The Effect of Heterogeneous Soil Moisture on a Summer Baroclinic Circulation in the Central United States*

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ABSTRACT

Thermally induced circulations, similar to sea breezes, may be established in the presence of horizontal gradients in soil moisture, soil type, vegetation, or snow cover. The expense of extensive observational networks and the relatively small-scale circulations involved has made examining these circulations very difficult. Recent numerical studies have indicated that sharp gradients in soil or vegetation properties may induce mesoscale circulations in the absence of synoptic forcing.

The current study employed a three-dimensional, hydrostatic mesoscale model to evaluate the effects of horizontally heterogeneous soil moisture and soil type on the passage of a summer cold front in the central United States. Grid-scale condensation, precipitation, latent heat release, and cumulus convection are not accounted for in this model; moisture was affected only by advection, diffusion, and evaporation. Numerical simulations demonstrated that evaporation of soil moisture significantly affected the boundary layer structure embedded in the baroclinic circulation. Although the position of the front was not altered, the thermal and momentum fields were affected enough to weaken the front near the surface. Evaporated soil moisture was advected ahead of the cold front, far from its source region. Moisture convergence was significantly enhanced in several locations, indicating that soil moisture may play an important role in modifying the spatial distribution and intensity of precipitation.

The impact of surface inhomogeneities in soil moisture and soil type on the atmosphere is expected to be highly dependent on the particular synoptic conditions.

1. Introduction

The partition of energy between the sensible and latent heat fluxes at the surface is a fundamental factor that determines the evolution of the planetary boundary layer. The intensity of circulations in the boundary layer is directly related to the magnitude and horizontal variation of the sensible and latent heat fluxes. The components of the surface energy budget at the earth's surface can be significantly altered by terrain inhomogeneities. Thermally induced mesoscale circulations forced by terrain inhomogeneities such as sea and land breezes, mountain and valley winds, and urban circulations have been studied extensively by observational and numerical techniques.

There has been an increasing recognition in the literature that other inhomogeneities in land characteristics may cause important circulations to develop. Thermally induced circulations established near horizontal gradients in soil type, soil moisture, vegetation, snow cover, or cloud cover, may also produce circulations similar in structure and magnitude to sea breezes. A nonclassical mesoscale circulation (NCMC) was defined by Segal et al. (1989) as a circulation produced by soil-moisture gradients, to distinguish it from the sea-breeze phenomena. This term will be used in this paper as well.

Most studies have focused on the potential impact of simple discontinuities in soil type, soil moisture, or vegetation, while neglecting synoptic forcing and three-dimensional effects (Avissar and Pielke 1989; Mahfouf et al. 1987; Ookouchi et al. 1984; Pinty et al. 1989; Segal et al. 1988; Yan and Anthes 1988). Since these studies neglect synoptic forcing, the impact of NCMCs on the boundary layer could be overpredicted in many situations. Large horizontal gradients in land characteristics assumed by these studies are resolved by a grid spacing of 5–10 km. Sharp horizontal gradients in soil moisture are not entirely unrealistic. Strong soil-moisture contrasts can be produced by persistent weather patterns, convective precipitation, topographic influences, and agricultural irrigation.

Synoptic forcing may be large enough to mask or suppress NCMCs in numerical simulations as suggested

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by Segal et al. (1989); however, it is possible that NCMCs interact with larger-scale circulations under the proper circumstances. Mesoscale phenomena primarily forced by instabilities in larger-scale synoptic systems, such as low-level jets, fronts, and convection bands, could be affected by NCMCs. McCorcle (1988) and Fast and McCorcle (1990) initialized a coupled earth–atmosphere numerical model with typical springtime synoptic conditions to examine the effect of inhomogeneous soil–moisture content on the Great Plains low-level jet. The magnitude and structure of the simulated nocturnal jet was very sensitive to sharp soil-moisture gradients. The Penn State–NCAR Mesoscale Model was used to examine the influence of soil-moisture variation in the southern Great Plains and the effect of the Mexican plateau on the evolution and structure of the dryline, elevated mixed layer, and the boundary layer in Lanicci et al. (1987). Variable soil moisture in the southern Great Plains was found to be important in determining differential heating and generation of low-level instability in the prefrontal environment.

While most studies of NCMCs assume homogeneous land characteristics within a grid element 10–100 km wide, large inhomogeneities of soil type, soil moisture, vegetation, and soil type are frequently observed on this scale. Wetzel and Chang (1988) reported that for mesoscale and global numerical models with a grid spacing on the order of 100 km, the subgrid-scale variability of soil moisture may be as large as the total mean available moisture content in a particular region. The effects of soil moisture on the boundary layer may be relatively transient because of the subgrid variability in evaporation rate. Vegetation effects also may be transient because evapotranspiration depends on soil moisture, density of vegetation cover, and stomatal, internal, and root resistance. Soil type remains constant for a particular location, but can vary substantially over a grid element. Avisissar and Pielke (1989) and Wetzel and Chang (1988) have addressed these problems by proposing subgrid-scale heterogeneous surface forcing parameterizations.

The expense of extensive observational networks and the relatively small-scale circulations involved have made observing NCMCs very difficult. Currently, it is easier to simulate the potential impacts of NCMCs with numerical models. The partition of energy between the sensible and latent heat fluxes at the surface is of prime importance in achieving accurate simulations of NCMCs in numerical models. Although mesoscale, synoptic, and climate models have been employing more complex surface energy budgets, the particular parameterization of the surface energy budget may significantly affect the magnitude of smaller-scale circulations, such as NCMCs. Avisissar and Pielke (1989) and Segal et al. (1988) have shown that a more complex representation of the soil layer and vegetation can produce significantly different results from simpler parameterizations. Additional research is necessary to determine how complex surface forcings need to be parameterized in order to adequately simulate the effects of NCMCs. It is apparent that more simultaneous observations of meteorological and biophysical parameters are needed to understand the complex relationships at the earth/air interface and to verify two- and three-dimensional models. This was the motivation for the recent FIFE (Sellers et al. 1988) and HAPEx–MOBILE (Andre et al. 1986) experiments. Some of the initial data from HAPEx–MOBILE can be found in Andre et al. (1988) and Pinty et al. (1989). Segal et al. (1989) presented a set of observations taken over irrigated areas in northeast Colorado along with a few three-dimensional simulations of the flow field. Despite the lack of soil-moisture data, the possible effects of NCMCs on relatively larger-scale circulations can be determined with hypothetical soil-moisture distributions as in Lanicci et al. (1987) and McCorcle (1988). This research will attempt to evaluate the intensity and the horizontal and vertical extent of NCMCs resulting from soil-moisture and soil-type distributions in the central United States. This investigation will differ from previous studies by determining whether a specific NCMC can significantly affect a baroclinic mesoscale circulation. By comparing simulations with, and without, any horizontally inhomogeneous land properties, an estimate of the effect of NCMCs on baroclinic circulations can be obtained. The thermal and moisture interaction of NCMCs in the boundary layer with circulations in the free atmosphere will also be evaluated. The diurnal boundary layer may be altered enough to affect the structure of larger-scale weather patterns such as low-level jets or fronts. Moisture-divergence fields may be altered by NCMCs to change the spatial distribution and intensity of convective precipitation.

