A numerical study on the interaction of nonclassical mesoscale circulations and baroclinic systems

Jerome Dean Fast

Iowa State University

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A numerical study on the interaction of nonclassical mesoscale circulations and baroclinic systems

Fast, Jerome Dean, Ph.D.
Iowa State University, 1990
A numerical study on the interaction of nonclassical mesoscale circulations and baroclinic systems

by

Jerome Dean Fast

A Dissertation Submitted to the Graduate Faculty in Partial Fulfillment of the Requirements for the Degree of DOCTOR OF PHILOSOPHY

Department: Geological and Atmospheric Sciences Major: Meteorology

Approved:

Signature was redacted for privacy.

Signature was redacted for privacy.

In Charge of Major Work

Signature was redacted for privacy.

For the Major Department

Signature was redacted for privacy.

For the Graduate College

Iowa State University
Ames, Iowa
1990
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GENERAL INTRODUCTION

Mesoscale circulations can be divided into two categories. The first category includes phenomena such as sea and land breezes, mountain and valley winds, and urban circulations which are forced by heterogeneous land characteristics. The second category includes circulations such as low-level jets, fronts, convection bands, and tropical cyclones that are primarily forced by instabilities in larger-scale synoptic systems. Those mesoscale systems in the second category have been difficult to resolve because many significant features fall between stations of the standard radiosonde network. As a result, numerical modeling has been employed to examine the detailed flow patterns predicted by the governing equations. A wide variety of numerical models have been reported in the literature to simulate the complex properties of mesoscale atmospheric phenomena.

History of the Model

The mesoscale model used in the present study has undergone many modifications to improve the numerical representation of the governing equations, the parameterization of the physics at the earth/atmosphere surface, and the dynamics of the flow above the boundary layer.

The two-dimensional form of the original boundary-layer model was developed by Paegle and McLawhorn (1983). The model was designed to predict the diurnal cycles of boundary-layer flows of synoptic horizontal scale above sloping terrain. It was found that the diurnal convergence cycle in the central United States was sensitive to soil parameters, absolute rotation parameterization, mixing parameterization, and longwave radiative flux divergence. Applications over complex North American terrain produced boundary-layer ascents that were generally in phase with observed summer diurnal thunderstorm distributions. The numerical model also was found to display parameter sensitivity similar to more idealized semi-analytic treatments (Paegle and McLawhorn, 1973; Paegle and Rasch, 1973; Paegle, 1978).

The model was extended to three dimensions by Astling et al. (1985) and was used to study the boundary-layer control of nocturnal convection. In this study, the model produced well-defined nocturnal convection that the operational LFM (Limited Fine Mesh) model did not predict. Sensitivity simulations suggested that the influences by topographically bound low-level circulations upon the short-term evolution of convection may be more predictable than larger scale influences.
Paegle et al. (1984) and Paegle (1984) used the model in a study of the low-level jet circulations near the Alps. They suggested that the observed strong nocturnal winds may have been a principal consequence of the enhanced nocturnal stratification that would tend to force more of the current to flow around the mountains at night than during the day. This study demonstrated that this mechanism may be important in the vicinity of smaller mountain ranges, such as the Alps, where inertial oscillations could not develop over such short horizontal distances. In an investigation of the model performance in the case of a low-level jet north of the Alps, Paegle and Vukicevic (1987a,b) found that the model was insensitive to initial data and determined that fixed forcings, such as topography, may enhance the predictability of these boundary-layer flows.

McCorcle (1988) added a soil moisture package to the model similar to the AFGL (Air Force Geophysics Laboratory) Soil Hydrology Model described by Mahrt et al. (1987) and examined the effects of surface moisture on the Great Plains low-level jet. The diurnal oscillations of the boundary-layer winds produced by the coupled earth-atmosphere model were very sensitive to surface moisture content and distribution.

Fast and McCorcle (1990) used the coupled earth-atmosphere model to determine the sensitivity of the Great Plains low-level jet to surface slope, latitude, soil type, and soil-moisture content and distribution. The magnitude and position of the simulated low-level jet was found to be very sensitive to surface slope, latitude, and soil-moisture content and distribution. The magnitude and position of the low-level jet was also sensitive to soil-type distribution only when moisture is incorporated in the soil layer.

It is known that some insect pests of corn are introduced each spring to the Midwest by the northward migration of populations that overwinter far to the south. Research by Kaster and Showers (1982) and Domino et al. (1983) has shown that nocturnal, long-range movement of noctuids, such as the black cutworm moth, is strongly correlated to particular low-level wind conditions that are common during the spring over the central United States. National Weather Service models lack the necessary vertical and horizontal resolution necessary to adequately forecast low-level wind and temperature fluctuations that are important in the prediction of insect migration. McCorcle and Fast (1990, 1989) and Showers et al. (1989, 1988) demonstrated that the boundary-layer model could be used to establish forecast criteria to accurately predict the movement of wind-transported pests. In order to predict the probable areas of infestation, a trajectory package and an advection-diffusion predictive equation were incorporated into the model. The boundary-layer model
was found to be superior to the National Weather Service models in forecasting the direction and distance of the insect migration.

Nonclassical Mesoscale Circulations

Horizontal gradients in soil moisture, soil type, vegetation, snow cover, or cloud cover may cause thermally-induced circulations similar to sea-breezes to develop. This type of circulation has been referred to as a nonclassical mesoscale circulation (NCMC) by Segal et al. (1989). The observation and numerical prediction of NCMCs 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.

Some studies indicate that normal variations of soil moisture have a more significant impact on boundary-layer characteristics than the more frequently observed changes in surface albedo, surface roughness, or soil thermal capacity. The presence of soil moisture and vegetation is expected to modify the surface thermal fluxes as compared to those of an equivalent bare soil surface under the same environmental conditions. Variations in surface moisture or vegetation can significantly alter the magnitudes of the sensible and latent heat fluxes to produce circulations almost as intense as a typical sea breeze.

One-dimensional numerical studies have shown that the presence of soil moisture and vegetation can alter the components of the surface heat budget, which ultimately affected the circulations in the boundary layer (Deardorff, 1977; Deardorff, 1978; Lin and Sun, 1986; Marht and Pan, 1984; Marht et al., 1987; McCumber and Pielke, 1981; Noilhan and Planton, 1989; Pan and Marht, 1987). The models in these studies vary in complexity, and it is not clear yet how detailed a model needs to be to adequately simulate the energy and moisture exchanges at the soil-atmosphere interface. Some of these studies compared the results of the soil-layer parameterizations with observations. Most of these models qualitatively reproduced the diurnal variation in temperature and changes in soil moisture that were observed.

There have been relatively fewer studies that incorporate soil moisture or vegetation parameterizations in two and three-dimensional mesoscale or boundary-layer models. This is the result of a lack of data on the horizontal and temporal scale needed to verify these numerical models. Subgrid-scale variability is particularly important in modeling evapotranspiration because large soil moisture gradients may occur on scales as small as a few meters (Avissar and Pielke, 1989). In the future, remote sensing by satellites may be
able to routinely determine high resolution horizontal soil moisture distributions (Wetzel and Chang, 1988). At the present time, numerical simulations of NCMCs in two and three-dimensional models have been forced to use simple, theoretical soil-moisture and vegetation distributions.

Ookouchi et al. (1984) evaluated the thermally-induced circulation over flat terrain due to nonuniform horizontal distribution of soil-moisture availability using a two-dimensional numerical model with no synoptic flow. The effect of soil-moisture distribution on the thermally-induced upslope flow also was examined.

Soil and vegetation parameterizations were incorporated into a two-dimensional mesoscale model that included a detailed representation of the boundary layer in Mahfouf et al. (1987). In this study, the atmospheric response to soil and vegetation inhomogeneities was examined over flat terrain with no synoptic flow.

Segal et al. (1988) evaluated the effect of vegetated surfaces on sea-breeze and daytime thermally-induced upslope flows. Simulations incorporating a simplified vegetation parameterization were compared with simulations that used a more complicated representation of vegetation. The generation of thermally-induced flow by vegetated areas contrasted with bare-soil areas was also examined. The two-dimensional numerical model contained no synoptic flow, except for one simulation.

Yan and Anthes (1988) used a two-dimensional numerical model to simulate circulations induced by horizontal variations in surface moisture availability. The model contained prognostic equations for water vapor, cloud water, and rain water, with a simple parameterization of cloud microphysical processes. Four geometric variations of surface moisture were examined: 1) an edge geometry which includes a land-water contrast and moist land adjacent to dry land, 2) a single strip of moist land surrounded by dry land, 3) alternating bands of moist and dry land, 4) a single strip of dry land surrounded by moist land. No synoptic forcing was considered.

Pinty et al. (1989) employed soil and vegetation parameterizations in a two-dimensional atmospheric model. In this investigation, significant circulations developed in response to a vegetation discontinuity when sufficient moisture was present in the soil and no synoptic forcing was imposed.

These two-dimensional studies indicate that soil moisture and vegetation parameterizations were found to have a significant effect on the structure of the simulated boundary layer. The resulting thermally-induced circulations were as intense as typical sea-
breezes when these models employed horizontal grid spacing between 5 and 20 km and imposed no synoptic flow.

Nonclassical mesoscale circulations could also have a significant effect on larger circulations, such as the Great Plains low-level jet, with an imposed background synoptic flow field. McCrindle (1988) examined the effect of evaporation of soil moisture on the Great Plains nocturnal jet by coupling a soil-hydrology system in a three-dimensional boundary-layer numerical model. The response of the low-level jet to evaporation from the soil was examined by 1) saturating the soil over the Rockies and by 2) saturating the soil over the eastern Great Plains. All of the simulations included a significant imposed synoptic flow field. Fast and McCrindle (1990) used typical springtime synoptic conditions to initialize a two-dimensional numerical model that examined the effect of soil type, soil-moisture content, and soil-moisture distribution on the Great Plains low-level jet.

A mesoscale model in one, two, and three-dimensions was used in Tjernstrom (1989) to examine the effect of soil moisture and vegetation on boundary-layer circulations. This study was primarily concerned with the effects of discontinuities of forest and crop areas in complex terrain with a very simple, uniform synoptic flow field.

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). Southwesterly flow over the plateau advected the warm, dry mixed layer northward over the cooler, moister air from the Gulf of Mexico. 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 pre-storm environment.

Climate models are also being tested with soil-layer and vegetation parameterizations in an effort to more realistically describe the surface energy budget (Dickinson, 1984; Dickinson et al., 1986; Meehl and Washington, 1988; Sellers and Dorman, 1987; Wilson et al., 1987). These studies primarily illustrate one-dimensional model parameterizations of the soil-layer physics that are intended for assimilation into three-dimensional climate models. Global climate models have predicted that increased carbon dioxide in the atmosphere may cause higher average temperatures and lower average precipitation in the central United States. Global climate modeling studies have also shown considerable sensitivity of their simulated climatologies to drastic changes in the formulation of soil evaporation and evapotranspiration; therefore, forecasts of mean temperature and precipitation patterns may be altered.
It is apparent that more simultaneous observations of meteorological and biophysical parameters are needed to understand the complex relationships at the earth/atmosphere interface and to verify the parameterizations used by two and three-dimensional models. This was the motivation for the recent HAPEX-MOBILHY (Andre et al., 1988) and FIFE (Sellers et al., 1988) experiments. HAPEX-MOBILHY was a detailed three-dimensional study of the boundary layer for a $10^4$ km$^2$ region in southwestern France that included surface and satellite observations of soil-moisture and vegetative characteristics as described by Andre et al. (1986). Some of the initial data from this experiment can be found in Andre et al. (1988) and Pinty et al. (1989). The meteorological and biophysical parameters at a 225 km$^2$ site in northern Kansas were observed using satellite and surface based instrumentation in FIFE (Sellers et al., 1988). Segal et al. (1989) presented a limited set of observations taken over irrigated areas in northeast Colorado along with a few three-dimensional simulations of the flow field. They concluded from their observations and numerical simulations that the effects of NCMCs were evident up to at least 500 m above the terrain. Numerical simulations also indicated that synoptic flow tended to mask the effects of NCMCs.

