Plains than for the case of relatively higher evaporation rates because the nocturnal low-level jet provides substantially stronger support for rainfall in the former case. Conclusions are summarized in section 5.

2. The ULAM

The earliest description of the ULAM is given by Paegle and McLawhorn (1983). That early version of the model was limited to boundary layer applications and emphasized the prediction of diurnal wind oscillations, nocturnal low-level jets, and topographically modulated flows. It has been used in real-data integrations for LLJ cases over the Great Plains (Arling et al. 1985; McCormick 1986) and in the vicinity of the Alps (Paegle et al. 1984). The model truncation error around
complex terrain is small (Waldron et al. 1996) and it is therefore expected to give accurate results in the present application.

More recent, fully tropospheric implementations of the model are described by Nicolini et al. (1993), Horel and Gibson (1994), and Waldron et al. (1996). These prognostic applications also include precipitation forecasts and validations against observed data.

a. Model equations

The model utilizes hydrostatic and anelastic equations in terrain-following coordinates (Paegle and McLawhorn 1983). Model equations are shown below in differential form:

\[ \frac{\partial u}{\partial t} + \frac{u}{r \cos \phi} \frac{\partial u}{\partial \lambda} + \frac{v}{r} \frac{\partial u}{\partial \phi} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho \cos \phi} \frac{\partial p}{\partial \lambda} - \frac{\rho \cos \phi}{\rho \cos \phi} \frac{\partial Z}{\partial \phi} + F_u, \quad (1) \]

\[ \frac{\partial v}{\partial t} + \frac{u}{r \cos \phi} \frac{\partial v}{\partial \lambda} + \frac{v}{r} \frac{\partial v}{\partial \phi} + w \frac{\partial v}{\partial z} = -\frac{1}{\rho} \frac{\partial p'}{\partial \phi} - \frac{\rho \cos \phi}{\rho \cos \phi} \frac{\partial Z}{\partial \phi} + F_v, \quad (2) \]

\[ \frac{\partial \theta}{\partial t} + \frac{u}{r \cos \phi} \frac{\partial \theta}{\partial \lambda} + \frac{v}{r} \frac{\partial \theta}{\partial \phi} + w \frac{\partial \theta}{\partial z} = F_\theta + H, \quad (3) \]

\[ \frac{\partial p}{\partial z} = -\rho g, \quad (4) \]

\[ \frac{\partial (\rho u)}{\partial z} = - \left( \frac{1}{r \cos \phi} \frac{\partial (\rho u)}{\partial \lambda} + \frac{1}{r} \frac{\partial (\rho \cos \phi)}{\partial \phi} \right), \quad (5) \]

\[ \rho = \rho R T. \quad (6) \]

Here \( u, v, \) and \( w \) are the zonal, meridional, and vertical wind components, respectively. The atmosphere density \( (\rho = \rho + \rho') \) and pressure \( (p = p + p') \) have been decomposed into basic state and deviations (primes) as described by Waldron (1994). Other symbols represent the gravitational acceleration \( g \), the terrain height \( Z_r \), and potential temperature \( \theta \), where

\[ \theta = T \left( \frac{P_0}{P} \right)^{R/C_p}. \quad (7) \]

Here \( P_0 \) is reference pressure (1013.25 mb), \( R \) is the atmospheric gas constant, \( C_p \) is the specific heat of dry air at constant pressure, and \( t, \lambda, \phi, \) and \( z \) are time, longitude, latitude, and height, respectively, while \( H \) in (3) represents heating. The friction terms \( F_u, F_v, \) and \( F_\theta \) can be specified by horizontal and vertical diffusions as

\[ F_u = \nabla_h(K_u \nabla u) + \frac{\partial}{\partial z} \left( K_z \frac{\partial u}{\partial z} \right), \quad (8) \]

with similar forms for \( F_v \) and \( F_\theta \). Here \( K_u \) and \( K_z \) are the horizontal and vertical turbulent conductivities, and \( \nabla_h \) is the horizontal gradient operator. In the current version of the model, the horizontal diffusion coefficient \( K_u \) is defined as