The coupled earth–atmospheric numerical model described by McCorcle (1988) has been modified to incorporate baroclinic initial conditions. Section 2 outlines the development of the present mesoscale model used to study NCMC’s embedded in baroclinic circulations.

An observed summer baroclinic circulation of a frontal passage in the central United States is used to initialize the numerical model. This front moved through the central United States during 21–23 June 1989 and is described in section 3. This system produced three regions of scattered showers in the Great Plains, with a few stations reporting moderate rainfall. The surface and upper-level characteristics of thermal, moisture, and momentum fields are presented.

The role of soil-moisture and soil-type parameterizations on boundary-layer and mesoscale circulations are examined by performing several control and sensitivity experiments. The results are presented in section 4. Several simulations are performed with no synoptic flow imposed to examine isolated NCMCs produced
by various soil-moisture and soil-type distributions. Then synoptic flow is imposed to examine potential effects of NCMCs on the baroclinic circulation. Difference fields are calculated for several variables by subtracting the results of the control simulations from the sensitivity simulations. Section 5 presents the conclusions of this study.

2. Numerical model

A hydrostatic, coupled earth–atmosphere numerical model described by McCorcle (1988) and Fast and McCorcle (1990) has been modified to simulate baroclinic mesoscale phenomena. The current version of the model assimilates observed surface and upper-air data to the three-dimensional numerical grid for the initial conditions.

a. Governing equations

The vertical coordinate for the atmospheric governing equations has been transformed from an orthogonal to a nonorthogonal, terrain-following vertical coordinate. The functional form of the vertical coordinate σ is defined by

$$\sigma = s \left( \frac{z - z_G}{s - z_G} \right), \quad (1)$$

where $z$ is the Cartesian vertical coordinate, $s$ is the constant height of the model top, and $z_G$ is the elevation of the terrain. This type of vertical coordinate has been used by several mesoscale models reported in the literature (Pielke 1984).

In addition, the model employs a lower layer of nodes in the domain that are logarithmically spaced. The governing equations are transformed in this layer to a new vertical coordinate $\xi$ defined as

$$\xi = \alpha \ln \left( \frac{\sigma}{\sigma_0} \right), \quad (2)$$

where $\alpha$ is a constant and $\sigma_0$ is the roughness length of the surface. This transformation is retained from the boundary-layer model formulation as described in Paegle and McLawhorn (1983).

The atmospheric portion of the coupled earth–atmosphere model is governed by an anelastic, hydrostatic system of equations. The governing equations are transformed into the nonorthogonal grid system for the atmospheric portion of the model by a procedure similar to Pielke (1984) and are listed in the Appendix.

To more precisely predict surface forcings, the model incorporates forecasts of both moisture and heat fluxes within the soil by using a soil-moisture forecast method similar to that employed by the Air Force Geophysics Laboratory Soil Hydrology Model as described by Mahrt and Pan (1984) and Pan and Mahrt (1987). A prognostic equation for the volumetric soil water content $\eta$ is used that contains terms for hydraulic conductivity, hydraulic diffusivity, evaporation, transpiration, and dewfall. A soil heat-flux equation is used to determine the soil temperature $T_{soil}$. Soil temperature forecasts are dependent on the thermal conductivity, which is highly dependent on soil moisture.

The present numerical model does not contain parameterizations for grid-scale condensation, precipitation, latent heat release, or cumulus convection. Evaporated soil moisture is simply advected by the wind field and there are no feedback processes that could reduce this atmospheric moisture. The effect of heterogeneous soil-moisture distributions on atmospheric circulations can still be examined, while neglecting these feedback processes as shown by Avissar and Pielke (1989), Mahfouf et al. (1987), Ookouchi et al. (1984), and Segal et al. (1988).

The prognostic and diagnostic equations are solved by a combination of finite-difference and finite-element techniques. The advection terms are approximated by a fourth-order scheme as described in Tremback et al. (1987). The vertical diffusion terms are discretized by a finite-element technique based upon Galerkin approximations. A Crank–Nicholson scheme is used to solve the time-dependent terms. The transformed vertical velocity is determined diagnostically by integrating the anelastic continuity equation [Eq. (A8)] from the roughness height to the model top. The vertical velocity $w$ is then determined from Eq. (A9). The hydrostatic equation, Eq. (A3), is integrated from the model top to the surface to determine the deviation pressure $p'$.

b. Boundary conditions

Radiative heating and cooling prescribed at the earth–atmosphere interface is one of the physical forcings in the model. The thermodynamic energy equation and soil heat-flux equation are coupled with the surface heat-balance equation at the soil roughness height $z_0$ to obtain

$$G - F_n + \rho C_p K_h \frac{\partial \theta}{\partial z} - \lambda_s \frac{\partial T_{soil}}{\partial z} - \rho_w L_v E = 0 \quad (3)$$

where $F_n$ is the longwave radiative flux, $\lambda_s$ the soil thermal conductivity, $\rho_w$ the density of water, $L_v$ the latent heat of vaporization, and $E$ the evaporation rate in meters per second. The longwave (terrestrial) radiative flux $F_n$ and the atmospheric flux divergence $Q$ that appear in Eq. (A5) are computed as functions of the water-vapor pathlength integrated through the atmosphere (Paegle and McLawhorn 1983). Equation (3) is a balance of the solar radiative flux, the longwave radiative flux, the sensible heat flux, the soil heat flux, and the evaporative flux. Solar radiation $G$, which appears in the surface energy budget equation, is calculated from
\[ G = 1353 \, \text{W m}^{-2} \times (1 - A)[\sin \phi \sin \delta + \cos \delta \cos \phi \sin(\pi t/12)]\tau \quad (4) \]

where \( A \) is the albedo, \( \phi \) is the latitude, \( \delta \) is the declination, \( t \) is time in hours (that varies in longitude), and \( \tau \) is the transmittance. The declination is a function of Julian day. Clear-sky conditions are assumed in the simulations described in this paper; therefore, \( \tau \) is set to 1.0.

For most simulations in this study, albedo varies according to summertime datasets taken from Matthews (1985). In addition, albedo may be determined as a function of soil moisture as described by Idso et al. (1975). This parameterization is valid only for loam soil and is given by the following relation:

\[ A = 0.31 - 0.34 \frac{\eta}{\eta_s} \quad \eta/\eta_s \leq 0.5 \]
\[ A = 0.14 \quad \eta/\eta_s > 0.5 \quad (5) \]

Equation (5) is used in several sensitivity simulations to examine its effect on boundary-layer circulations.