All of these numerical studies indicate that horizontal discontinuities in soil moisture or vegetation could induce significant discontinuities in surface thermal forcing and, consequently, mesoscale circulations. Such circulations may play an important role in patterns related to local meteorology and climatology, cumulus convection, and air quality. Since it is increasingly being recognized that atmospheric processes are inherently connected to energy exchanges at the ocean and earth surface, realistically coupling the earth and atmosphere in numerical models will be an important topic for many years.

Explanation of Dissertation Format

The development of the anelastic, hydrostatic mesoscale model numerical study and the numerical study of baroclinic circulations in the central United States was written in the alternate dissertation format. Two papers have been written that will be submitted to professional atmospheric science journals. The first paper employs the mesoscale model to simulate the effects of horizontally inhomogeneous land characteristics on a frontal passage in the central United States. In the second paper, the mesoscale model is used to simulate the effect of baroclinicity and inhomogeneous land characteristics on the transport of insect pests. A summary and discussion of the two studies is presented following the two papers.
PAPER 1.
THE EFFECT OF HETEROGENEOUS SOIL MOISTURE
ON A SUMMER BAROCLINIC CIRCULATION IN THE
CENTRAL UNITED STATES
THE EFFECT OF HETEROGENEOUS SOIL MOISTURE
ON A SUMMER BAROCLINIC CIRCULATION IN THE
CENTRAL UNITED STATES

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(to be submitted to Monthly Weather Review)
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. 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.
I. 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 over-predicted in many situations. Large horizontal gradients in land characteristics assumed by these studies are resolved by a grid spacing of 5 to 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 by Segal et al. (1989); however, it is possible that NCMCs interact with larger-scale circulations under the proper circumstances. 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 prestorm environment.

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.

While most studies of NCMCs assume homogeneous land characteristics within a grid element 10 to 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. Avissar 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 has 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. Avissar 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/atmosphere interface and to verify two and three-dimensional models. This was the motivation for the recent FIFE (Sellers et al., 1988) and HAPEX-MOBILHY (Andre et al., 1986) experiments. Some of the initial data from
HAPEX-MOBILHY 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 II outlines the development of the present mesoscale model used to study NCMSs 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 June 21 - 23, 1989 and is described in Section III. 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 IV. 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 V presents the conclusions of this study.
II. 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, \( \sigma \), is defined by

\[
\sigma = s \left[ \frac{z - z_\alpha}{s - z_\alpha} \right] \quad (1)
\]

where \( z \) is the cartesian vertical coordinate, \( s \) is the constant height of the model top, and \( z_\alpha \) 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

\[
\xi = \alpha \ln \left[ \frac{\sigma}{z_\alpha} \right] \quad (2)
\]

where \( \alpha \) is a constant and \( z_\alpha \) 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
\[
\begin{align*}
\frac{Du}{Dt} &= \frac{\partial u}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial u}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial u}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial u}{\partial \xi} \frac{\alpha}{\sigma} \right) + f_v \\
- f^2 \left[ \frac{1}{\alpha \cos \phi} \frac{\partial z_o}{\partial \lambda} \left( \frac{s - \sigma}{s} \right) + u + \frac{1}{a \delta \phi} \left( \frac{s - \sigma}{s} \right) v + \left( \frac{s - z_o}{s} \right) \right] - \frac{1}{\rho a \cos \phi} \frac{\partial p}{\partial \lambda} \\
+ g \frac{1}{\alpha \cos \phi} \frac{\partial z_o}{\partial \lambda} \left( \frac{\sigma - s}{s} \right) + \nabla \cdot (K_d \nabla u) \\

\frac{Dv}{Dt} &= \frac{\partial v}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial v}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial v}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial v}{\partial \xi} \frac{\alpha}{\sigma} \right) - fu - \frac{1}{\rho a \delta \phi} \\
+ g \frac{1}{a \delta \phi} \left( \frac{\sigma - s}{s} \right) + \nabla \cdot (K_d \nabla v) \\

\left( \frac{s}{s - z_o} \right) \frac{\partial p'}{\partial \xi} = -p' g \\

\frac{Dq}{Dt} &= \frac{\partial q}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial q}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial q}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial q}{\partial \xi} \frac{\alpha}{\sigma} \right) + \nabla \cdot (K_d \nabla q) \\

\frac{D\theta}{Dt} &= \frac{\partial \theta}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial \theta}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial \theta}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial \theta}{\partial \xi} \frac{\alpha}{\sigma} + Q \left( \frac{P_c}{C_r} \right)^{k/c} \right) \\
+ \nabla \cdot (K_d \nabla \theta) \\

\frac{Dx}{Dt} &= \frac{\partial x}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial x}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial x}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial x}{\partial \xi} \frac{\alpha}{\sigma} \right) \\
- \frac{8}{K_d} \frac{\partial}{\partial \xi} \left( \frac{s}{s - z_o} \right) + \nabla \cdot (K_d \nabla e) \\

\frac{Dy}{Dt} &= \frac{\partial y}{\partial t} + u \frac{1}{\alpha \cos \phi} \frac{\partial y}{\partial \lambda} + v \frac{1}{a \delta \phi} + \omega \frac{\partial y}{\partial \xi} \frac{\alpha}{\sigma} = \left( \frac{s}{s - z_o} \right)^2 \frac{\partial}{\partial \xi} \left( K_m \frac{\partial y}{\partial \xi} \frac{\alpha}{\sigma} \right) + S_y + \nabla \cdot (K_d \nabla y) \\

p \left[ \frac{1}{\alpha \cos \phi} \frac{\partial x}{\partial \lambda} + \frac{1}{\delta \phi} \frac{\partial (\omega \cos \phi)}{\partial \phi} \right] + \frac{\partial (\omega p)}{\partial \xi} \frac{\alpha}{\sigma} = \rho_s \left[ \frac{u}{\cos \phi} \frac{\partial z_o}{\partial \lambda} + \frac{\nu}{\delta \phi} \frac{\partial z_o}{\partial \phi} \right] = 0
\end{align*}
\]
\[ \frac{\omega}{\cos \phi} \partial \frac{\partial z_0}{\partial \lambda} \left( \frac{\sigma - s}{s - z_0} \right) + \frac{\cos \phi}{\partial \phi} \left( \frac{\sigma - s}{s - z_0} \right) + \frac{w}{s - z_0} s = (11) \]

\[ \rho' = \frac{p' - \rho_{0}RT'}{RT'} (12) \]

\[ T' = \theta \left( \frac{p}{p_{0}} \right)^{\frac{B}{C}} - T_{b} + 0.61qT_{b} (13) \]

\[ p = p_{0} + p' (14) \]

The primary variables are \( u, v, w, p, p', q, \theta, \chi, T', \) and \( \rho' \). 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, \( \chi \) the particulate concentration, \( \rho' \) the deviation density, and \( T' \) the deviation temperature that is adjusted for moisture. The constants used in these equations include \( g \) the gravitational force, \( f \) the Coriolis parameter, \( Q \) the diabatic heating, \( R \) the gas constant for dry air, \( p_{0} \) the reference pressure, the specific heat capacity for dry air, \( T_{b} \) the basic state temperature, \( p_{0} \) the basic state pressure, and \( \rho_{0} \), 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. (5), is sufficiently accurate.

Closure for the prognostic equations is based on K-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, \( K_{q}, K_{h}, K_{n}, \) and \( K_{e} \) are a function of \( K_{m} \). The horizontal exchange coefficient \( K_{e} \) is calculated from the deformation rate. In Eq. (8), \( \beta \) is a constant and \( l \) is a length scale. Additional details of the turbulence parameterizations and closure methods are described in Paegle and McLawhorn (1983) and McCorcle (1988).

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. Modest changes in soil moisture can alter the thermal structure of the lower boundary layer and ultimately the entire dynamic field (Mahfouf et al., 1987; Ookouchi et al., 1984).

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. 10) from the roughness height to the model top. The vertical velocity, $w$, is then determined from Eq. (11). The hydrostatic equation, Eq. (5) is integrated from the model top to the surface to determine the deviation pressure, $p'$.

The present numerical model does not contain a parameterization for cumulus convection. Evaporated soil moisture is simply advected by the wind field and there are no feedback processes that could reduce this atmospheric moisture.

### 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, $z_o$, to obtain

$$G - F_n + \rho C_v K_s \frac{\partial \theta}{\partial z} - \lambda \frac{\partial T_{soil}}{\partial z} - \rho \lambda L E = 0$$

where $F_n$ is the longwave radiative flux, $\lambda$, the soil thermal conductivity, $\rho$ the density of water, $L$, the latent heat of vaporization, and $E$ the evaporation rate in m s$^{-1}$. The longwave (terrestrial) radiative flux, $F_n$, and the atmospheric flux divergence, $Q$, that appears in Eq. (7) are computed as functions of the water-vapor path length integrated through the atmosphere (Paegle and McLawhorn, 1983). Equation (15) 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

$$G = 1353 \, \text{Wm}^{-2} \times (1 - A) \times (\sin \phi \cos \delta + \cos \phi \sin \pi t/12) \tau$$

(16)
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.

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:

\[ A = 0.31 - 0.34\eta/\eta_s \quad \eta/\eta_s \leq 0.5 \]
\[ A = 0.14 \quad \eta/\eta_s > 0.5 \] (17)

Equation (17) is used in several sensitivity simulations to examine its effect on boundary-layer circulations.

The transmittance in Eq. (16) is a function of cloud cover and is parameterized following Anthes et al. (1987). Cloudy skies could reduce daytime temperatures at the surface so that evaporation from the soil may be reduced or eliminated, and the potential effects of NCMCs would be diminished. To more clearly isolate NCMCs, clear-sky conditions are assumed in the simulations described in this paper; therefore, $\tau$ is set to 1.0.

Temperature continuity is assumed at the roughness height such that

\[ T_{\text{air}} = T_{\text{soil}} \quad \text{at} \ z = z_o. \] (18)

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

\[ W_{\text{air}} = W_{\text{soil}} \quad \text{at} \ z = z_o \] (19)
\[ W_{\text{air}} = \rho K_s \frac{\partial q}{\partial z} \] (20)
\[ W_{\text{soil}} = \rho_e E \] (21)

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.

Specification of lateral boundary conditions has always been a problem of limited-area numerical models. Mesoscale circulations in the region of interest can be adversely influenced by improper specification of lateral boundary conditions. The best solution is to move the lateral boundaries as far from the region of interest as possible; however, this usually increases the number of grid points. Dirichlet, Neumann, or radiation lateral boundary conditions may be chosen in the current mesoscale model, depending upon the particular application.
The boundary-layer model described by McCorcle (1988) assumed Neumann type 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 which neglect synoptic flow.

Dirichlet lateral boundary conditions can be used to hold an initial value constant through a simulation. Dirichlet lateral boundary conditions can also specify time-dependent values obtained from larger-scale numerical models or objectively analyzed observation fields. 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 smoothing near the lateral boundaries; however, a simple low-pass filter is used near the lateral boundaries (Pielke, 1984) in the present model. Time varying lateral boundary conditions based on observed values are used in the baroclinic numerical simulations in this study.

The radiation boundary condition formulation used in the model is based on a formulation similar to Orlanski (1976). The boundary value of a prognostic variable, $\Phi$, is determined from

$$\frac{\partial \Phi}{\partial t} = -c \frac{\partial \Phi}{\partial n}$$

where \(n\) is direction normal to a lateral boundary. The phase speed, \(c\), may be set to either a fixed value or determined from Eq. (22) using values of \(\Phi\) from the previous time step.

Dirichlet 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, $T_o$, 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.

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 using a single-pass Barnes scheme. At the model top, pressure is determined by the hydrostatic relationship from the observed 300 mb height field. The interior pressure is obtained by the integration of the hydrostatic equation, Eq. (5). The initial winds are geostrophic, except below 324 m, where the winds are forced to logarithmically approach zero at the roughness height.