\[ K_u = \alpha (\Delta S)^{\frac{3}{2}} \left[ \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)^2 + \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right)^2 \right]^{\frac{1}{2}}, \quad (9) \]

where \( \Delta S \) is the grid interval and \( \alpha \) is a constant with given value of 0.36. The definition of vertical diffusion coefficient \( K_z \) follows Yamada and Bunker (1989), as described by Waldron (1994). The equation governing specific humidity \( q \) is

\[ \frac{\partial q}{\partial t} + \frac{u}{r \cos \phi} \frac{\partial q}{\partial \lambda} + \frac{v}{r} \frac{\partial q}{\partial \phi} + w \frac{\partial q}{\partial z} = F_q - P + E_a, \quad (10) \]

where \( F_q \) represents turbulent diffusion, \( P \) is condensation rate, and \( E_a \) is evaporation rate in the atmosphere.

b. Model description

The current version of the model includes forecasts of turbulent kinetic energy, solar and longwave radiative heating of the atmosphere and surface, cloud radiation interactions, and stable and convective radiative processes. Convective and stable precipitation are based on the parameterization schemes used in the National Center for Atmospheric Research’s CCM1. Convective precipitation is parameterized via convective adjustment, which relaxes a supersaturated atmosphere toward the moist adiabat, while preserving total moist energy. Solar radiation calculations are performed every time step, but longwave fluxes are computed only once each hour. The radiation processes used in the model are summarized in Nicolini et al. (1993).

Land and water surfaces are distinguished by their roughness height, specific heat, density, and conductivity. Table 2 of Paegle and McLawhorn (1983) provides soil parameters. The conductivity of water is several orders of magnitude larger than that for soil, so that the surface temperature above water shows virtually no change during the forecast. Surface evaporation is specified over land and water. In the present simulations there is no surface evaporation over land at night (e.g., Fig. 11 of Sellers 1987). Over the ocean, the lower boundary condition specifies \( q = q_s \) (saturation value). Evaporation of falling precipitation is also included following the method used in the NCEP eta model in early 1994. This evaporation is allowed below cloud base until liquid disappears or relative humidity falls below 100%.

The horizontal latitude-longitude grid contains 65 \( \times \) 65 grid points, spaced 0.5° latitude by 0.5° longitude, and the southwest corner of the domain is situated at 25°N, 110°W. There are 17 vertical levels positioned
at the surface, 1, 10, 100, 300, 500, 1000, 2000, 3000, 4000, 5500, 7000, 8500, 10 000, 11 500, 13 000, and 14 500 m above the surface. Five levels are used below the surface to predict the soil temperature. The atmosphere and soil are coupled with a heat balance condition at the soil–atmosphere interface:

\[ C_a \rho_a K_i \frac{\partial \theta}{\partial z} = C_s \rho_s K_s \frac{\partial T_{\text{sub}}}{\partial z} + G_i - F_u - \rho_s L_e E = 0, \]

(11)

where \( \rho_a, \rho_s, \) and \( \rho_w \) are air, soil, and water density, respectively, \( C_a \) and \( C_s \) are atmosphere and soil heat capacity, and \( K_i \) is soil conductivity. The parameter \( G_i \) is solar radiation and \( F_u \) is longwave radiation, considered positive for downward and upward flux, respectively. Surface evaporation rate \( E \) is described later, and \( L_e \) is latent heat of vaporization.

The model uses a time step of 150 s, a closed upper boundary condition similar to that imposed by Innocenti et al. (1993), and lateral boundary conditions and relaxation of the large-scale (internal waves 0–2) horizontal wind field to values obtained from the NGM model above 3 km (Waldron et al. 1995).