Temperature continuity is assumed at the roughness height such that

\[ T_{\text{air}} = T_{\text{soil}} \text{ at } z = z_o. \quad (6) \]

A similar continuity relation exists for the moisture flux across the interface so that

\[ W_{\text{air}} = W_{\text{soil}} \text{ at } z = z_o \quad (7) \]
\[ W_{\text{air}} = \rho K_r \frac{\partial q}{\partial z} \quad (8) \]
\[ W_{\text{soil}} = \rho_w E \quad (9) \]

where \( W_{\text{air}} \) and \( W_{\text{soil}} \) are the vertical moisture fluxes in the atmosphere and soil, respectively. At the bottom of the soil layer, the temperature is held to its initial value.

The boundary-layer model described by McCorscle (1988) assumed zero-gradient lateral boundary conditions for all of the prognostic variables. This boundary condition sets the derivative of a prognostic variable normal to a lateral boundary equal to zero. The current mesoscale model makes this assumption for the simulations that neglect synoptic flow.

Time-varying lateral boundary conditions based on objectively analyzed observation fields are used in the baroclinic numerical simulations in this study. When this lateral boundary condition is employed, there may be unwanted numerical instabilities near the lateral boundaries. Several methods have been proposed to remove this numerical noise, such as additional horizontal diffusion near the lateral boundaries; however, the present model used a simple low-pass filter (Pielke 1984) on the three outermost nodes near the lateral boundaries.

Time-varying boundary conditions are used at the model top for all of the prognostic variables in the atmospheric portion of the earth-atmosphere model. For the simulations that neglect synoptic forcing, geopotential height gradients are assumed to be zero at the model top so that no horizontal wind is forced. Potential temperature and specific humidity are held constant in time. For the baroclinic simulations, the prognostic variables are allowed to change in time. At the model top, spline interpolation of observed fields is used to update the prognostic variables and pressure in time. The horizontal wind field at the model top is determined from the geostrophic relationship.

c. Initial conditions

The initial basic-state temperature \( T_0 \) is specified by assuming a vertical lapse rate of 6.5°C km\(^{-1}\) with a sea-level temperature of 298°C. The Poisson equation is used to determine the basic-state pressure. The basic-state density fields are then determined from the equation of state.

For the simulations with no imposed synoptic flow, barotropic initial conditions are used in the mesoscale model. The initial deviation pressure and wind fields are set to zero. The specific humidity is determined by employing the Clausius–Clapeyron relationship and assuming a 75% relative humidity throughout the entire domain. Deviation temperature is then diagnosed from a combination of the equation of state and the virtual temperature relationship. Potential temperature is determined from the Poisson equation. Since there is no synoptic flow in the barotropic simulations, the only forcing will come from the diurnally varying surface energy budget. Variable surface characteristics will

FIG. 1. Model topography, contour interval of 150 m.
cause horizontally inhomogeneous thermal fields to develop which, in turn, will induce circulations.

Initialization techniques using barotropic initial conditions may be adequate when simulating idealized atmospheric circulations; however, baroclinic initial conditions are needed to more accurately simulate realistic events.

For the baroclinic simulations, observed surface and upper-air potential temperature and specific humidity are objectively analyzed to the three-dimensional
model grid. At the model top, pressure is determined by the hydrostatic relationship from the gridded analysis of the observed 300-mb height field. The interior pressure is obtained by the integration of the hydrostatic equation (A3). The initial winds are geostrophic, except below 324 m, where the winds are forced to logarithmically approach zero at the roughness height. In the baroclinic simulations, the synoptic field imposed at the top and lateral boundaries will force the model, in addition to the diurnally varying surface energy

FIG. 3. The surface specific humidity (g kg⁻¹) on (a) 1200 UTC 21 June 1989 and (b) 1200 UTC 22 June 1989.
budget. The thermally induced circulations due to the variable surface characteristics will be modified by the synoptic flow.

An ageostrophic wind field could have been used for the initial conditions; however, a dynamic initialization technique would have been necessary to balance the mass and momentum fields. The model uses a Newtonian nudging technique described by Anthes (1974) and Hoke and Anthes (1976), but a preforecast adjustment period greatly increases the computational

![Map of 300 mb, 12 UTC June 21, 1989](image)

![Map of 300 mb, 12 UTC June 22, 1989](image)

**Fig. 4.** The 300-mb height field (10^4 m) and selected wind barbs for (a) 1200 UTC 21 June 1989 and (b) 1200 UTC 22 June 1989.
Fig. 5. Observed 24-h precipitation on 1200 UTC 22 June 1989. The open circles denote stations reporting a trace to 0.5 in and the filled circles denote stations reporting more than 0.5 in.

Fig. 6. Crop moisture index for 24 June 1989.
time necessary for a single simulation. This technique requires an additional term added to the prognostic equations [Eqs. (A1), (A2), and (A5)] that nudges the numerical results toward the objective analysis of the observed data to bring the dynamic and thermal fields into balance as much as possible. During this adjustment process, the diurnal forcings are removed and the synoptic forcing at the boundaries are held constant in time.

Sensitivity tests with the Newtonian nudging technique showed that a preforecast period of 12 h was necessary to balance the mass and momentum fields. As in Anthes et al. (1982), initialization with unbalanced temperatures and winds produced no discernible increase in noise when compared to simulations that employed balance fields. Since the objective of this study is to qualitatively simulate baroclinic circulations, not forecast observed events, dynamic initialization was not used.

The baroclinic initial conditions of the mesoscale model are based on observations obtained from the Unidata SDM (Scientific Data Management) system (Sherrett and Fulker 1988). A procedure has been developed to create an objective analysis of the horizontal wind components, specific humidity, potential temperature, and height fields from radiosonde data for arbitrary horizontal grids for every standard observation level. This procedure incorporated a single-pass Barnes (Barnes 1964) objective analysis technique with

![Map](image)

**Fig. 7.** Initial volumetric soil-moisture distributions representing the relatively wet and dry regions indicated by the crop moisture index where (a) distribution SM1 comprises of a gradual horizontal soil-moisture gradient and (b) distribution SM2 comprises of a sharp horizontal soil-moisture gradient.

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**Table 1. Summary of the numerical simulations.**

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<th>Synoptic</th>
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<th>Soil type</th>
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<sup>a</sup> ST1 = loan in entire domain  
<sup>b</sup> A1 = albedo from summertime datasets (Matthews 1985)  
<sup>c</sup> SM1 = distribution shown in Fig. 7a  
<sup>d</sup> SM2 = distribution shown in Fig. 7b  
<sup>e</sup> SM3 = 0.284 in entire soil layer  
<sup>f</sup> A2 = a function of soil moisture (Idso et al. 1975)  
<sup>g</sup> ST2 = distribution shown in Fig. 8  
<sup>h</sup> A3 = a function of soil color (Wilson and Henderson-Sellers 1985)
exponential weights. The objective analysis fields used every available radiosonde station in North America with an average grid station spacing of 300–400 km. The objectively analyzed variables are then interpolated to the vertical levels of the mesoscale model. Surface and upper-air data are received continually by the SDM system from a satellite link, so that near real-time simulations of mesoscale circulations can be made. These data are continually archived so that simulations of past events also can be made.

d. Model domain

The domain used in this study is the central United States as depicted in Fig. 1. The model domain employs 25 nodes in both horizontal directions with a grid spacing of 104 km. A time step of 360 s is used.