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 time necessary for a single simulation. This technique requires an additional term added to the prognostic equations (Eqs. (3), (4), (6), and (7)) 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. The largest adjustment in the prognostic variables occurred near the surface where there was a greater imbalance in the wind field than the temperature or specific humidity field during the preforecast integration. Since pressure is dependent on the temperature, it also is adjusted to the thermal field during this period. After the preforecast period, the results beyond the
6 h forecast were very similar to the results from a simulation using geostrophic winds and no preforecast adjustment period. 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 (Sherretz 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 RAOB data for arbitrary horizontal grids for every standard observation level. 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. To examine the sensitivity of the model to horizontal scales of motion, several simulations employed a smaller domain and grid spacing. The model domain for both grids employs 25 nodes in both horizontal directions. A horizontal grid spacing of 104 km and a time step of 360 s is used for the larger domain. For the smaller domain, a horizontal grid spacing of 74 km and a time step of 180 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. 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 m 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 to 0.96 m below the surface. Because temperature 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.
III. 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 12 UTC June 21, 1989. The lateral and top boundary conditions incorporate synoptic data from 12 UTC June 21 to 12 UTC June 23, 1989. Numerical results will focus on the first 24 h period; therefore, only the details of the synoptic situation during this period are described here.

Surface fronts, pressure, temperature, and winds for 12 UTC June 21 and 22 are shown in Fig. 2a-b. At 12 UTC June 21, 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 00 UTC, daytime heating produced temperatures in excess of 30° C from the desert southwest to southeastern Missouri ahead of the front (not shown). After 24 h at 12 UTC June 22, 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 during the next 24 h (not shown).

The evolution of the specific humidity fields at the surface for this frontal passage is shown in Fig. 3a-b. At 12 UTC June 21, a strong moisture gradient existed just behind the cold front from North Dakota to northern Texas. At 00 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 00 UTC and 12 UTC June 22, 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 Fig. 4a-b, reveal a 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 12 UTC June 21 to 60 m s\(^{-1}\) on 12 UTC June 22. 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 June 22, 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.

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.
IV. NUMERICAL RESULTS

This section summarizes some of the simulations that are performed to isolate the mechanisms by which NCMCs could affect 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 III for 12 UTC June 21, 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, 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 June 21. 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 74 km and 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 Fig. 7a-b. Soil-moisture distribution SM1 is representative of the relatively wet and dry regions from the June 24 Crop Moisture Index (Fig. 6) and incorporates a gradual horizontal gradient in soil-moisture content. A uniform horizontal initial soil moisture region of \( \eta = 0.284 \) (about 65% the saturation value for loam soil) is used for soil-moisture distribution SM2. 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 data set of soil type for the entire United States. Most of the simulations in this paper use loam soil throughout the domain (distribution ST1) since the principal 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. Neumann lateral boundary conditions were used as described in Section II.

1. Dry-soil simulations

Figures 9a-b depict the wind, temperature, and specific humidity fields for the 12-h forecast valid for 1800 LST June 21, 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 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 rotated clockwise 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 June 22, 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 affected circulations with a horizontal scale of 140 km in the boundary-layer model of McCorcle (1988) and in global climate models (Dickinson, 1984; 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 Fig. 10a-f.

The most significant changes in the forecast variables occurred for the sharp moisture gradient of distribution SM2 as seen in Fig. 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 reduction 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 Fig. 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 properites of soils varies 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 was highly dependent on soil type as shown in Fast and McCorcle (1990).

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 Fig. 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. (17) 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. (16) did not differ significantly from experiments NS2 - NS4. The forecasted variables produced difference fields similar to those shown in Fig 10a-f.

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

The smaller grid experiments for soil-moisture distribution SM2 produced the same magnitude of NCMC as experiment NS3 as seen in Fig 12a-b. It appeared that the model was not particularly sensitive to the horizontal scales used here, although it is expected that the NCMCs would have a larger magnitude if the horizontal grid spacing was reduced further.

These simulations with no synoptic flow indicate the general structure and magnitude of the NCMCs 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 12 UTC June 21, 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 Neumann, Dirichlet, and radiation lateral boundary conditions. Results indicated that Neumann conditions quickly affected the results in the model interior and contaminated all of the features of the observed front. Both Dirichlet and radiation lateral boundary conditions, as described in Section II, retained the observed frontal features during this case. The simulated position of the front was slightly superior when Dirichlet conditions were used; therefore, Dirichlet lateral boundary conditions were employed for all of the baroclinic circulations reported in this section. Dirichlet lateral boundary conditions are used for all the prognostic variables, except for potential temperature which used Neumann 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 13a-f and 14a-f depict the wind, temperature, specific humidity, and moisture-convergence fields for the 12- and 24-h forecasts 2 m above the ground. The results portrayed in Figs. 14a-f are for the same synoptic situation, except that the smaller grid in Fig. 1 was employed. The front in simulation S1 has moved to northern Minnesota, central Iowa, and on into central Oklahoma and northern Texas by 1800 LST June 21, as seen in Fig. 13a. A warm pocket of air in excess of 30° C stretched from southern Texas to southern Kansas ahead of the front. The coldest air was located well behind the front in Wyoming. A sharp moisture gradient was evident near the frontal boundary in the central United States in Fig 13c. The large gradients at the southeast and northern boundaries resulted from the lateral boundary conditions employed by the mesoscale model. At this time, the model predicted significantly lower humidities near the boundaries than the observed values.

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

\[-q \nabla \cdot \nabla - \nabla \cdot \nabla q = - \nabla \cdot (q \nabla)\]  

(23)

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 Waldsteicher (1989), convection often develops downwind of a MFC maxima, where MFC increases rapidly in time, and where the gradient of MFC is increasing.

Figure 13e 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. (23).

During the following 12 h the simulated front weakened considerably and moved slightly southeastward as seen in Fig 13b. 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. 13d). In Fig. 13f, the convergence of moisture has weakened significantly along the southern portions of the front.

The results of the smaller grid simulation (S11) are similar to those of the larger grid, except that the front pushed 200 km further east into central Wisconsin and western Illinois as seen in Fig. 14a-b. The smaller grid simulation predicted the surface temperature to be about 2° C warmer throughout the domain, so that the 30° C isotherm extended into southern Iowa. The specific humidity field was very similar to simulation S1, except that the air mass behind the front was significantly moister throughout the period. The MFC patterns in Fig. 14e-f are quite different from those in Fig. 13e-f. The moisture convergence was evident all along the front from Oklahoma to northern Minnesota on 1800 LST June 21 in simulation S11. The moisture convergence diminished during the evening along the front as in simulation S1. A discrepancy in MFC fields between grid sizes arises not only because the wind fields are different, but also because MFC is a mathematical derivative, and is dependent on the spatial scale (Waldstreicher, 1989).

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. 15a). 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. 15c) 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 15d.

The results of the soil-moisture distribution (SM2) in Fig. 16a-d are similar to those in Fig. 15a-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 16a and 1.2 m s\(^{-1}\) after 24 h in Fig. 16b. 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 June 21 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. 16c-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 Fig. 17a-d. As with the corresponding simulation 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. 17c-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. 15 to 17 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. This 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 which intensifies the nonclassical circulation. This latter mechanics was 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 S2 (sharp moisture gradient) as seen in Figs. 18a-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. (23) is now due to the moisture advection 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 Fig. 19a-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 Fig. 20a-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. 20c.

The vertical cross-section plots in Figs. 20e-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. (17) 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 Fig 10a-f.

The time evolution of the potential-temperature and specific-humidity profiles for simulations S1 and S3 are shown in Figs. 21a-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 Fig. 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. 22a-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.

The difference fields for the small-grid simulations in Fig. 23a-f once again show a similar pattern and magnitude. It appeared that the grid spacing did not significantly affect the magnitude of the reduction in temperature in the boundary layer or the change in the wind vector. Substantial amounts of moisture were advected northward ahead of the front into eastern Iowa. The MFC difference fields are more detailed than the corresponding large-grid simulations (S1 and S3). A maximum moisture-convergence region existed in southern Iowa after 12 h as seen in Fig. 23a. After 24 h, significant convergence still existed all along the front with three enhanced areas in northern Minnesota, eastern Missouri and western Illinois, and western Oklahoma. These results predict much more favorable conditions for precipitation than the larger-grid simulations.

Even though the larger-grid simulations were able to produce a similar NCMC, it is not obvious from these simulations what resolution is superior when determining MFC.
V. CONCLUSIONS

Atmospheric processes are inherently connected to energy exchanges at the ocean and earth surface. The observation and numerical prediction of NCMCs 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 a bare-soil surface under the same environment 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 to 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 McCrclle (1988) 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 which were set equal to their geostrophic value.

Observations from 12 UTC June 21, 1989 of a frontal passage were used to initialize the three-dimensional 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 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 to 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 were found to have an even greater effect than in the no-synoptic-flow 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 cumulus and precipitation 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 over-predicted.

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 observations soil moisture 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.

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.
VI. ACKNOWLEDGMENTS

This research was supported by the Iowa State University Agricultural and Home Economics Experiment Station under project 2804. The data and some of the software for this study were made available through the Unidata program, which is sponsored by the National Science Foundation.
VII. REFERENCES


VIII. APPENDIX A: DEVELOPMENT OF THE GOVERNING EQUATIONS

A. Transformation to the Nonorthogonal Coordinate System

Spherical coordinates are used in atmospheric models to account for the earth's curvature when simulating relatively large-scale synoptic or mesoscale phenomena. Also, the meteorological data used to initialize a model often are available on a grid system defined in terms of latitude and longitude. The present model incorporates spherical coordinates for the governing equations. The horizontal derivatives in this system are

$$\frac{\partial}{\partial x} = \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda}, \quad \frac{\partial}{\partial y} = \frac{1}{a} \frac{\partial}{\partial \phi}$$

where $a$ is the earth's radius, $\lambda$ is the longitude, and $\phi$ is the latitude. For most atmospheric models, it is customary to transform the vertical coordinate, $\sigma$. The functional forms of $\sigma$ normally used by atmospheric models are the isentropic, isobaric, or sigma representations. In the present model, the cartesian coordinate system $(x, y, z)$ used in McCorcle (1988) has been transformed into a sigma, nonorthogonal terrain-following coordinate system $(x, y, \sigma)$. The relation between the spatial coordinates for the nonorthogonal system and the cartesian system is given by

<table>
<thead>
<tr>
<th>Nonorthogonal:</th>
<th>Cartesian:</th>
</tr>
</thead>
<tbody>
<tr>
<td>$x = x$</td>
<td>$x = x$</td>
</tr>
<tr>
<td>$y = y$</td>
<td>$y = y$</td>
</tr>
<tr>
<td>$\sigma = \frac{z - z_0}{s - z_0}$</td>
<td>$z = \sigma \left( s - z_0(x, y) \right) + z_0(x, y)$</td>
</tr>
</tbody>
</table>

In order to preserve the invariance of the physical representation of the governing equations from one coordinate system to another, the tensor analysis method was used to transform the prognostic equations and the diagnostic equation for continuity (Pielke, 1984).

The individual velocity components in the nonorthogonal coordinate system defined in Eq. (25) can be expressed in terms of the velocity components in the cartesian system as
Strong vertical gradients in the prognostic variables commonly occur in the boundary layer, especially near the surface. In order to adequately resolve the structure of many boundary layer phenomena, a fine vertical grid spacing is needed near the surface. The vertical coordinate, \( \sigma \), is transformed into a new vertical coordinate, \( \xi \) (Eq. (2)), so that the nodes near the surface are logarithmically spaced (Paegle and McLawhorn, 1983). Then vertical derivatives for this system are

\[
\begin{align*}
\frac{\partial u}{\partial \sigma} &= \frac{\partial u}{\partial \xi} \frac{\partial \xi}{\partial \sigma} = \alpha \frac{\partial u}{\partial \xi}, \\
\frac{\partial^2 u}{\partial \sigma^2} &= \frac{\partial^2 u}{\partial \xi^2} \left( \frac{\partial \xi}{\partial \sigma} \right)^2 + \frac{\partial u}{\partial \xi} \frac{\partial^2 \xi}{\partial \sigma^2} = \left( \frac{\alpha}{\sigma} \right)^2 \frac{\partial^2 u}{\partial \xi^2} - \frac{\alpha}{\sigma} \frac{\partial^2 u}{\partial \xi^2} \frac{\partial \xi}{\partial \sigma} 
\end{align*}
\]

Above some predetermined level, the nodes are evenly spaced above the logarithmic layer to the top of the model.

### B. Closure

Equations (3) - (14) contain the variables \( u, v, \omega, w, p, p', q, \theta, e, \chi, p', T', K_\omega, K_e, K_q, K_r, K_a, \) and \( K_d \) that are unknown. There are twelve equations and eighteen unknowns; therefore, the system is unsolvable mathematically and it must be closed by parameterizing the exchange coefficients in some manner.