The present goal is to use the model as a regional, high-resolution space–time interpolator for data extracted from the NCEP Nested Grid Model initial conditions (at 0000 and 1200 UTC) and NGM 6-h forecasts (0600 and 1800 UTC). Thus, the large-scale data are available to the ULAM at 6-h intervals on the 80 km by 80 km NGM grid. The boundary conditions are obtained from the NGM analyses/forecasts interpolated linearly to the time step of the local model. These conditions are applied directly at the lateral and top boundaries of the model, and the method of Davies (1976) nudging is used in a thin layer (five grid points) adjacent to each lateral boundary.

Deviations of predicted fields from the background NGM state are zero on the boundaries, where NGM values are imposed. These periodic fields are projected onto Fourier series that are filtered as described by Waldron et al. (1996). The forecast equations retain terms that nudge the longer resolved internal waves spectrally toward their NGM values.

3. Precipitation forecasts

The model was integrated from 27 June through 31 July using NGM analyses/forecasts to supply large-scale and boundary information. This is labeled experiment 1 and serves as the control. During this period, the circulation pattern exhibits a quasi-stationary trough over the United States (Mo et al. 1995). The lower boundary conditions specify zero flow and surface heat balance as outlined in section 2b.

The surface evaporation rate \( E \), of the GDAS 0–6-h forecasts was used to specify evaporation over land. Evaporation was fixed to a value equal to 1.5\( E \), during the day and zero at night. The resulting daily surface evaporation is displayed in Fig. 2a. Evaporation rates exceed 0.0025 m per day for much of the flood region but are less than 0.001 m per day over the southwestern plains. The GDAS may underestimate the evaporation during the summer season, but there is no other data source available. Most fields were saved every 6 h, but the vertically integrated zonal and meridional water vapor fluxes (\( Q_u \) and \( Q_v \)) were accumulated every time step during the integration.

a. Comparison with the NGM and eta results

Figure 3 displays ULAM precipitation for the first 24 h for the control experiment starting from 0000 UTC 27 June. The four panels show the 6-h rainfall subtotals in the first (1800–0000 CST), second (0000–0600 CST), third (0600–1200 CST), and fourth (1200–1800 CST) 6-h periods. The forecast is typical of all other forecasts for the subsequent week in that it predicts nocturnal precipitation over the central Plains, centered in this case over southwest Iowa.

Closer examination of the forecast low-level wind shows that the model nocturnal rain was supported by the acceleration of a nocturnal LLJ southwest of the rain region (results not shown). In the event of Fig. 3, the 850-mb flow accelerated from about 10 m s\(^{-1}\) at 1800 CST to almost 20 m s\(^{-1}\) during the night. The accompanying low-level convergence enhancement downwind of the jet core induced nocturnal precipitation over Iowa in this forecast similar to the nocturnal precipitation event studied by Nicolini et al. (1993).

Regional forecasts produced by the NGM and eta models at NCEP for this period produced qualitatively similar results. These also predicted substantial nocturnal jet accelerations and rainfall. Figures 4 and 5 display results for NGM and eta model forecasts for 24-h predictions initialized at 0000 UTC. Comparison with the ULAM forecasts of Fig. 3 shows that the rainfall in that model has somewhat patchier, smaller-scale structure and produces locally heavier amounts than the NCEP models. This is consistent with the higher resolution of the ULAM, which uses an approximately 50-km grid size compared to about 80 km for the NGM and eta models. All three models predict the nocturnal precipitation increase around Iowa for this and all other events of the subsequent week.

b. Total precipitation

The resulting total precipitation produced by the ULAM for the analysis period is shown in Fig. 6. Precipitation maxima are centered over Iowa and Ohio. The ULAM simulates well the maximum precipitation amount and location (compare Figs. 6 and 1a), though it underforecasts precipitation over the central Plains and the Midwest. The relatively dry conditions over Texas are also well represented. The low precipitation
Fig. 2. Evaporation rates for experiments 1 (a), 2 (b), and 3 (c). Contour interval 0.0005 m per day.
over the Great Plains is consistent with the low evaporation rate in that area (Fig. 2a). The model generates excessive rain near the eastern boundary. This may be due to the imbalance between the ULAM and the NGM fields.

c. Precipitation time series

Various time averages of the precipitation forecasts have been compared to observed time-averaged precipitation and precipitation fields from the eta and NGM models. The comparisons (results not shown) indicate that the ULAM produces time-averaged precipitation with broadly similar features to the observed precipitation and to precipitation forecasts by the eta and NGM models. The forecast and observed precipitation display strong diurnal fluctuations. The following discussion describes these fluctuations in the ULAM.