There are 27 nodes in the vertical in the atmospheric portion of the model, and 10 of those nodes are logarithmically spaced below 324 m. A grid spacing of 557 m is employed between 324 m and the model top at 9793 m. Most mesoscale models set the level of the model top higher in the atmosphere; however, no convective effects are included in this model. The vertical grid in the soil portion of the model consists of 15 soil-temperature computation levels, spaced equally 0.04 m apart, extending from the roughness height \( z_0 \) at 0.04 to 0.52 m below the surface. The soil hydrology consists of a two-layer method to update soil-moisture content. The upper layer is 0.08 m deep and the lower layer extends 0.96 m below the surface. Because temper-

![Fig. 8. Soil-type distribution ST2 for the central United States based on a general soil-type map depicted in Foth and Schafer (1980).](image)

![Fig. 9. Numerical results from dry soil, no-synoptic-flow simulation NS1 2 m above the surface predicted for 1800 LST 21 June. (a) Wind and temperature fields, contour interval of 2°C and (b) specific humidity field, contour interval of 1 g kg\(^{-1}\).](image)

perature forecasts depend on soil-moisture content to calculate soil thermal conductivity and heat capacity, updated volumetric soil-moisture values are interpolated to match the soil-temperature forecast levels.
3. Case description

Model performance and sensitivity to soil-moisture distributions were examined by using a typical summer cold-front passage. The model is initialized with observed surface and upper-air data from 1200 UTC 21 June 1989. The lateral and top boundary conditions incorporate synoptic data every 12 h from 1200 UTC.
1200 UTC 21 June a weak low pressure system in southern South Dakota and northern Nebraska is located on a cold front that extended from northeastern North Dakota to southern Colorado. The front was stationary from Colorado to central California. The strongest southerly surface winds were 8.8 m s\(^{-1}\) in northern Texas, and the strongest northerly surface winds behind the front were 6.0 m s\(^{-1}\) in central South Dakota. A relatively uniform warm air mass existed ahead of the front with a surface temperature between 18° and 23°C. The coldest air temperature of 5°C was located well behind the front in Wyoming and Idaho.

During the next 24 h, the front slowly advanced and weakened as it progressed southeastward across the central United States. At 0000 UTC 22 June daytime heating produced temperatures in excess of 30°C from the desert southwest to southeastern Missouri ahead of the front (not shown). At 1200 UTC 22 June the front extended from northern Minnesota through southwest Missouri to western Texas, and the temperature gradient across the front was greatly reduced. The front continued to move southeastward, but virtually dissipated by 1200 UTC 23 June (not shown).

The evolution of the specific humidity fields at the surface for this frontal passage is shown in Figs. 3a,b. At 1200 UTC 21 June a strong moisture gradient existed just behind the cold front from North Dakota to northern Texas. At 0000 UTC the moisture gradient in the northern plains increased due to advection (not shown). Moist air was advected from the Gulf of Mexico by the southerly winds between the surface and the 700-mb level ahead of the front. Between 0000 and 1200 UTC 22 June the 850-mb flow changed from the south to the southwest, cutting off the moisture advection to the north-central United States.

The 300-mb heights and wind fields, depicted in Figs. 4a,b, reveal a nearly stationary trough over the western United States. During the 24-h period, the trough moved southeastward and deepened slightly. The highest wind speeds occurred over North Dakota and increased from 40 m s\(^{-1}\) on 1200 UTC 21 June to 60 m s\(^{-1}\) on 1200 UTC 22 June. Lack of a rapidly propagating upper-air system was responsible for the relatively slow movement of the surface front.

This synoptic system produced some isolated thundershowers along the front in northeast Minnesota, eastern Nebraska and Kansas, and the panhandle of Texas. The observed 24-h precipitation for 22 June 1989 is plotted in Fig. 5. Cloudy conditions existed along the central and northern portions of the front and in the northern plains during most of the day. Clouds are important modulators of both solar and longwave radiation; however, the present mesoscale model cannot simulate their potential effects on the boundary-layer circulations because the model does not utilize a parameterization for clouds.

This particular case is chosen, not only for the frontal
passage, but also for the soil-moisture conditions present. During most of June 1989, the southeastern United States and the Great Lakes region experienced abundant precipitation. As shown in Fig. 6, these regions were reporting relatively wet soil conditions. Most of the other areas in the central United States were reporting abnormally dry conditions, with excessive dryness reported in Nebraska and southern Texas. For this reason, this case seemed appropriate to examine the potential effects of a soil-moisture distribution on the forecast variables of a baroclinic circulation.

4. Numerical results

This section summarizes some of the simulations that are performed to isolate the mechanisms by which NCMCs could effect larger-scale circulations. This study consists of several control and sensitivity experiments to demonstrate that inhomogeneities in soil moisture, and soil type can significantly modify typical mesoscale circulations. A brief description of these experiments is summarized in Table 1.

One set of control experiments uses the initial conditions described in section 3 for 1200 UTC 21 June 1989. The control simulations of the frontal passage are made with dry, bare soil under clear-sky conditions. The sensitivity experiments are similar to the control simulations, except that soil moisture, soil type, and albedo characteristics are modified.

Another set of control experiments simulate the circulations that develop over the domain without any synoptic forcing. These simulations are made with solar radiation attributes from 21 June. Numerical studies, such as Ookouchi et al. (1984), have shown that
NCMCs may be as significant as the sea-breeze phenomena in the absence of synoptic forcing with a horizontal grid spacing of 10 km. The purpose of this set of experiments is to demonstrate the potential magnitude of NCMCs without the complicating effects of synoptic flow patterns using a horizontal grid spacing of 104 km.

By comparing the sensitivity and control experiments, the effect of the simulated NCMCs can be evaluated in detail. This is accomplished by subtracting the results from the control experiments from the results for the sensitivity experiments. Most of the figures in this section depict these difference fields to demonstrate the structure and extent of the secondary circulations caused by surface inhomogeneities.