For the present model, \( K_a \) is given by the following relation:

\[
K_a = (0.2 \pi)^{1/3} l
\]

where \( l \) is the mixing length. The vertical diffusion for heat is assumed to be equal to the vertical eddy diffusion for moisture and is
The vertical diffusion of kinetic energy is given by

\[ K_v = 1.35 K_m \]  

(29)

Finally, the horizontal diffusion coefficient is calculated as a function of the deformation rate, \( D_n \), as in Smagorinsky (1963),

\[ K_x = 0.36 D_n \Delta^2 \]  

(31)

where \( \Delta \) is the horizontal grid spacing.

A reference length scale, \( l_o \), based on a vertically integrated turbulence kinetic energy profile is calculated as in Yamada and Mellor (1975).

\[ l_o = 0.1414 \int \frac{e z dz}{e dz} \]  

(32)

Then the length scale, \( l \), is determined using similarity functions similar to those in Zdunkowski et al. (1976) such that

\[ l = \frac{1}{\frac{1}{l_o} + \frac{\phi(z/L)}{kz}} \quad \text{and} \quad L = \frac{u^2}{\frac{k}{\theta_{efc}} K_v \frac{\partial \theta}{\partial z_{efc}}} \]  

(33)

where \( k \) is von Karman's constant, \( L \) is the Monin-Obukov length, \( u^2 \) is the friction velocity, and \( \phi(z/L) \) is a similarity function determined from

- If \( z/L > 0 \) (stable) \( \phi(z/L) = 5.1 \)
- If \( 0 < z/L < 0.82 \) (stable) \( \phi(z/L) = 1 + 5\phi(z/L) \)
- If \( z/L < -10 \) (neutral) \( \phi(z/L) = \left( \frac{1}{161} \right)^{1/3} \)
If \(-10 < z/L < 0\) (unstable) \[
\phi(z/L) = \left(\frac{1}{1 - 16\phi(z/L)}\right)^{1/3}
\] (34)

These similarity functions are described in Zdunkowski et al. (1976).

C. Treatment of Perturbation Temperature and Pressure

In the present mesoscale model, pressure, density, and temperature are composed of basic state thermodynamic variables (subscript \(s\)) and deviation thermodynamic variables (primed) (Paegle and McLawhorn, 1983). The basic state pressure, density, and temperature fields are assumed to vary only in height and are not a function of atmospheric moisture.

Equation (5) is obtained by substituting \(p = p_s + p'\) and \(\rho = \rho_s + \rho'\) into the hydrostatic equation. The total pressure is used in the horizontal momentum equations, Eqs. (3) and (4), and is updated each time step after integrating Eq. (5). The equation of state can be written as

\[
p_s - p' = (\rho_s + \rho')R(T_s - T')
\]

\[
p_s + p' = \rho_sRT_s + \rho_sRT' + \rho'RT_s + \rho'RT'
\]

\[
p' = \rho_sRT' + \rho'RT_s + \rho'RT'
\] (35)

Virtual temperature, \(T_v\), is defined as \(T_v = T_s + 0.61q\). The mesoscale model of Paegle and McLawhorn (1983) adjusts the deviation temperature assuming that the temperature, \(T\), can be replaced by the virtual temperature in the following equation such that:

\[
T' = T_v - T_s
\]

Then the deviation temperature can be written as:

\[
T' = T + 0.61qT_s - T_s
\]

\[
T' = T + 0.61q(T_s + T') - T_s
\]

\[
T' = T + 0.61qT_s - T_s
\]
where the Poisson equation has been used for the dry-air temperature. When Eq. (36) is used in Eq. (35), one obtains the expression for deviation pressure presented in Paegle and McLawhorn (1983).

D. Governing Equations for the Soil

The soil portion of the mesoscale model employs forecasts of soil moisture and soil temperature similar to a method described by Mahrt and Pan (1984) and Pan and Mahrt (1987). The details of the soil hydrology in this model can also be found in McCorcle (1988) and Fast and McCorcle (1990). The soil system includes a transport equation for the volumetric soil-moisture content, \( \eta \), given by

\[
\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z} \left[ D_s \frac{\partial \eta}{\partial z} \right] + \frac{\partial K_s}{\partial z} - E,
\]

and a soil heat flux equation:

\[
C \frac{\partial T_{\text{soil}}}{\partial t} = \frac{\partial}{\partial z} \left[ \lambda_s \frac{\partial T_{\text{soil}}}{\partial z} \right],
\]

where \( D_s \) is the soil diffusivity, \( K_s \) the hydraulic conductivity, \( E \) the change in moisture due to evapotranspiration or dewfall, \( C \) the volumetric heat capacity, \( \lambda_s \) the soil thermal conductivity, and \( T_{\text{soil}} \) is the soil temperature. \( D_s \) is the soil diffusivity and \( K_s \) is the hydraulic conductivity, defined by:

\[
D_s = \frac{b K_w W_s}{\eta_s} \left( \frac{\eta}{\eta_s} \right)^{b+3} \tag{39}
\]

\[
K_s = K_w \left( \frac{\eta}{\eta_s} \right)^{2b+3} \tag{40}
\]
where $K_w$ and $\eta_w$ are saturation values of hydraulic conductivity and soil-moisture content, respectively. $W_s$ is the saturation moisture potential, and $b$ is a coefficient dependent on soil type. All these parameters are empirically defined as in Clapp and Hornberger (1978) and are soil-type dependent. For the top soil level, precipitation may be added to Eq. (37) but is not included in this research. The soil thermal conductivity is given by:

$$\lambda_s = \exp\left(-(\log_{10} \Psi + 2.7)\right) \quad \text{for } \log_{10} \Psi < 5$$

$$\lambda_s = 0.00041 \quad \text{for } \log_{10} \Psi > 5.1 ,$$

according to Al Nakshabandi and Kohnke (1965), where $\Psi$ is the soil-moisture potential. The parameters $D_v$, $K_s$, $C$, $W_s$, and $\lambda_s$ are strong functions of soil type and soil-moisture content. Equation (42) implies that the soil thermal conductivity is the same for all soil types if they are dry. This is not entirely true (Hillel, 1980), but it is a good first-order approximation.

The volumetric heat capacity of the soil is estimated as a weighted function of the moisture content such that

$$C = (1 - \eta_s)C_i + \eta_sC_w$$

In this relation, $C_i$ is the heat capacity for the dry soil, and $C_w$ is the heat capacity of water.

For the ocean, the thermal conductivity is defined as

$$\lambda_ocean = \rho_o C_w K_w ,$$

where $\rho_o$ is the density of water, and $K_w$ the vertical-eddy-mixing coefficient, is chosen as $1.3 \times 10^4$ m$^2$ s$^{-1}$ (McCormick, 1988). This large value assures that sea surface temperatures are almost constant during the forecast period.

Evaporation in the model is calculated as a flux of moisture from a bare soil surface. The model can calculate evaporation and transpiration from a canopy, but they are not included in the present simulations. Evaporation is expressed as a function of potential evaporation, $E_p$, which is defined as the evaporation from a well-watered surface under a given set of atmospheric conditions. Potential evaporation in the model is similar to a Penman (1948) formulation that includes the effects of solar radiation, longwave radiation, soil heat flux, wind speed, and atmospheric stability.
The direct evaporation from the soil is assumed to occur at the potential rate if the soil moisture is above an "air-dry" value, $\eta_a$ (Nimah and Hanks, 1973).

\[
\text{If } \eta_s > \eta_{ec} > \eta_a \text{, then } E_{\text{soil}} = E_p \text{ at } z = z_s ,
\]

where $\eta_s$ is the saturation soil-moisture content. Below this value, the electrostatic forces in the soil prevent evaporation from continuing at the potential rate. Evaporation is then determined by the flux from the soil and is consequently less than the potential rate. Evaporation is assumed to cease when the soil moisture is sufficiently depleted at $\eta = \eta_e$.

\[
\text{If } \eta_e < \eta_{ec} < \eta_a \text{, then } E_{\text{soil}} = D_s \frac{\partial \eta}{\partial z} + K_s < E_p ,
\]

\[
\text{If } \eta_{ec} < \eta_e \text{, then } E_{\text{soil}} = 0 .
\]

The simulations in the present study use sand, loam, clay, and clay loam soil types. The parameters $\eta_s$, $\eta_{ec}$, and $\eta_e$ listed in Table 2 depend only on the soil type and are obtained from Clapp and Hornberger (1978).
IX. APPENDIX B: APPLICATIONS OF THE MODEL

A wide range of atmospheric phenomena can be simulated by the governing equations and physical parameterizations in the coupled earth-atmosphere mesoscale model. Yet, the model possesses many assumptions that must be considered before a numerical simulation is performed. The following is a list of some of the major constraints and characteristics of the mesoscale model.

- In general, the horizontal spacing of the grid must be larger than 10 km and the terrain slope must be smaller than 5° to satisfy the hydrostatic assumption. Smaller grid spacings may be employed for weak synoptic flow situations or when the terrain is flat.

- The first-order closure assumption, limits the application of the model to larger-scale flows where the turbulence structure can be adequately simulated by this method. This closure approximation is often called gradient-transport theory or K-theory and is based on a local, small-eddy assumption. Although it is one of the simplest parameterizations, it can frequently fail in convective mixed layers where large eddies and counter-gradient transports exist.

- Barotropic initial conditions can be used to simulate idealized boundary-layer circulations. This type of initial condition may not be appropriate when simulating observed atmospheric circulations with strong synoptic forcing.

- Baroclinic initial conditions can be used to simulate observed atmospheric circulations; however, the numerical results tend to be more unstable and may require a smaller time step.

- A dynamic-initialization technique that employs a preforecast adjustment period may be necessary to properly balance the mass and momentum fields when baroclinic initial conditions are used. Preliminary simulations indicated that this technique did not significantly improve the forecasts of the prognostic variables.

- Neumann lateral boundary conditions for the horizontal wind components have been shown to contaminate the numerical results in the interior of the domain when simulating observed atmospheric circulations with baroclinic initial conditions.

- Time-dependent Dirichlet lateral boundary conditions may produce very sharp potential temperature gradients at the lateral boundaries. The potential temperature predicted from the surface energy budget at the surface may be significantly different
than the observed potential temperature at the lateral boundaries. This feature is most pronounced during the day and may be produced by processes or surface characteristics the model cannot incorporate, such as accurate soil-moisture observations. It may also be influenced by the value of $K_0$ in the surface energy budget (Eq. (15)). This problem can be eliminated by assuming Neumann lateral boundary conditions for potential temperature when Dirichlet lateral boundary conditions are assigned for the other prognostic variables. There can also be discontinuities between the model results on the first point from the boundary and the observed values at the lateral boundary for other prognostic variables; however, this problem is not serious.

Considering the form of the governing equations (Section IIA), the boundary conditions (Section IIB), the initialization procedure and the data assimilation technique (Section IIC), and the assumptions of the mesoscale model listed above, some appropriate applications of the model could include:

- Near-real time simulations of non-precipitating, baroclinic and barotropic mesoscale phenomena in the continental United States.
- Diagnostic studies of past weather events in the continental United States.
- Simulation of the diurnal oscillations in the planetary boundary layer over flat or complex terrain.
- Evolution of the boundary layer (stable, unstable, or neutral) and its effect on larger-scale synoptic flow patterns.
- Evolution of the low-level jet over flat or complex terrain.
- Effects of soil moisture and vegetation distributions on boundary-layer and mesoscale circulations.
- Long-range transport of air-pollution or biotic agents in the atmosphere.