The time evolution of 6-h rainfall totals averaged over a subdomain extending over $33^\circ$–$45^\circ$N, $102^\circ$–$85.5^\circ$W during the analysis period is shown in Fig. 7a. This subdomain includes most of the region of the observed flooding. There are approximately 25 local maxima in this time series, corresponding to diurnal rainfall fluctuations, whose maxima generally occur in the late afternoon to early morning, while the minima occur more commonly around midday (compare with evaporation values that are zero at night). Such diurnal oscillations are also locally found. Figure 8 shows the five-week (27 June–31 July 1993) total precipitation accumulated during four 6-h intervals. Maximum precipitation over Iowa is found between midnight and 0600 local time (0600–1200 UTC).

Figure 7a also shows the meridional moisture flux through the south boundary of the subdomain ($33^\circ$–$45^\circ$N, $102^\circ$–$85.5^\circ$W) every 6 h. The flux has been divided by the area of the subdomain and converted to units of meters per day in order to facilitate comparison with the rainfall rates. Approximately 30 maxima appear in this curve. These often correspond...
to the maxima in the precipitation curve and suggest a strong diurnal cycle in the moisture flux, featuring nocturnal maxima. This is evident when diurnal oscillations are removed as shown in Fig. 7b. This is done by using only values at 0000 UTC and interpolating linearly between these values for other times.

It is noteworthy that the moisture influx is substantially stronger than the rainfall during most of the simulation, suggesting that horizontal moisture inflow toward the flood region may suffice as a moisture source for the rainfall. Local surface evapotranspiration could modify the precipitation but may not be as essential as the dynamical processes, allowing the already plentiful moisture supply to be released. The next section addresses this question in greater detail.

4. Evaporation experiments

The suggestion that surface evaporation may not lead to increased rainfall is superficially at variance with studies emphasizing the importance of precipitation recycling. An alternative explanation (McCormick 1988) is that drier surface conditions increase the surface thermal response to the diurnal solar cycle, because less energy is used in latent heat of evaporation and more of the solar heating is realized in surface warming. Cook (1994) advances a similar hypothesis substantiated by GCM experiments.

The warmer daytime surface conditions lead to greater low-level buoyancy and produce stronger LLJs over the southern Plains, with stronger moisture convergence to support rainfall over the central Plains. To test this hypothesis, two experiments were made to study the linkage between the LLJ and rain in the central United States. Both experiments were initialized using the NGM analyses at 0000 UTC 27 June 1993 and were integrated for 14 days. All parameters were those of experiment 1 except for the surface evaporation rate. Figure 2b displays the daily surface evaporation in one experiment where the values everywhere north of 39°N are the same as those in experiment 1.
South of 39°N, \( E_s \) is specified to be 0.005 m in 1 day, which gives a daily surface evaporation substantially greater than the values in the control, particularly in the LLJ entrance region over the southern Plains. This relatively wet surface case will be referred to as experiment 2. The final experiment maintains surface evaporation over the LLJ entry region identical to that of experiment 1, but north of 41°N, \( E_s \) is set at 0.005 m in 1 day. This experiment will be called experiment 3 and the resulting daily surface evaporation is shown in Fig. 2c.

a. Rainfall

Figure 9 displays the total precipitation forecasts of the control and differences between the experiments and the control for this two-week period. It can be seen that the control (experiment 1), which has the smallest soil evaporation below the LLJ core, produces the heaviest rainfall, while experiment 2, which has the largest land–soil evaporation below the LLJ core, produces the least rainfall in Iowa. Experiment 3, with the same surface evaporation rates over the jet core, simulates a slightly shifted rain pattern. Rain in Iowa is reduced, but there is more rain in Illinois and Indiana. Overall there is a regional redistribution of precipitation according to the local evaporation rate.