As indicated in Table 1, three initial soil-moisture distributions are used. Distributions SM1 and SM2 are depicted in Figs. 7a,b. Soil-moisture distribution SM1 is representative of the relatively wet and dry regions from the 24 June crop moisture index (CMI) (Fig. 6). Wet CMI values are arbitrarily assigned a value of $\eta = 0.35$ (about 80% of the saturation value for loam soil) in SM1. Excessively dry CMI values are assigned a value of $\eta = 0.05$. This distribution then incorporates a gradual horizontal gradient in soil-moisture content between the wet and dry regions. The CMI is available during the growing season every two weeks. The 24 June field is used because it remained relatively constant during June and early July. In SM2, two uniform horizontal initial soil moisture regions of $\eta = 0.284$ (about 65% the saturation value for loam soil) and $\eta = 0$ are assumed. The relatively wet and dry regions are located in the same areas as in Fig. 7a; however, a sharp horizontal gradient in soil moisture is used. Soil-moisture distribution SM3 sets $\eta = 0.284$ in the soil layer in the entire domain.

Estimating initial soil-moisture distributions for mesoscale models can lead to large uncertainties because of large spatial irregularities in the domain of interest and the transience of contrast lines. Only theoretical or plausible soil-moisture distributions can be incorporated into the present mesoscale model. More research is needed to develop routine procedures that assimilate quantitatively the effects of soil moisture into short-range forecasts, such as derived soil-moisture values from satellite data (Wetzel and Chang 1988). Observations of daily variations of soil moisture throughout the United States are needed to verify results from these forecasts.

It is possible to use a subgrid-scale weighting technique similar to Avissar and Pielke (1989) to determine soil type; however, this would require an accurate dataset of soil type for the entire United States [gridded soil texture and albedo data are available from NCAR data archives as discussed in Wilson and Henderson-Sellers (1985)]. Most of the simulations in this paper use loam soil throughout the domain (distribution ST1) since the principle objective is to examine the effect of soil moisture. In several simulations, soil type is allowed to vary horizontally as shown in Fig. 8, but it is homogeneous within a grid cell. This distribution is based on a general soil-type map depicted in Foth and Schafer (1980). When this soil-type distribution is used, albedo is allowed to vary according to soil color as described by Wilson and Henderson-Sellers (1985).

a. No-synoptic-flow experiments

Each of the simulations listed in Table 1 for no synoptic flow are integrated for a period of 48 h, although all of the figures present results of the 12-h forecast. The resulting circulations for the second day were very similar to those from the first 24-h period because soil moisture did not sufficiently dry out during the first day. Neumann lateral boundary conditions were used as described in section 2.

1) Dry-soil simulations

Figures 9a,b depict the wind, temperature, and specific humidity fields for the 12-h forecast valid for 1800 LST 21 June, 2 m above the surface for control experiment NS1. At this time, upslope wind speeds in excess of 2.0 m s$^{-1}$ were predicted in the west-central Great Plains near the surface. The model predicted upslope flow for most of the day and a maximum upslope wind speed of 3.7 m s$^{-1}$ occurred at 1800 LST, 43 to 119 m above the surface in northwest Texas. The wind direction veered during the evening due to Coriolis forcing to produce a nocturnal southerly jet of 5.1 m s$^{-1}$, 119 m above the surface in northwest Texas between 2100 LST and midnight (not shown). By 0600 LST 22 June, a downslope westerly wind was evident. The specific-humidity distribution shown in Fig. 9b did not change considerably during the simulation because the winds were relatively light over most of the period. No significant moisture advection occurred from the Gulf of Mexico.

Holton (1967) demonstrated that the diurnally oscillating slope flow was an important mechanism of the Great Plains low-level jet. Even the relatively large horizontal scales and gentle slopes used in the present study can produce significant slope flows (Fast and McCorcle 1990). The slope flow predicted by the model resembled the type of flow that can occur over smaller terrain features with much steeper slopes.

2) Effect of heterogeneous soil moisture and type

The addition of soil moisture can produce sea-breeze-type circulations when simulated by numerical models using horizontal scales between 5 and 10 km as shown by Avissar and Pielke (1989), Mahfouf et al.
(1987), and Ookouchi et al. (1984). Evaporation of soil moisture also effected circulations with a horizontal scale of 140 km in the boundary-layer model of McCorcle (1988) and in global climate models (Dickinson et al. 1986; Meehl and Washington 1988; Wilson et al. 1987). The differences in the forecast variables due to the various soil-moisture and soil-type distributions in this study are shown in Figs. 10a-f.
duction in temperature had the effect of producing a mesohigh over the regions of moist soil. A weak sea-breeze–like circulation developed near the boundary of the warmer, dry-soil areas and the cooler, moist-soil areas. Wind speeds at 1800 LST differed from the dry-soil simulation by as much as 1.5 m s$^{-1}$ in northwest Texas 119 m above the terrain. The response of the model to soil distribution SM1 as seen in Figs. 10a,b was similar to SM2, except that the resulting NCMC was weaker due to the smaller gradients in soil moisture. The maximum difference in wind speed between the dry-soil simulation was 1.0 m s$^{-1}$ in northeast Arkansas from 43 to 119 m above the terrain.

Evaporation rates are highly dependent upon soil type. Because the water holding and retention properties of soils vary by more than 300% with soil type (Taylor and Ashorof 1972), the soil properties can be significant in surface energy exchange. Evaporation rates in the present soil-hydrology model were highly dependent on soil type as shown in Fast and McCorcle (1990). This was due, in part, to the initialization of soil-moisture content. For instance, relatively large evaporation rates were produced over loam soil when $\eta = 0.284$ in soil-moisture distribution SM2. Evaporation rates were much smaller over clay soil with $\eta = 0.284$ since this soil-moisture content is just above the wilting point ($\eta = 0.208$). Clay soils have very fine pore sizes and consequently more energy was required to evaporate the water from this soil.

The NCMC produced in simulation NS9 using soil-moisture distribution SM2, soil-type distribution ST2, and albedo A2 also demonstrated that evaporation can proceed at different rates for different soil types. As depicted in Figs. 10e,f, evaporation proceeded readily over north Texas, Oklahoma, and Kansas, while less evaporation occurred over the nonloam regions. The horizontal extent of the NCMC was much less than simulation NS3 with uniform loam soil, although the circulation was just as intense.

Soil-moisture distribution SM3 profoundly altered the flow field (not shown) because it resulted in temperature reductions of 1.7° to 3.1° C everywhere in the domain. This damped the magnitude of the slope flow by as much as 3.8 m s$^{-1}$. The reduction in the upslope component is qualitatively similar to the results of the terrain simulations reported in Ookouchi et al. (1984).

The simulations that represent albedo by Eq. (5) according to soil moisture (Idso et al. 1975) in experiments NS5–NS7 did alter the albedo somewhat; however, the horizontal potential-temperature field determined indirectly by Eq. (4) did not differ significantly from experiments NS2–NS4. The forecasted variables produced difference fields similar to those shown in Figs. 10a–f, which indicated that soil-moisture differences were more important than albedo differences.