These applications may be simulated by the model in one, two, or three-dimensions.
### APPENDIX C: LIST OF SYMBOLS

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a$</td>
<td>Radius of the earth</td>
</tr>
<tr>
<td>$A$</td>
<td>Albedo</td>
</tr>
<tr>
<td>$b$</td>
<td>Coefficient for soil diffusivity and hydraulic conductivity</td>
</tr>
<tr>
<td>$c$</td>
<td>Phase speed</td>
</tr>
<tr>
<td>$C$</td>
<td>Volumetric heat capacity</td>
</tr>
<tr>
<td>$C_l$</td>
<td>Heat capacity of dry soil</td>
</tr>
<tr>
<td>$C_p$</td>
<td>Specific heat of air at constant pressure</td>
</tr>
<tr>
<td>$C_w$</td>
<td>Heat capacity of water</td>
</tr>
<tr>
<td>$D_t$</td>
<td>Deformation rate</td>
</tr>
<tr>
<td>$D_s$</td>
<td>Soil diffusivity</td>
</tr>
<tr>
<td>$e$</td>
<td>Turbulence kinetic energy</td>
</tr>
<tr>
<td>$E$</td>
<td>Bare-soil evaporation</td>
</tr>
<tr>
<td>$E_p$</td>
<td>Potential evaporation</td>
</tr>
<tr>
<td>$f$</td>
<td>Coriolis parameter = $2\Omega \sin \phi$</td>
</tr>
<tr>
<td>$f^*$</td>
<td>$2\Omega \cos \phi$</td>
</tr>
<tr>
<td>$F_o$</td>
<td>Longwave radiative flux</td>
</tr>
<tr>
<td>$g$</td>
<td>Acceleration of gravity</td>
</tr>
<tr>
<td>$G$</td>
<td>Solar radiation</td>
</tr>
<tr>
<td>$k$</td>
<td>von Karman constant</td>
</tr>
<tr>
<td>$K_d$</td>
<td>Horizontal exchange coefficient</td>
</tr>
<tr>
<td>$K_e$</td>
<td>Vertical exchange coefficient for turbulence kinetic energy</td>
</tr>
<tr>
<td>$K_h$</td>
<td>Vertical exchange coefficient for heat</td>
</tr>
<tr>
<td>$K_m$</td>
<td>Vertical exchange coefficient for momentum</td>
</tr>
<tr>
<td>$K_q$</td>
<td>Vertical exchange coefficient for specific humidity</td>
</tr>
<tr>
<td>$K_w$</td>
<td>Vertical-eddy-mixing coefficient for the ocean</td>
</tr>
<tr>
<td>$K_n$</td>
<td>Vertical exchange coefficient for particulate concentration</td>
</tr>
<tr>
<td>$K_{sw}$</td>
<td>Saturation hydraulic conductivity</td>
</tr>
<tr>
<td>$l$</td>
<td>Length scale</td>
</tr>
<tr>
<td>$l_o$</td>
<td>Reference length scale</td>
</tr>
<tr>
<td>$L$</td>
<td>Monin-Obukhov length</td>
</tr>
<tr>
<td>$L_v$</td>
<td>Latent heat of vaporization</td>
</tr>
<tr>
<td>$n$</td>
<td>Direction normal to lateral boundary</td>
</tr>
</tbody>
</table>
\(p\)  Pressure
\(p'\)  Deviation pressure
\(p_o\)  Reference pressure
\(p_s\)  Basic-state pressure
\(q\)  Specific humidity
\(Q\)  Diabatic heating
\(R\)  Gas constant for dry air
\(s\)  Constant height of the model top
\(S_s\)  Source term for particulate concentration
\(t\)  Time
\(T\)  Temperature
\(T'\)  Deviation temperature that is corrected for moisture
\(T_s\)  Basic-state temperature
\(T_{soil}\)  Soil temperature
\(T_v\)  Virtual temperature
\(u\)  East/west velocity component
\(u_*\)  Friction velocity
\(v\)  North/south velocity component
\(V\)  Horizontal velocity vector
\(w\)  Vertical velocity component
\(W_{air}\)  Vertical moisture flux in the atmosphere
\(W_s\)  Saturation moisture potential
\(W_{soil}\)  Vertical moisture flux in the soil
\(z\)  Cartesian vertical coordinate
\(z_s\)  Height of the ground above sea-level
\(z_o\)  Roughness length
\(\alpha\)  Constant in logarithmic vertical coordinate transformation
\(\beta\)  Constant in dissipation term for turbulence kinetic energy
\(\delta\)  Declination
\(\Delta\)  Horizontal grid spacing
\(\eta\)  Volumetric soil moisture-content
\(\eta_s\)  Volumetric soil moisture-content where evaporation ceases
\(\eta_d\)  Air-dry volumetric soil-moisture content
\(\eta_s\)  Saturation volumetric soil-moisture content
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\eta_{\text{ref}}$</td>
<td>Reference volumetric soil-moisture content</td>
</tr>
<tr>
<td>$\eta_{\text{wilt}}$</td>
<td>Wilting-point volumetric soil-moisture content</td>
</tr>
<tr>
<td>$\theta$</td>
<td>Potential temperature</td>
</tr>
<tr>
<td>$\theta_{sfc}$</td>
<td>Potential temperature at the roughness height</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Longitude</td>
</tr>
<tr>
<td>$\lambda_s$</td>
<td>Soil thermal conductivity</td>
</tr>
<tr>
<td>$\xi$</td>
<td>Logarithmic vertical coordinate</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Density</td>
</tr>
<tr>
<td>$\rho'$</td>
<td>Deviation density</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Basic-state density</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>Density of water</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Terrain-following nonorthogonal vertical coordinate</td>
</tr>
<tr>
<td>$\tau$</td>
<td>Transmittance</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Latitude</td>
</tr>
<tr>
<td>$\phi(z/L)$</td>
<td>Similarity function</td>
</tr>
<tr>
<td>$\chi$</td>
<td>Particulate concentration</td>
</tr>
<tr>
<td>$\Psi$</td>
<td>Soil-moisture potential</td>
</tr>
<tr>
<td>$\Psi_s$</td>
<td>Saturation soil-moisture potential</td>
</tr>
<tr>
<td>$\Omega$</td>
<td>Angular velocity of the earth</td>
</tr>
<tr>
<td>$\omega$</td>
<td>Transformed vertical velocity component</td>
</tr>
</tbody>
</table>
Table 1. Summary of the numerical simulations

<table>
<thead>
<tr>
<th>Case</th>
<th>Synoptic</th>
<th>Soil Moisture</th>
<th>Soil Type</th>
<th>Albedo</th>
<th>Grid Spacing</th>
</tr>
</thead>
<tbody>
<tr>
<td>NS1</td>
<td>no</td>
<td>none</td>
<td>ST1*</td>
<td>A1b</td>
<td>104 km</td>
</tr>
<tr>
<td>NS2</td>
<td>no</td>
<td>SM1*</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>NS3</td>
<td>no</td>
<td>SM2*</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>NS4</td>
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<td>SM3*</td>
<td>ST1</td>
<td>A1</td>
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</tr>
<tr>
<td>NS5</td>
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<td>SM1</td>
<td>ST1</td>
<td>A2f</td>
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</tr>
<tr>
<td>NS6</td>
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<td>SM2</td>
<td>ST1</td>
<td>A2</td>
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</tr>
<tr>
<td>NS7</td>
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<td>SM3</td>
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<td>A2</td>
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</tr>
<tr>
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<td>A3b</td>
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<tr>
<td>NS9</td>
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<td>ST2</td>
<td>A3</td>
<td>104 km</td>
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<td>SM3</td>
<td>ST2</td>
<td>A3</td>
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<tr>
<td>NS11</td>
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<td>none</td>
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<td>A1</td>
<td>74 km</td>
</tr>
<tr>
<td>NS12</td>
<td>no</td>
<td>SM2</td>
<td>ST1</td>
<td>A1</td>
<td>74 km</td>
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<tr>
<td>S1</td>
<td>yes</td>
<td>none</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>S2</td>
<td>yes</td>
<td>SM1</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>S3</td>
<td>yes</td>
<td>SM2</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>S4</td>
<td>yes</td>
<td>SM3</td>
<td>ST1</td>
<td>A1</td>
<td>104 km</td>
</tr>
<tr>
<td>S5</td>
<td>yes</td>
<td>SM1</td>
<td>ST1</td>
<td>A2</td>
<td>104 km</td>
</tr>
<tr>
<td>S6</td>
<td>yes</td>
<td>SM2</td>
<td>ST1</td>
<td>A2</td>
<td>104 km</td>
</tr>
<tr>
<td>S7</td>
<td>yes</td>
<td>SM3</td>
<td>ST1</td>
<td>A2</td>
<td>104 km</td>
</tr>
<tr>
<td>S8</td>
<td>yes</td>
<td>SM1</td>
<td>ST2</td>
<td>A3</td>
<td>104 km</td>
</tr>
<tr>
<td>S9</td>
<td>yes</td>
<td>SM2</td>
<td>ST2</td>
<td>A3</td>
<td>104 km</td>
</tr>
<tr>
<td>S10</td>
<td>yes</td>
<td>SM3</td>
<td>ST2</td>
<td>A3</td>
<td>104 km</td>
</tr>
<tr>
<td>S11</td>
<td>yes</td>
<td>none</td>
<td>ST1</td>
<td>A1</td>
<td>74 km</td>
</tr>
<tr>
<td>S12</td>
<td>yes</td>
<td>SM2</td>
<td>ST1</td>
<td>A1</td>
<td>74 km</td>
</tr>
</tbody>
</table>

*ST1 = loam in entire domain.
*A1 = albedo from summertime data sets (Matthews, 1985).
*SM1 = distribution shown in Fig. 7a.
*SM2 = distribution shown in Fig. 7b.
*SM3 = 0.284 in entire soil layer.
*A2 = a function of soil moisture (Idso et al., 1975).
*ST2 = distribution shown in Fig. 8.
Table 2. Volumetric soil-moisture content parameters as a function of soil textural class as given by Clapp and Hornberger (1978)

<table>
<thead>
<tr>
<th>Soil type</th>
<th>$\eta_r$</th>
<th>$\eta_d$</th>
<th>$\eta_s$</th>
<th>$\eta_{rel}$</th>
<th>$\eta_{wil}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sand</td>
<td>0.050</td>
<td>0.100</td>
<td>0.395</td>
<td>0.100</td>
<td>0.025</td>
</tr>
<tr>
<td>loamy sand</td>
<td>0.058</td>
<td>0.116</td>
<td>0.410</td>
<td>0.116</td>
<td>0.033</td>
</tr>
<tr>
<td>sandy loam</td>
<td>0.079</td>
<td>0.158</td>
<td>0.435</td>
<td>0.158</td>
<td>0.050</td>
</tr>
<tr>
<td>silt loam</td>
<td>-</td>
<td>-</td>
<td>0.485</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>loam</td>
<td>0.130</td>
<td>0.267</td>
<td>0.451</td>
<td>0.267</td>
<td>0.100</td>
</tr>
<tr>
<td>sandy clay loam</td>
<td>-</td>
<td>-</td>
<td>0.420</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>silty clay loam</td>
<td>0.200</td>
<td>0.300</td>
<td>0.477</td>
<td>0.300</td>
<td>0.133</td>
</tr>
<tr>
<td>clay loam</td>
<td>0.220</td>
<td>0.317</td>
<td>0.476</td>
<td>0.317</td>
<td>0.150</td>
</tr>
<tr>
<td>sandy clay</td>
<td>-</td>
<td>-</td>
<td>0.426</td>
<td>-</td>
<td>-</td>
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<td>silty clay</td>
<td>-</td>
<td>-</td>
<td>0.492</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>clay</td>
<td>0.250</td>
<td>0.325</td>
<td>0.482</td>
<td>0.325</td>
<td>0.208</td>
</tr>
</tbody>
</table>

Table 3. Soil parameters as a function of soil textural class as given by Clapp and Hornberger (1978)

<table>
<thead>
<tr>
<th>Soil type</th>
<th>$C_i$</th>
<th>b</th>
<th>$\Psi_s$</th>
<th>$K_m$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sand</td>
<td>0.3500</td>
<td>4.050</td>
<td>0.121</td>
<td>0.000176000</td>
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<tr>
<td>loamy sand</td>
<td>0.3550</td>
<td>4.380</td>
<td>0.090</td>
<td>0.000156330</td>
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<td>sandy loam</td>
<td>0.3200</td>
<td>4.900</td>
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<td>silt loam</td>
<td>0.3133</td>
<td>5.300</td>
<td>0.785</td>
<td>0.000007200</td>
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<td>loam</td>
<td>0.3066</td>
<td>5.390</td>
<td>0.478</td>
<td>0.000006950</td>
</tr>
<tr>
<td>sandy clay loam</td>
<td>0.3000</td>
<td>7.120</td>
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<tr>
<td>silty clay loam</td>
<td>0.2933</td>
<td>7.750</td>
<td>0.356</td>
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<tr>
<td>clay loam</td>
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<td>0.630</td>
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<tr>
<td>sandy clay</td>
<td>0.2800</td>
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</tr>
<tr>
<td>silty clay</td>
<td>0.2650</td>
<td>10.400</td>
<td>0.490</td>
<td>0.000001033</td>
</tr>
<tr>
<td>clay</td>
<td>0.2500</td>
<td>11.400</td>
<td>0.404</td>
<td>0.000001283</td>
</tr>
</tbody>
</table>
Fig. 1. Model topography, contour interval of 150 m. The domain for the finer grid resolution simulations is bounded by the smaller box.