Figure 10 displays precipitation differences between experiments 2 and 1 accumulated from 0000 to 1200 UTC and from 1200 to 2400 UTC. A strong diurnal signal is evident with largest differences found during late evening to early morning when rainfall is largest.

b. Circulation response

Figure 11a shows the vertically integrated meridional water vapor transport averaged over the two-week period \( \dot{Q}_v \) for experiment 1, with maximum values over Texas and Oklahoma as in observations (Fig. 1b). The LLJ extends from the gulf coast of Texas and Mexico to Oklahoma and then shifts eastward to Missouri. There are small \( \dot{Q}_v \) differences between the experi-
Helfand and Schubert (1995), with maximum wind speed larger in the ULAM experiments than in their simulation. This may be due to heavy rainfall during the two weeks of the current experiments and the higher ULAM resolution than used in their model. The difference (Fig. 13b) shows weaker diurnal wind oscillations in experiment 2 than in experiment 1 from 500 to 1000 m above the surface.

There are about ten instances, mostly nocturnal, where the control produces significantly stronger LLJs than does experiment 2. This causes a strengthened diurnal cycle of moisture transport \((Q_u)\) and its convergence as shown in Fig. 14, favoring nocturnal precipitation occurrences in experiment 1. The effect is also

![Flux (33N) Precip and Evap](image)

**Fig. 7.** (a) Meridional moisture flux (open dots) through the south boundary of the subdomain \((33^\circ-45^\circ N, 102^\circ-105.5^\circ W)\), area-averaged precipitation rate (full dots), and evaporation rate (thin solid line) versus time (at 6-h intervals) in units of 0.01 m accumulated in 1 day. The flux has been divided by the area of the subdomain. (b) Same as (a) but with diurnal variations removed.
Fig. 8. Same as Fig. 6 but for rainfall accumulated during 0000–0600, 0600–1200, 1200–1800, and 1800–2400 UTC.
Fig. 9. (a) Total precipitation simulated by the ULAM for 27 June–10 July 1993 for experiment 1 and differences (b) between experiments 2 and 1 and (c) between experiments 3 and 1. Contour interval is 0.05 m for experiment 1 is with values greater than 0.05 m shaded. Contour 0.02 m also included. Differences contoured every 0.01 m with values less than −0.02 m shaded.
5. Conclusions

Regional summertime atmospheric conditions of 1993 have been analyzed over North America by nesting a high-resolution forecast model within the NGM and nudging boundary values and large internal scales of the local model toward NGM initial analyses and forecasts archived at 6-h intervals. The approach permits the local model (ULAM) to develop finer-scale structures in the precipitation and circulation forecasts than those resolved by the NGM, and the methodology allows reexamination of the physical basis of the 1993 flooding.

The present investigation has focused on the influence of surface evaporation upon the rainfall and low-level flow in the regional simulation. Much of the rainfall simulated in the control experiment occurred from the late afternoon to early morning hours, with a pronounced midday minimum over the flood region. This agrees with observational analyses of the climatology of summer rainfall over the central Plains by Wallace (1975), which display a distinct nocturnal maximum in the region, and provides evidence that this maximum increases with intensity of precipitation over locations such as Iowa.