The time evolution of the potential temperature profile for experiments NS1 and NS3 located in eastern

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**Fig. 12. (Continued)**

The most significant changes in the forecast variables occurred for the sharp moisture gradient of distribution SM2 as seen in Figs. 10c,d. Surface temperatures decreased by as much as 3.0°C in northern Arkansas due to evaporative cooling. Evaporation from the soil layer in the moist regions increased the specific humidity by as much as 5.6 g kg$^{-1}$ in western Arkansas. The re-
Oklahoma is shown in Fig. 11a. At midday, the evaporation of soil moisture reduced the temperature of the mixed layer by 2°C. This reduction in temperature stabilized the boundary layer somewhat and diminished the vertical mixing to lower the boundary-layer height by 500 m at 1200 LST. During the evening, a stable layer 20 m in depth developed due to radiational cooling in the dry-soil simulation. The addition of soil moisture reduced the radiational cooling at night. This resulted in warmer temperatures near the surface and

![Diagram A](image1)

**FIG. 13.** Predicted difference fields (simulation S2 - S1) 2 m above the surface. (a) Wind and temperature difference fields for 1800 LST 21 June, contour interval of 0.5°C, (b) as in (a), but for 0600 LST 22 June, (c) specific-humidity difference field for 1800 LST 21 June, contour interval of 1 g kg⁻¹, (d) as in (c), but for 0600 LST 22 June.
reduced the lapse rate in the lowest 100 m when compared to the dry-soil simulation. The time evolution of the corresponding specific humidity profiles for the same location in Fig. 11b show the increase in moisture evaporated from the soil. At 1200 LST, the dry-soil simulation was slightly moister 880 m above the surface because of the greater strength of the daytime boundary layer that mixed moisture upward. Since the boundary-layer depth was suppressed somewhat in the moist-soil simulations, moisture accumulated near the surface initially, and was not transported above the 324-m level.

These simulations with no synoptic flow indicate the general structure and magnitude of the NCMCs

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**FIG. 14.** As in Fig. 13, except for simulation S3 - S1.
that could develop with the soil-moisture and soil-type distributions used in this study. Now the effect of these circulations on synoptic flows can be evaluated.

b. Synoptic-flow experiments

Each of the simulations listed in Table 1 incorporating synoptic flow were initialized with observed data taken from 1200 UTC 21 June 1989 and were integrated for a period of 48 h. Most of the figures depict results from the first 24-h period when the model was most sensitive to surface inhomogeneities.

Preliminary simulations of this frontal passage where performed with zero-gradient, time-varying, and radiation lateral boundary conditions. Results indicated

FIG. 15. As in Fig. 13, except for simulation S9 - S1.
that time-varying conditions quickly affected the results
in the model interior and contaminated all of the fea-
tures of the observed front. Both time-varying and ra-
diation lateral boundary conditions, as described in
section 2, retained the observed frontal features during
this case; however, the simulated position of the front
was slightly superior when time-varying conditions
were used. Time-varying lateral boundary conditions
are used for all the prognostic variables, except for po-
tential temperature, which used zero-gradient boundary
conditions.

1) DRY-SOIL SIMULATIONS

The results for the dry-soil simulation S1 indicated
that the numerical model was able to qualitatively
simulate the thermal and dynamic fields associated with
this particular front.

Figures 12a–f depict the wind, temperature, specific
humidity, and moisture-convergence fields for the 12-
and 24-h forecasts 2 m above the ground. The front in
simulation S1 has moved to northern Minnesota, cen-
tral Iowa, and on into central Oklahoma and northern
Texas by 1800 LST 21 June as seen in Fig. 12a. The
simulated front advanced 100–200 km farther to the
southeast than the observed front on 0000 UTC 22
June (not shown). A warm pocket of air in excess of
30°C stretched from southern Texas to southern Kan-
sas ahead of the front. The coldest air was located well
behind the front in Wyoming. A sharp moisture gra-
dient was evident near the frontal boundary in the cen-
tral United States in Fig. 12c. The large gradients at
the southeast and northern boundaries resulted from
the lateral boundary conditions employed by the me-
oscale model. At this time, the model predicted sig-
nificantly lower humidities near the boundaries than
the observed values.

Moisture-flux convergence (MFC) is a useful diag-
nostic quantity because it can be used to identify the
locations of potential thunderstorm development
(Waldstreicher 1989). Here, MFC is defined by

$$-q \nabla \cdot V - V \cdot \nabla q = -\nabla \cdot (qV),$$

(10)

where the first term is mass divergence and the second
term is moisture advection. A large positive value of
MFC does not guarantee thunderstorm development
because strong capping inversions may be present. As
described in Waldstreicher (1989), convection often
develops downwind of a MFC maxima, where MFC
increases rapidly in time, and where the gradient of
MFC is increasing.

Figure 12e depicts convergence of moisture along
the southern frontal boundary, with a local maximum
in central Texas. There is a divergence of moisture
along the northern portions of the front in Minnesota.
Most of the convergence or divergence of moisture in
this simulation is due to the first term in Eq. (10).

During the following 12 h the simulated front weak-
ened considerably and moved slightly southeastward
as seen in Fig. 12b. The position of the simulated front
agrees quite well with the observed front on 1200 UTC

![Diagram](image-url)

**Fig. 16.** Predicted MFC difference fields (simulation S3 - S1) 2 m above the ground, contour interval $2 \times 10^4 \text{s}^{-1}$ for (a) 1800 LST 21 June and (b) 0600 LST 22 June. Positive values indicate greater convergence in simulation S3 and negative values indicate greater divergence in simulation S3.
22 June as shown in Fig. 2b. The moisture gradient remained relatively strong in the northern portions of the front, with the driest air located just behind the front in the western plains (Fig. 12d). In Fig. 12f, the convergence of moisture has weakened significantly along the southern portions of the front.

2) EFFECT OF HETEROGENEOUS SOIL MOISTURE AND TYPE

The addition of soil moisture in the sensitivity simulation with a gradual gradient in soil moisture (SM1) cooled the boundary layer over northwest Arkansas.
and northwest Mississippi by 4.5°C (Fig. 13a). The wind field in the moist simulation differed from the dry-soil simulation by as much as 1.7 m s⁻¹ at 12 h and 1.0 m s⁻¹ at 24 h. As with the simulations having no synoptic flow, the effect of soil moisture was to produce a weak mesohigh over the moist regions. Specific humidity increased by as much as 5 g kg⁻¹ in Oklahoma and 6 g kg⁻¹ in western Tennessee (Fig. 13c) after 12 h. During the evening most of the additional moisture evaporated during the day was advected to the front where it converged over Oklahoma, as seen in Fig. 13d.