Fig. 2. The sea-level pressure (mb) and surface temperature (°C) fields on (a) 12 UTC June 21, 1989 and (b) 12 UTC June 22, 1989.
Fig. 3. The surface specific humidity (g kg⁻¹) on (a) 12 UTC June 21, 1989 and (b) 12 UTC June 22, 1989

Fig. 4. The 300 mb height field (10¹ m) and selected wind barbs for (a) 12 UTC June 21, 1989 and (b) 12 UTC June 22, 1989
Fig. 5. Observed 24-hour precipitation on 12 UTC June 22, 1989. The open circles denote stations reporting a trace to 0.5 inches and the filled circles denote stations reporting more than 0.5 inches.

Fig. 6. Crop Moisture Index for June 24, 1989.
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.

Fig. 8. Soil type distribution ST2 for the central United States based on a general soil type map depicted in Foth and Schafer (1980).
Fig. 9. Numerical results from dry soil, no-synoptic-flow simulation NS1 2 m above the surface predicted for 1800 LST June 21. (a) Wind and temperature fields, contour interval of 2° C. (b) Specific humidity field, contour interval of 1 g kg⁻¹
Fig. 10. Predicted difference fields 2 m above the surface for 1800 LST June 21. (a) Wind and temperature difference fields (simulation NS2 - NS1), contour interval of 0.5° C. (b) Specific humidity difference field (simulation NS2 - NS1), contour interval of 1 g kg⁻¹. (c) As in (a), except for simulation NS3 - NS1. (d) As in (b), except for simulation NS3 - NS1. (e) As in (a), except for simulation NS9 - NS1. (f) As in (b), except for simulation NS9 - NS1.
Fig. 10. (continued)
Fig. 11. Evolution of the simulated boundary layer in eastern Oklahoma from 1200 LST June 21 to 0600 LST June 22. (a) Potential temperature profiles (°K), where ○ denote results from dry-soil simulation NS1 and □ denote results from moist-soil simulation NS3. (b) As in (a), but for specific humidity (g kg⁻¹)
Fig. 12. Predicted difference fields (simulation NS12 - NS11) 2 m above the surface for 1800 LST June 21. (a) Wind and temperature difference fields, contour interval of 0.5° C. (b) Specific humidity difference field, contour interval of 1 g kg⁻¹.
Fig. 13. Numerical results from dry soil, frontal passage simulation 12 m above the surface. (a) Wind and temperature fields for 1800 LST June 21, contour interval of 2° C. (b) As in (a), but for 0600 LST June 22. (c) Specific humidity fields for 1800 LST June 21, contour interval 1 g kg⁻¹. (d) As in (c), but for 0600 LST June 22. (e) MFC field for 1800 LST June 21 where positive values indicate moisture convergence and negative values indicate moisture divergence, contour interval 25 * 10⁷ s⁻¹. (f) As in (e), but for 0600 LST June 22.
Fig. 13. (continued)
Fig. 14. As in Fig. 13, but for dry soil, frontal passage simulation S11
Fig. 14. (continued)
Fig. 15. Predicted difference fields (simulation S2 - S1) 2 m above the surface. (a) Wind and temperature difference fields for 1800 LST June 21, contour interval of 0.5°C. (b) As in (a), but for 0600 LST June 22. (c) Specific humidity difference field for 1800 LST June 21, contour interval of 1 g kg⁻¹. (d) As in (c), but for 0600 LST June 22.
As in Fig. 15, except for simulation S3 - S1
Fig. 17. As in Fig. 15, except for simulation S9 - S1
Fig. 18. Predicted MFC difference fields (simulation S3 - S1) 2 m above the ground, contour interval 20 * 10^9 s^-1 for (a) 1800 LST June 21 and (b) 0600 LST June 22. Positive values indicate greater convergence in simulation S3 and negative values indicate greater divergence in simulation S3.
Fig. 19. Predicted specific humidity difference fields (simulation S3 - S1) 880 m above the ground for (a) 1200 LST June 21, (b) 1800 LST June 21, (c) 0000 LST June 22, and (d) 0600 LST June 22, contour interval 1 g kg$^{-1}$
Fig. 20. Vertical cross-sections difference fields (simulation S3 - S1). (a) Specific humidity difference field on 1800 LST June 21 corresponding to line A-A' in Fig. 19b, contour interval of 0.5 g kg\(^{-1}\). (b) As in (a), but for 0600 LST June 22. (c) Specific humidity difference field on 1800 LST June 21 corresponding to line B-B' in Fig 19b, contour interval of 0.5 g kg\(^{-1}\). (d) As in (c), but for 0600 LST June 22. (e) Temperature difference field 1800 LST June 21 corresponding to line A-A' in Fig. 19b, contour interval of 0.5° C. (f) As in (e), but for 0600 LST June 22.
Fig. 20. (continued)
Evolution of the simulated boundary layer in eastern Oklahoma from 1200 LST June 21 to 0600 LST June 22. (a) Potential temperature profiles (°K), where ○ denote results from dry-soil simulation S1 and □ denote results from moist-soil simulation S3. (b) As in (a), but for specific humidity (g kg⁻¹).
Fig. 22. As in Fig. 21, but for boundary layer in western Illinois
Fig. 23. Predicted difference fields (simulation S12 - S11) 2 m above the surface. (a) Wind and temperature difference fields for 1800 LST June 21, contour interval of 0.5°C. (b) As in (a), but for 0600 LST June 22. (c) Specific humidity difference field for 1800 LST June 21, contour interval of 1 g kg⁻¹. (d) As in (c), but for 0600 LST June 22. (e) MFC difference field for 1800 LST June 21, contour interval of 20 * 10⁻⁹ s⁻¹. Positive values indicate greater convergence in simulation S12 and negative values indicate greater divergence in simulation S12. (f) As in (e), but for 0600 LST June 22.
Fig. 23. (continued)
PAPER 2.

PREDICTION OF THE TRANSPORT OF INSECT PESTS:
A COMPARISON OF TWO NUMERICAL MODEL FORMULATIONS
PREDICTION OF THE TRANSPORT OF INSECT PESTS: 
A COMPARISON OF TWO NUMERICAL MODEL FORMULATIONS

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Ames, Iowa 50011

(to be submitted to Journal of Applied Meteorology)
ABSTRACT

Entomological studies have shown that certain insect pests of corn, that overwinter near the Gulf of Mexico, can be transported to the Midwest by the prevailing winds in the spring. The larval progeny of these insects may cause serious economic damage to a corn crop. Since the transport of these insects is highly dependent on the meteorological conditions, conventional numerical techniques can be employed to predict insect transport. Known insect behavior can be incorporated into atmospheric numerical models to aid in planning for insecticide application and other integrated pest management decisions.

This research evaluates the predicted regions of infestation by two different numerical model formulations. The forecasts of insect transport by advection-diffusion and trajectory methods were found to be sensitive to the particular initial conditions, boundary conditions, surface inhomogeneities, and numerical formulations used by the models. Both models produced excellent forecasts of the wind direction. The boundary-layer model overpredicted the wind speed throughout the nocturnal periods when insect transport occurs. The mesoscale model overpredicted the wind speeds near the surface late in the afternoon, but the wind speeds were too low during most of the nocturnal periods when important low-level jets occurred. Based on observational trapping data, the mesoscale model forecasted the transport of the black cutworm moth to the Corn Belt better than the boundary-layer model for the two periods examined in this study.
I. INTRODUCTION

It is known that some insect pests of corn are introduced each spring to the Midwest by the northward migration of populations that overwinter far to the south. Research by Kaster and Showers (1982) and Domino et al. (1983) has shown that nocturnal, long-range movement of noctuids, such as the black cutworm moth, is strongly correlated to particular low-level wind conditions that are common during the spring over the central United States. Showers et al. (1989b) have observed that these moths are nocturnal fliers that will take off shortly after sunset if the proper atmospheric conditions exist. The larval progeny of the migrant moths feed on stalks of young corn plants and may cause serious economic damage to a corn crop.

A boundary-layer model was applied by McCorcle and Fast (1990) (MF) to predict the introduction of wind-transported insect pests to the Corn Belt for several periods during the spring of 1988. In this model, known insect behavior was applied to air-parcel trajectory and advection-diffusion forecast methods to determine possible regions of infestation. This model has been used to aid insecticide planning and other integrated pest management decisions since 1988 (Showers et al., 1988; Showers et al., 1989a). Results from MF indicated there was close agreement between black cutworm moth trapping data and forecasted moth movement.

Observational studies by Kaster and Showers (1982) and Showers et al. (1989b) have indicated that the black cutworm moth could be transported as much as 700 km in a single night. The Great Plains low-level jet is the most likely phenomena to carry moths for such a long distance during one nocturnal period. Any numerical model that attempts to forecast the transport of insects by dispersion methods must be able to simulate this phenomena. It is widely recognized that the Great Plains low-level jet is caused primarily by the diurnal oscillation of frictional forces in the boundary layer (Blackadar, 1957) and by the diurnal oscillation of buoyancy forces over the gentle slopes in the central United States (Holton, 1967; McNider and Pielke, 1981). The boundary-layer model employed by MF has been shown by McCorcle (1988) to qualitatively produce the observed features of the Great Plains low-level jet as described by Bonner (1968).

The movement of insect pests from the overwintering area in the Gulf coast area to specific regions in the Corn Belt depends on the nocturnal wind speed and direction over the Great Plains. Any errors in the prediction of wind speed or direction will significantly affect trajectory and dispersion forecasts in an atmospheric model. Forecasts of the transport of
insect pests made by the boundary-layer model in MF are highly dependent upon the model's assumptions and formulations. The dynamic and thermal fields simulated by atmospheric models have been shown by a number of investigators to be very sensitive to initial conditions, boundary conditions, and numerical formulations (for a discussion see Pielke, 1984).

Determining the precise location and timing of infestation is a primary goal of integrated pest management; therefore, accurate forecasts of wind speed and direction are necessary to ascertain the most probable regions of infestation. It is worthwhile to compare results from different numerical models, especially if prediction of the movement of insect pests is being routinely done in an operational mode. It is important to test different model formulations to (1) evaluate the different wind fields that are predicted and (2) find any systematic errors in the model results. As reported by Pielke (1984), there have been relatively few model intercomparison studies.

Both the boundary-layer model employed by MF and mesoscale model described by Fast and McCrorcle (1990b) are used in this investigation to simulate the transport of black cutworm moths to the Corn Belt. The purpose of this study is to evaluate the predicted regions of infestations simulated by these two numerical models. By comparing the forecasted wind speed, direction, and low-level jet structure calculated by these models with observations, an estimate of the error in forecasted infestation regions can be obtained. The forecasted regions of infestation can also be compared with trapping data to determine model performance.

A brief description of the atmospheric models used in this study is presented in Section II. Meteorological data for two specific transport dates examined in this study for the spring of 1988 are outlined in Section III. Flow conditions for these dates are simulated using both numerical model formulations and the results of these simulations are presented in Section IV.
II. NUMERICAL METHODS

A. Two Numerical Formulations for the Atmospheric System

Possible infestation regions by black cutworm moths may be predicted using current meteorological forecast techniques, as shown in MF. Forecasts of this kind will depend on the particular model formulation. The assumptions and limitations of two numerical model formulations used in this study are described in this section.

The characteristics of the hydrostatic, coupled-earth atmosphere, boundary-layer model used by MF that are relevant to advection-diffusion and trajectory forecast methods include:

- NGM 850-mb geopotential height field forecasts used at model top to force time-dependent Dirichlet top boundary condition for pressure
- geostrophic wind imposed at the model top
- potential temperature and specific humidity fixed at the model top
- barotropic initial conditions
- Neumann lateral boundary conditions
- advection terms approximated by a second-order finite-difference scheme
- diffusion terms approximated Galerkin finite-element scheme
- orthogonal terrain-following cartesian coordinate system

The domain in that study extends from 20° N to 50° N latitude and from 112.5° W to 82.5° W longitude. This domain contains 25 nodes in both horizontal directions with a grid spacing of 1.25°. Nine nodes are logarithmically spaced below 119 m. Above this level, 10 additional nodes are equally spaced 205 m apart to the model top 2170 m above the surface. A time step of 600 s is used.