The northward LLJ moisture flux provides much of the moisture source for the precipitation and appears to be important for its diurnal modulation as suggested in many previous studies (see Nicolini et al. 1993 and references therein). The net moisture influx through the southern boundary is significantly larger than the rainfall rate over the flood region. As a consequence, modifications of surface evaporation apparently are relatively more important in changing the buoyancy—the resulting strength of LLJ and its convergence pattern—than they are in providing additional moisture to the already plentiful moisture influx from the Gulf of Mexico. The effect of evaporation works through the temperature field and not directly through the water vapor distribution. This suggests that accurate surface evaporation in the Great Plains is necessary for accurate simulation of the LLJ and its effect on rainfall production.
Fig. 11. (a) Vertically integrated meridional water vapor transport averaged for 27 June–10 July 1993 for experiment 1 and differences between (b) experiments 2 and 1, and (c) experiments 3 and 1. Contour intervals are (a) 50 kg (m s)^{-1}, and for (b) and (c) are 10 kg (m s)^{-1}.
Fig. 12. (a) Meridional wind component at 500 m above surface averaged between 0600 and 1800 UTC, contoured every 2 m s\(^{-1}\). Values larger than 8 m s\(^{-1}\) are shaded. Meridional wind component differences between (b) experiments 2 and 1 and (c) experiments 3 and 1, contoured every 1 m s\(^{-1}\). Zero contour omitted. Values smaller than \(-2\) m s\(^{-1}\) are shaded.
The LLJ, and especially its diurnal oscillation, increases for drier surface conditions in the vicinity of the jet core, thus providing conditions more conducive to rainfall in these cases than in cases of stronger surface evaporation. This appears to be a more important mechanism for rainfall release over the Mississippi River basin than moistening through local evapotranspiration, although the latter also contributes to more rainfall when this occurs downwind of the jet core.

Results shown represent a relatively small sample of our experiments. Similar experiments were performed for other model settings. All of these display a tendency for stronger rainfall with drier surface conditions over the southern Plains. The rainfall intensity and the degree of its modulation by surface conditions exhibited some sensitivity to day-to-night distribution of surface evaporation and to the type of upper boundary conditions used in the model. The experiments selected for display were performed retaining only daytime evaporation over land surfaces and used a rigid top lid. Model assumptions regarding these processes produced the greatest sensitivity for the predicted rainfall, and the present selection of these boundary conditions produced relatively small precipitation response compared to other boundary options.

The present results are superficially at variance with other model studies that demonstrate increased precipitation with increased surface evaporation. On a global average, for atmospheric equilibrium conditions, precipitation must balance surface evaporation for sufficiently long times, and enhanced global precipitation requires larger global evaporation. In small regions, it is likely that most of the precipitating water originates from water vapor advected from other areas rather than evaporated locally from the surface. In such instances, the dynamical support of advective processes and vertically lifting may be more critical than local surface evaporation to local rainfall. Our results suggest that this was the case for the Mississippi River floods of 1993.

The present results support the hypothesis presented by McCorcle (1988), who suggested that relatively dry conditions preceding a severe nocturnal flood event over Tulsa, Oklahoma, may have contributed to the dynamical support of that event. The southern Great Plains are commonly dry during summer and therefore produce strong diurnal temperature oscillations that support substantial nocturnal jets. Most summers produce much less rainfall than occurred in 1993 because the LLJ is only one of several factors contributing to
Fig. 14. The vertically integrated meridional water vapor transport difference between 0000 and 0600 UTC accumulation and between 1200 and 1800 UTC accumulation averaged for 27 June–10 July 1993. Panels (a), (b), and (c) correspond to experiments 1, 2, and 3, respectively. Contour interval is 30 kg (m s)⁻¹.
rainfall. Other influences, including large-scale synoptic support for ascending motion and anomalous persistence of the synoptic support over a restricted region, are also important for local floods.

Mo et al. (1995) present one explanation of the persistent synoptic support for the 1993 event. Nicolini et al. (1993) show the important feedback effect that the LLJ-triggered nocturnal precipitation has on the nocturnal LLJ. McCorcle (1986) demonstrates substantial sensitivity of the LLJ and its diurnal oscillations to the ambient synoptic-scale circulation, showing that strong LLJs are favored in synoptic situations with stronger cyclonic flows over the Rocky Mountains and relatively dry surface conditions around the jet core. The present study and our earlier investigation (Mo et al. 1995) suggest that these influences and relatively dry surface conditions upwind of the flood all contributed to the anomalous Mississippi Basin rainfall episode of summer 1993.

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