The results of the soil-moisture distribution SM2 in Figs. 14a–d are similar to those in Figs. 13a–d, except that the simulated NCMC is significantly stronger. The surface temperature was reduced by as much as 5.5°C in southern Oklahoma. The effects of the NCMC in the boundary layer extend as far as 200–300 km north of the soil-moisture gradient into northern Missouri. A larger mesohigh produced wind-speed differences near the surface of 2.6 m s⁻¹ at 12 h in Fig. 14a and 1.2 m s⁻¹ after 24 h in Fig. 14b. The position of the front was not altered by the addition of soil moisture, but the wind-speed modifications resulted in a weakened front near the surface. In the sensitivity simulation, the largest increase in specific humidity was 8 g kg⁻¹ that occurred at 1200 LST 21 June over the moist-soil region southwestern Missouri. Moisture evaporated over the wet region also was advected northward to produce significantly higher humidities far from the moisture-transition region. Evaporated moisture converged to the front much sooner than in simulation S2 as seen in Figs. 14c–d because of the greater moisture availability in soil-moisture distribution SM2.

The effects of horizontally varying soil-type and soil-moisture distribution of simulation S12 are shown in Figs. 15a–d. As with the corresponding simulations having no synoptic flow (NS12), the structure of the NCMC was significantly different from those cases that used uniform loam soil. Even though the area of intense evaporation was smaller, significant specific humidity differences developed after 12 h as seen in Figs. 15c–d. Once again the evaporated moisture converged into Oklahoma ahead of the front, but in slightly less quantities than for the other soil-moisture distributions.

The NCMCs resulting from the soil-moisture gradients in Figs. 13–15 are significantly stronger to those produced in the absence of synoptic forcing (Fig. 10). This was due, in part, because of the specification of the initial conditions in the two sets of experiments. The synoptic-flow simulations incorporated a more realistic initial temperature distribution. During the afternoon periods of the model integration the predicted surface temperatures in the southern plains were as much as 7°C warmer than the no-synoptic-flow simulations; therefore, evaporation occurred at a much higher rate in the synoptic-flow experiments.

Eventually forced stronger NCMCs to develop. The horizontal distribution of the soil moisture in the sensitivity simulations also had an effect on the strength of the NCMCs. Drier, warmer air was advected into the southern plains. This situation leads to an intensification of the horizontal pressure gradients that intensifies the nonclassical circulation. These latter mechanisms were also illustrated in the synoptic flow simulation in Avissar and Pielke (1989).

The modification of the specific humidity and wind fields by the presence of soil moisture also altered the MFC fields in the central United States for case SM2 (sharp moisture gradient) as seen in Figs. 16a,b. The interaction of the synoptic circulation and the NCMC enhanced the moisture convergence just behind the front in Kansas and western Iowa, and ahead of the front in Iowa, Wisconsin, and northern Illinois after 12 h. The enhanced divergence in the southern states demonstrates that moisture evaporated in those regions advected north towards the front. The enhanced convergence is the same order of magnitude as the moisture convergence in the dry-soil simulation S1. An important difference from the dry-soil simulations is that a significantly greater portion of the MFC predicted by Eq. (10) is now due to the moisture-advecting term. Observations have shown that moisture advection can significantly contribute to the development and subsequent intensification of storms (Bothwell 1986).

The time evolution of the specific humidity 880 m above the surface is shown in Figs. 17a–d. After 6 h, the specific humidity in the moist-soil simulation is less than the dry-soil simulation at this level because the cooler boundary layer reduced the mixing as in Fig. 11a. By 12 h, daytime heating has allowed the boundary layer to grow above 880 m so that the additional moisture is transported from the surface. The maximum value is 3.5 g kg⁻¹, about a 30% increase at this level, in northeastern Oklahoma and southeastern Kansas. During the next 12 h, this additional moisture was advected northeastward ahead of the front into northern Wisconsin.

While the most profound modifications in the boundary-layer structure occurred near the surface, the NCMCs caused by the soil-moisture distribution were noticeable up to 2500 m above the terrain. The vertical cross-section plots in Figs. 18a–d demonstrate a specific humidity increase up to 1 km above the terrain after 12 h. During the evening, vertical mixing was reduced, but some of the additional moisture appears to have reached 2500 m above the surface. This was probably due to the passage of the front and synoptic-scale vertical motions ahead of the front that transport this moisture. Near the surface, the evaporated moisture was advected 300 km north of the soil-moisture–contrast zone as seen in Fig. 18c.

The vertical cross-section plots in Figs. 18e, f clearly depict the reductions in temperature in the boundary
layer. The changes in the temperature profile are not constant even though the initial soil-moisture region for this case was uniform (SM2).

As in the no-synoptic-flow simulations, albedo calculated by Eq. (5) in experiments S5–S7 was somewhat altered by soil moisture; however, the horizontal potential temperature field did not differ significantly from experiments S2–S4. The forecasted variables produced difference fields similar to those shown in Figs. 10a–f.
The time evolution of the potential temperature and specific-humidity profiles for simulations S1 and S3 are shown in Figs. 19a,b for a point in eastern Oklahoma. The effect of moisture on the temperature and specific-humidity profiles over a moist surface were similar to the corresponding no-synoptic-flow simulations in Figs. 11a,b, except that the effect was larger. The temperature in the daytime was reduced by as much as 4°C. The boundary-layer heights in the sensitivity simulation were as much as 500 m lower than the dry-soil simulation because the reduced temperature at the surface suppressed vertical mixing near the surface.

The time evolution of the potential temperature and the specific humidity over northern Illinois, a dry region, are shown in Figs. 20a,b. Moisture was not advected to that area until after 12 h. The most significant increase in specific humidity occurred after 18 h. The potential temperature was not reduced significantly because daytime heating before the advection of moisture kept the mixed layer relatively warm.

5. Conclusions

Atmospheric processes are inherently connected to energy exchanges at the ocean and earth surface. The observation and numerical prediction of NCMECs has received growing attention in the research literature because they may be as important as other more thoroughly examined mesoscale phenomena, such as sea and land breezes, mountain and valley winds, and urban circulations. The presence of soil moisture or vegetation is expected to modify the surface thermal fluxes when compared to a bare-soil surface under the same environmental conditions. Two- and three-dimensional numerical studies have indicated that horizontal discontinuities in soil moisture or vegetation could induce significant discontinuities in surface thermal forcing and, consequently, mesoscale circulations. Most numerical studies have simulated the resulting mesoscale circulations, that are similar to sea breezes, with horizontal grid spacing of approximately 5–15 km with no imposed synoptic flow. Such circulations may play an important role in patterns related to local meteorology and climatology, cumulus convection, and air quality.

A major task of this research has been to expand the coupled earth–atmosphere model described by McCorcle (1988) in order to include dynamics above the boundary layer, baroclinic initial conditions, and various boundary conditions. These changes were necessary to examine the effect of surface inhomogeneities on the thermal and momentum properties of baroclinic circulations. The mesoscale model is governed by an anelastic, hydrostatic system of equations that are transformed to a nonorthogonal grid system. For the baroclinic simulations in this study, the lateral boundary conditions varied in time and were based on the objective analysis of observed data. The prognostic variables at the model top also varied in time and were determined from an objective analysis of observed data, except for the horizontal wind components that were set equal to their geostrophic value.