Several types of errors in the predicted wind field can arise from this type of model formulation. The boundary-layer model will be subject to the errors associated with the forecasted 850-mb synoptic height field from the larger-scale NGM model. These height field errors will lead to incorrect wind speeds and directions at the model top that are calculated from the geostrophic relationship. In addition, there are many times when the actual 850-mb flow is significantly ageostrophic over the central United States. Ageostrophic winds at 850 mb may occur during frontal passages and periods of strong convection.
The boundary-layer model has been shown by Astling et al. (1985) and Berri and Paegle (1990) to be insensitive to initial random noise in the wind field in the prediction of low-level circulations when significant topographic forcing is present; nevertheless, systematic errors in the initial conditions may significantly affect the numerical forecasts. Since barotropic initial conditions are employed by the boundary-layer model, the initial thermal and dynamic fields may differ substantially from the observed fields. It takes several simulation hours to "spin-up" the model to produce baroclinic flows. Hoke and Anthes (1976) demonstrated that two-dimensional simulations of the jet stream were highly dependent upon the particular initial temperature and wind field. While the model has been shown to qualitatively simulate flows over complex terrain, such as the Great Plains low-level jet (McCorcle, 1988), barotropic initial winds that lack vertical shear could ultimately lead to significant differences between the predicted flow fields and the observed mesoscale circulations.

Atmospheric models can produce serious errors if the lateral boundary conditions are incorrectly specified, as demonstrated by Anthes-and Warner (1978). Neumann boundary conditions may be incorporated into limited area models if the lateral boundaries are far from the region of interest. The migration of black cutworm moths normally occurs from southern Texas to the Corn Belt, well within the domain used by MF.

The boundary-layer model employed by MF has been modified in Fast and McCorcle (1990b) and consists of baroclinic initial conditions to more accurately simulate mesoscale phenomena. Some of the characteristics of this mesoscale model include:

• inclusion of upper atmospheric processes to approximately the 300-mb level
• time-dependent Dirichlet lateral and top boundary conditions used for the prognostic variables that is based on the objective analysis of observed fields
• geostrophic wind imposed at the model top
• baroclinic initial conditions
• advection terms approximated by a fourth-order difference scheme
• diffusion terms approximated Galerkin finite-element scheme
• nonorthogonal, terrain-following sigma coordinate system

The domain used by this model is smaller than in MF because of the density of data needed to use baroclinic initial conditions. The lack of data over the Gulf of Mexico and portions of Mexico may produce unrealistic thermal and dynamic fields for the initial conditions. The
domain for this model extends from 27.5° N to 50° N latitude and from 102.75° W to 86.5° W longitude. This domain also contains 25 nodes in both horizontal directions, but has a smaller grid spacing of 0.94°. The vertical resolution is much coarser than the boundary-layer model of MF, except near the surface. Ten nodes are logarithmically spaced below 324 m. Above this level, 17 additional nodes are equally spaced 550 m apart to the model top about 10 km above the surface. The coarser resolution in the vertical is necessary to reduce the overall computational time.

Baroclinic initial conditions are expected to approximate the initial thermal and dynamic structure of the atmosphere with a higher degree of accuracy. The mesoscale model does not require several simulation hours to "spin-up" to attain baroclinic flow fields. Also, the incorporation of observed fields at the lateral boundaries had a profound impact on the interior solution as suggested by Fast and McCorcie (1990b). Nevertheless, some numerical noise may be introduced when there are discontinuities between the model results near the lateral boundaries and the observed fields.

Profound differences between the solution of the advection equation for second and fourth-order techniques were found by Trembeck et al. (1987). The fourth-order scheme was clearly superior and more accurately represented advection in a simple numerical test. The numerical representation of the advection terms on the numerical results of a mesoscale model may not be as pronounced, since the prognostic variables are highly dependent on many other physical processes in the governing equations.

B. Formulation for the Insect Transport

Both of the models described in the previous section employ advection-diffusion and trajectory methods to predict the transport of insects. Insects are treated as a relative concentration, since the exact initial density of moths is impossible to determine. The prediction of the relative concentration in the boundary-layer model is made by following equation written in spherical coordinates:

$$\frac{D\chi}{Dt} = \frac{\partial\chi}{\partial t} + \frac{u}{a\cos\phi} \frac{\partial\chi}{\partial \phi} + \frac{v}{a} \frac{\partial\chi}{\partial \phi} + \frac{w}{\partial \xi} + \frac{\partial}{\partial \xi} \left( K_a \frac{\partial \chi}{\partial \xi} \right) \nabla \cdot \chi + S_x + \nabla \cdot \left( K_d \nabla \chi \right)$$

where \(\chi\) is the relative concentration, \(u\) and \(v\) the horizontal velocity components, \(w\) the vertical velocity component, \(S_x\) the source term, \(K_x\) the vertical diffusion coefficient for the moth plume, \(K_d\) is the horizontal diffusion coefficient, and \(\alpha\) and \(L\) are parameters in the
vertical grid transformation. The prediction of relative moth concentrations throughout the
domain of the mesoscale model is made using a similar equation:

\[
\frac{\partial \alpha}{\partial t} + \frac{1}{\cos \phi} \frac{\partial \alpha}{\partial x} + \frac{1}{a} \frac{\partial \alpha}{\partial \phi} + \frac{\omega}{\chi} \frac{\partial \alpha}{\partial \chi} = \left[ \frac{s}{s-z_o} \right]^2 \frac{\partial}{\partial \chi} \left( K_x \frac{\partial \alpha}{\partial \chi} \right) \frac{\alpha}{\sigma} + S_x + \nabla \cdot (K_v \nabla \chi) \tag{2}
\]

where \( \omega \) is the transformed vertical velocity component, \( s \) the constant height of the model
\text{top}, \( \sigma \) the terrain-following coordinate, and \( z_o \) the elevation of the terrain.

In both numerical models, the transport of black cutworm moths is calculated only
during the nocturnal periods. At sunset, the insects are assumed to migrate to a cosine-
squared vertical distribution about an average 500 m above the ground in the boundary layer
model. This assumption is also employed in the mesoscale model, except that an average
height of 880 m is used because of the coarser resolution in the vertical. To simulate moth
migration on southerly winds, flight is prohibited for grid points with northerly wind
components in the boundary-layer model. The mesoscale model takes this assumption one
step further, and removes these insects to the ground to simulate the fallout potential of
moths near frontal zones.
III. CASE STUDIES

Two transport periods from the 1988 spring growing season are used to demonstrate the similarities and differences predicted by the numerical models. These two periods are chosen because southerly winds shortly after sunset in southern Texas indicate that the transport of black cutworm moths was possible. In addition, significant nocturnal low-level jets are evident over much of the central United States. The effect of baroclinic initial conditions on the simulated flow fields can be evaluated by the mesoscale model because strong cold fronts are present during both of these periods. Barotropic initial conditions cannot reproduce the detailed vertical gradients in the initial dynamic and thermal fields in these circumstances.

The first case examined in this study covers two nocturnal periods from March 22 to 24, 1988, and represents the first significant transport date of that spring. The second case that is examined covers three nocturnal periods during May 6 to 9, 1988, and probably represents the last economically significant transport case of the spring. A brief description of the surface synoptic features that are relevant to the transport of the moths is presented here.

At 00 UTC March 23, a low-pressure system in northern Iowa was located on a cold front that extended from Iowa through central Kansas into northern Texas. Strong southeasterly winds ahead of the front existed at sunset in southern Texas. During the next 12 h the low moved to the northeast into the upper peninsula of Michigan as the cold front advanced slowly to the southeast. Southerly winds were common near the surface ahead of the front throughout the first evening. The following day the cold front became stationary and finally dissipated by 00 UTC March 24 as a second cold front was pushing into the northwest portion of the Great Plains. During the next evening, the second cold front pushed into the central United States and extended from North Dakota through central Kansas into northern New Mexico. Ahead of the front, southerly surface winds were again present throughout the central and southern Great Plains. The migration of black cutworm moths for this period was also examined by MF and the synoptic features are described in more detail in that study.

At 00 UTC May 7, a low-pressure system in eastern Montana was located on a front that extended from Montana through eastern Colorado into eastern New Mexico. During the first evening, the low remained stationary and the northern portions of the front became occluded. The rest of the cold front pushed slowly into eastern South Dakota through
western Kansas into western Texas. Strong southeasterly winds ahead of the front were present from the Gulf coast into northern Minnesota. This cold front remained relatively stationary during the next day and second evening from 12 UTC May 7 and 12 UTC May 8. The low-pressure system moved only to the western Dakotas by 12 UTC May 8. The following day, the front slowly pushed to the east and extended from Minnesota through eastern Missouri into central Texas by 00 UTC May 9. Strong westerly winds moved into the northern and central plains as the cold front accelerated and moved to the southeast during the third evening. By 12 UTC May 9, the low moved into northern Minnesota, and the cold front stretched from Michigan through western Tennessee into central Texas.

For both of these periods, southerly winds over the overwintering area of the black cutworm moth near the Gulf coast in southern Texas indicated the initiation of migration was possible. The persistent southerly flows near the surface and the low-level jets over the southern plains indicated that these moths could be transported a great distance. The cold fronts in the March case never moved into the southern Great Plains during the two-day time period. The northeast-to-southwest orientation of the front in the central plains may have acted as a barrier to hinder transport of moths to the central and western portions of the Corn Belt. The cold front in the May case was relatively stationary for the first two days, which permitted strong southerly flows well into Canada for an extended period. This persistent southerly flow presented ideal conditions for the transport of moths to the Corn Belt. The cold front eventually accelerated and pushed towards the southeast through the southern plains, and the chances of transport of moths to the Corn Belt on the third evening of the May case were greatly diminished.
IV. NUMERICAL RESULTS

A brief description of the simulations performed in this study is summarized in Table 1. Each of the simulations listed in Table 1 is integrated for 48 h, except for simulation 5 which was integrated for 36 h. The numerical results have been compared to a select group of RAOB stations in the central United States which are depicted in Fig. 1.

A. Simulations for March 22 - 24, 1988

The mesoscale model was initialized with the observed temperature, specific humidity, and wind fields from 12 UTC March 22, 1988 (simulation 1 from Table 1). The top boundary was forced with the time-dependent 300-mb observed geopotential height, potential temperature, and specific humidity fields from 12 UTC March 22 to 12 UTC March 24. The winds at the model top were diagnosed from the geostrophic relationship. Time-dependent Dirichlet lateral boundary conditions are used for all the prognostic variables, except for potential temperature which used Neumann boundary conditions.

The boundary-layer model was initialized using the 850-mb heights from 12 UTC March 22, 1988 (simulation 4 from Table 1). The initial interior specific humidity is determined by the Clausius-Clapeyron relationship with an assumed 75% relative humidity. Potential temperature is then diagnosed by employing the equation of state, virtual temperature relationship, and Poisson's equation. The initial pressure distribution is assumed to be barotropic and does not vary with height. The initial winds are diagnosed from the geostrophic relationship, except in the lowest 200 m where they are forced to logarithmically approach zero at the surface. The top boundary was forced with the time-dependent 850-mb geopotential height fields and the winds were set equal to their geostrophic value. Neumann lateral boundary conditions are employed for all prognostic variables.

Clear-sky conditions and dry soil was assumed for both models.

1. Wind field forecasts

The predicted wind fields from the mesoscale model and the observations at approximately 881 m above the terrain are shown in Figs. 2 and 3. This level is chosen because (1) it falls within the climatological average of 500 to 1000 m above the terrain for low-level jets in the Great Plains (Bonner, 1968) and (2) it is the average height of the initial vertical concentration of insects. Values at the individual RAOB stations were interpolated
to the 881 m level above the terrain for comparison purposes. The mesoscale model was able to qualitatively forecast the overall structure of the wind field when compared to observations. Table 2 lists the differences of wind speed and direction between the mesoscale model and the observations at the individual RAOB stations. The wind speeds are consistently underpredicted at this level, except at 00 UTC March 24 where the wind speeds were slightly faster than the observations. While there are some large errors in the wind speed, which are probably due to the forecasted location of the frontal zone in the Central Plains, there are many instances where the model quantitatively predicts the wind speed. The model results were within 2.5 m s\(^{-1}\) of the observed value for 28 data points, out of 68, as listed in Table 2. The mesoscale model was able to reproduce the observed maximum wind speed region in northeast Texas, but the magnitude was 5 m s\(^{-1}\) too low.