Observations from 1200 UTC 21 June 1989 of a frontal passage were used to initialize the three-dimen-
sional model. This particular case was chosen, not only for the frontal passage, but also for the horizontal distribution of abnormally dry and wet soil-moisture conditions present. The sharp horizontal variations in soil moisture indicated that surface inhomogeneities may significantly affect the thermal, moisture, and momentum fields associated with this front.

Two sets of soil-moisture numerical experiments were executed to determine the magnitude and structure of the simulated NCMCs. One set of experiments consisted of several soil-moisture and soil-type distributions with no imposed synoptic flow. The second set of experiments used the same surface characteristics, except that baroclinic initial conditions were used.

Numerical results from the no-synoptic-flow experiments showed that soil-moisture and soil-type distributions could significantly affect the boundary layer even for relatively large horizontal scales. Evaporation from the soil increased the specific humidity by as much as $6.1 \text{ g kg}^{-1}$ and cooled the surface by as much as 3.0°C. The NCMC resembled a mesohigh wind field with a magnitude of 1.0–2.0 m s$^{-1}$. This altered the wind direction and speed of the slope flows over the terrain in the central United States. The effects of evaporation on the thermal and moisture fields were observed up to 1 km above the terrain.

The evaporation of soil moisture also affected the boundary layer structure embedded in the baroclinic circulation. Evaporation from the soil increased the specific humidity by as much as 10 g kg$^{-1}$ and lowered the surface temperature by as much as 6°C. As in the no-synoptic-flow experiments, a mesohigh wind field was produced by the altered thermal field with wind speeds between 1.5 and 3.0 m s$^{-1}$ near the surface. Some studies have indicated that significant synoptic flow patterns could mask or reduce the potential effects of surface inhomogeneities. In this study, soil-moisture and soil-type distributions in the synoptic-flow exper-
iments were found to have an even greater effect than in the no-synoptic-flow experiments; however, some of this effect was due to the different initial conditions used in the two sets of experiments. While the most significant effects occurred near the surface, evaporated soil moisture was advected horizontally far from its source and transported vertically into the free atmosphere by nonlinear synoptic-scale circulations.

Moisture flux convergence was used in this study to demonstrate the potential impact of horizontally heterogeneous soil moisture on the spatial distribution and intensity of precipitation. This could be examined in more detail by mesoscale models that include grid-scale condensation, precipitation, latent heat release, and cumulus convection parameterizations; nevertheless, the possible effects of soil-moisture distributions on mesoscale circulations can still be addressed using the present mesoscale model. It is important to note that the present mesoscale model does not contain a cumulus parameterization that might act as a feedback mechanism for the evaporated soil moisture; therefore, the magnitude of the resulting NCMCs may be overpredicted.

This research also demonstrates the need for routine, accurate observations of soil moisture content and distribution in the United States. These data are necessary because the parameterization of horizontal heterogeneous land characteristics in operational models may significantly influence short-range forecasts. Global climate models have shown considerable sensitivity to drastic changes in the formulation of soil evaporation and evapotranspiration; therefore, local climatological changes may not be predicted correctly. More routine soil-moisture observations on the horizontal scale are needed to verify two- and three-dimensional simulations of NCMCs. The parameterization of the effects of surface inhomogeneities in studies reported in the literature vary in complexity, and it is not clear how detailed a model needs to be to adequately simulate the energy and moisture exchanges at the soil–atmosphere interface.

The present mesoscale model does not incorporate
a vegetation-layer parameterization. Evaporation from the bare-soil surface could be greatly reduced, depending on the type of vegetation, due to shading. In addition, evapotranspiration from a vegetation layer can increase the moisture and decrease the temperature near the surface. Horizontally inhomogeneous distributions in plant types would greatly increase the complexity of energy and moisture exchanges between the surface and the atmosphere.

It is anticipated that the present mesoscale model will be used in the future to simulate mesoscale flow patterns with observed atmospheric and soil-layer data. This would require executing the model with a much smaller spatial resolution so that data from experiments such as HAPEX—MOBILHY could be employed. Possible forecast errors due to initial conditions, boundary conditions, grid resolution, and surface parameterizations could be evaluated in more detail.

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APPENDIX

Governing Equations

The governing equations described by McCorcle (1988) have been transformed into the nonorthogonal grid system for the atmospheric portion of the model and are

\[
\frac{Dq}{Dt} = \frac{\partial q}{\partial t} + u \frac{\partial q}{\partial \lambda} + v \frac{\partial q}{\partial \phi} + \omega \frac{\partial q}{\partial \xi} + \left( \frac{s}{s - z_G} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial q}{\partial \xi} \right) + \nabla \cdot (K_d \nabla q) \tag{A4}
\]

\[
\frac{D\theta}{Dt} = \frac{\partial \theta}{\partial t} + u \frac{\partial \theta}{\partial \lambda} + v \frac{\partial \theta}{\partial \phi} + \omega \frac{\partial \theta}{\partial \xi} \tag{A5}
\]

\[
\frac{D\varepsilon}{Dt} = \frac{\partial \varepsilon}{\partial t} + u \frac{\partial \varepsilon}{\partial \lambda} + v \frac{\partial \varepsilon}{\partial \phi} + \omega \frac{\partial \varepsilon}{\partial \xi} \tag{A6}
\]

The primary variables are \( u, v, \omega, w, p, p', q, \theta, e, x, \) and \( p' \). Here, \( u \) and \( v \) are the horizontal velocity components, \( \omega \) the transformed vertical velocity component, \( w \) the vertical velocity component, \( p \) the total
pressure, $p'$ the deviation pressure, $q$ the specific humidity, $\theta$ the potential temperature, $e$ the turbulent kinetic energy, $X$ the particulate concentration, $p'$ the deviation density, and $T'$ the deviation temperature that is adjusted for moisture. The constants used in these equations include $g$ the gravitational force, $f$ and $f$ the Coriolis parameters, $Q$ the diabatic heating, $R$ the gas constant for dry air, $p_0$ the reference pressure, $C_v$ the specific heat capacity for dry air, $T_b$ the basic-state temperature, $p_b$ the basic state pressure, and $\rho_b$ the basic state density.

For numerical grids that incorporate terrain slopes less than 5° and have horizontal scales much larger than the vertical scales, the hydrostatic approximation, Eq. (A3), is sufficiently accurate.

Closure for the prognostic equations is based on $K_m$ theory. The vertical exchange coefficient for momentum $K_m$ is determined from mixing-length theory and the turbulence kinetic energy. The other vertical exchange coefficients, $K_x$, $K_y$, $K_z$, and $K_w$ are a function of $K_m$. The horizontal exchange coefficient $K_d$ is calculated from the deformation rate. In Eq. (A6), $\beta$ is a constant and 1 is a length scale. Additional details of the turbulence parameterizations and closure methods are described in Paege and McLawhorn (1983) and McCorcle (1988).

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