The height of the simulated low-level jet occurred 324 m above the ground throughout much of the domain for both nocturnal periods. The height of the observed low-level jet in the south-central United States at 12 UTC for both morning periods occurred between 434 to 1185 m above the terrain. This may also indicate that the coarse vertical resolution of the mesoscale model may be able to predict the height of the low-level wind speed maximum.

The model was able to predict the wind direction at the 881 m level throughout the period with a much higher degree of accuracy than wind speed as shown in Table 2. The simulated wind directions were within 10° for many of the stations, and most of the model results were within 35° of the observations. For example, for 00 UTC March 23, 5 stations were within 10° and 12 stations where within 10° to 35°. The model results differed from the observations by more than 35° for only 16 data points, out of a total of 68 for the entire period.

The predicted wind fields by the boundary-layer model of MF at approximately 529 m above the surface are depicted in Fig. 4. This level was chosen because (1) it is the average height of the initial vertical concentration of insects and (2) it is close to the 881-m level used in Fig. 3. Even though the levels in Figs. 3 and 4 differ by about 350 m, it is apparent that a much stronger low-level jet was produced by the boundary-layer model at that level. The height of the low-level jet predicted by the boundary-layer model was 529 to 733 m above the surface, so that these resulting wind speeds in Fig. 4 are expected to be faster than those of the 881 m level in the mesoscale model. While the wind speed results at 00 and 12 UTC March 23 agree quite well with observations, the wind speeds near the jet core on March 24 from southern Texas to Illinois were overpredicted. As in the mesoscale model, the wind direction was predicted with a higher degree of accuracy than the wind speed.
There are several reasons why the low-level jet predicted by the boundary-layer model was significantly stronger than the one in the mesoscale model. Paegle et al. (1984) have suggested that low-level jets can be influenced by topographic flow-blocking resulting from increased nocturnal stratification. A large portion of the domain of the boundary-layer model extended over Mexico and the Gulf of Mexico. The acceleration of the nocturnal winds between the southwest side of the Bermuda High and the Mexican Plateau could influence the development and magnitude of the Great Plains low-level jet, as originally suggested by Wexler (1961). Holton (1967) demonstrated that the buoyancy driven flows across the gently sloping Great Plains may create ageostrophic wind components, which when coupled to the diurnal frictional oscillation in the boundary-layer (Blackadar, 1957) may result in low-level jets. Thermally-induced ageostrophic components predicted by the boundary-layer model over the eastern Mexican plateau may have also enhanced the simulated low-level jet. In the mesoscale model, the Mexico and the Gulf of Mexico were omitted from the domain because of the lack of upper-air data. This assumption may have restricted important mechanisms that influence the magnitude of observed low-level jets in the Great Plains. The stronger low-level jet predicted by the boundary-layer model may also result from the higher vertical resolution used in that model.

To examine the performance of these models in the vertical, the predicted wind speeds and directions are plotted for a node near the Monett, Missouri RAOB. Monett is one of several stations that are situated between the overwintering area of the black cutworm moth in southern Texas, and the Corn Belt. The strength of the jet predicted by these models is most noticeable in its vertical structure. The vertical profiles for the mesoscale model are depicted in Fig. 5a-b and the vertical profiles from the boundary-layer model are shown in Figs. 6a-b.

The mesoscale model predicted a shallow jet to form on the first night by 2100 LST March 22, as illustrated in Fig. 5a, that has a peak wind speed of 16 m s$^{-1}$ at 0000 LST March 23. The forecasted wind speed and direction profiles for the 1800 LST March 22 period show an excellent agreement with the observations in the lowest 3 km, except for a region above 1 km where the model wind speed by 5 m s$^{-1}$ to slow. At 0600 LST March 23, the jet speed was 4 m s$^{-1}$ slower and about 100 m lower than the observed profile. The model predicted a more southerly wind component above 1 km than the observations. In the late afternoon at 1800 LST March 23, the low-level wind maximum was nearly 10 m s$^{-1}$ faster than the observed profile as shown in Fig. 5b. A 19 m s$^{-1}$ nocturnal low-level jet developed 324 m above the surface by 2100 LST March 23 that remained relatively constant throughout the
rest of the evening. The simulated jet was only 2 m s\(^{-1}\) faster than the observed jet at 0600 LST March 24, but the jet height was 450 m too low. The wind directions were nearly the same as the observations.

The vertical profiles near Monett predicted by the boundary-layer model are clearly different in structure than the mesoscale model results as shown in Figs. 6a-b. In general, the boundary-layer wind speed profiles were much stronger above 1 km and the wind direction profiles did not exhibit as strong vertical shears near the surface when compared to the mesoscale model. The simulated wind speed profile at 1800 LST Mar 22 was overpredicted at most levels. At 0600 LST Mar 23, the boundary-layer model predicted the wind speed at the observed low-level jet maximum better than the mesoscale model; however, both models did not simulate the strong baroclinicity above 1 km at this time. The boundary-layer model predicted the wind speed to be as much as 12 m s\(^{-1}\) faster than the observations at 1800 LST March 23. By 0600 LST March 24, the simulated wind speed profile was qualitatively similar to the observed profile and correctly predicted the height of the nocturnal jet, but the model-simulated jet was as much as 5 m s\(^{-1}\) too strong.

The profiles of wind speed and direction for the boundary-layer model demonstrated the possible errors associated with forcing the model top from the larger-scale NGM 850-mb forecasts. At 12 UTC March 23 and 00 UTC March 24, the wind speed and direction was significantly different than the observations near the model top. This happened because (1) the forecasted NGM 850-mb height gradient was too strong, or (2) the actual winds were significantly ageostrophic, or (3) a combination of (1) and (2). The higher momentum forced at the model top was transported downward so that stronger wind speeds were present throughout the upper 1 km of the model domain of the boundary-layer model.

Upon comparison of the numerical results and observations in Figs. 2 - 6, it is clear that each model had its own advantages and disadvantages. The mesoscale model was able to simulate the position of the front and the surface winds during the period much better than the boundary-layer model (not shown) and exhibited stronger baroclinicity near the surface. The boundary-layer model lacked sharp vertical gradients in the prognostic variables so that the predicted surface fields lagged behind the observed fields. From this analysis, it is difficult to determine which model formulation would better simulate the long-range transport of insects.
2. Relative concentration and trajectory forecasts

As in MF, the initial relative concentration region for the mesoscale model at 1800 LST March 23 was located in southern Texas. The predicted position of the insects for the 24 h forecast valid at 0600 LST after the first night of migration is shown in Fig. 7a. The maximum concentration was located in Arkansas, although it is possible that transport could have taken place on the first night into southern Missouri and Illinois. The predicted position of the insects at the 42 h forecast valid at 0000 LST March 24 during the second nocturnal period of migration showed that transport of insects could have occurred into northern Illinois and Indiana and eastern Wisconsin. By the end of the second evening the maximum concentration was advected outside the eastern border of the model.

A summation of the relative concentration at the 881 m level for the 18, 24, 42, and 48 h forecasts of the mesoscale model is shown in Figure 8a to more clearly depict the path the black cutworm moth may have taken. Results of the advection-diffusion method agree with the significant captures in Illinois; however, this method does not explain the captures in western Iowa, southeastern Nebraska, and northeastern Kansas. The trajectory forecast for this period from the mesoscale model is shown if Fig. 8b. The final trajectories indicate that transport to southern Iowa and eastern Kansas was possible (which was not predicted by the advection-diffusion method). At the beginning of the second nocturnal period at 00 UTC March 23, a high-pressure ridge extended from Florida to northern Minnesota. The flow diverged around the ridge over Missouri and Iowa so that southerly winds persisted from Oklahoma to Minnesota and southwesterly winds occurred from Arkansas to Indiana. Since the maximum concentration at the end of the first night was located in eastern Arkansas (Fig 7a), the majority of the insects were predicted to move towards the northeast into Illinois and Indiana through the central portion of the ridge. Some of the air-parcel trajectories were located a couple of hundred of kilometers to the west in western Arkansas and eastern Oklahoma at the beginning of the second evening; therefore, these trajectories followed the flow field on the western side of the ridge into Kansas and Iowa.

The boundary-layer model predicted maximum concentrations over eastern Missouri and western Illinois after the two-night period (Fig. 9a). Trajectory forecasts from this model depicted in Fig. 9b indicated transport only into Illinois. The differences between Figs. 9b and 8b may also be partially attributed to the smaller initial trajectory region used by the boundary-layer model. Some of the initial trajectory locations in the mesoscale model were located about 100 km to the north and west of those in the boundary-layer model. The air-parcel trajectories in those locations were subject to more southerly winds. Neither the
advection-diffusion or the trajectory forecasts by the boundary-layer model predicted the observed trapping data in southern Iowa and eastern Kansas.

Based on the concentration and trajectory forecasts in Figs. 7 to 9, the mesoscale model produced a superior forecast of the transport of insects based on trapping data for this case.

B. Simulations for May 5-9, 1988

For this transport period, the mesoscale model was executed with two different starting times (simulations 2 and 3 from Table 1). The first simulation was initialized with observed data from 12 UTC, May 5 and the second simulation was initialized with data from 12 UTC, May 6. These two simulations are performed because the initiation of moth migration from southern Texas was probably occurring on both evenings. The baroclinic initialization procedure and the treatment of the lateral boundaries was the same as the March 22 case.

The boundary-layer model was initialized using the 850-mb heights from 00 UTC March 23, 1988 (simulation 5 from Table 1) because data were not available for the previous 12-h period. The barotropic initialization procedure and the treatment of the lateral boundaries was the same as the March 22 case.

1. Wind-field forecasts

The predicted wind fields from the mesoscale model and the observations at approximately 881 m above the terrain are depicted in Figs. 10 and 11. As in the March 22 case, the values of the individual RAOB stations were interpolated to the 881 m level above the terrain. The mesoscale model was able to qualitatively forecast the overall structure of low-level winds for both days. During the late afternoon, at 00 UTC, the model overpredicted the wind speeds for most of the stations as shown in Table 3; however the model underestimated the wind speeds for the early morning at the 12 UTC periods. The wind speed errors are particularly large over the Central Plains where the jet core was located. For example, the model underestimated the wind speed by 8.1 m s\(^{-1}\) at Topeka, 4.9 m s\(^{-1}\) at Omaha, and by 5.8 m s\(^{-1}\) at St. Cloud at 12 UTC May 7. At 12 UTC May 8, the model underestimated the wind speed by 10.1 m s\(^{-1}\) at Topeka, 7.5 m s\(^{-1}\) at Monett, and by 7.5 m s\(^{-1}\) at Oklahoma City as shown in Table 3.

The mesoscale model was able to predict the wind direction throughout the period with a higher degree of accuracy than the March 22 case. The model results were within 10°
for 41 data points out of a total of 77 (Table 3). There were only 11 data points where the model results differed with the observations by more than 35°.

The predicted wind fields from the boundary-layer model for the same period are depicted in Fig. 12. The boundary-layer model underpredicted the wind speed at 00 UTC May 7, but at 12 UTC, May 7 the magnitude of the wind field was closer to the observed valued than the mesoscale model. Both the boundary-layer and mesoscale model overpredicted the wind speed at 00 UTC May 8; however, the boundary-layer model again produced as superior forecast by 12 UTC May 8. As in the March 22 case, the boundary-layer model produced a much stronger low-level jet.

Vertical profiles of the simulated wind speed and direction for a node near the Monett, Missouri RAOB were again examined to evaluate model performance. The vertical profiles for the mesoscale model are depicted in Fig. 13a-b and the profiles for the boundary-layer model are shown in Figs. 14a-b.

In the mesoscale model, a weak jet formed on the first night by 2100 LST May 6 that had a peak magnitude of 13 m s⁻¹ 324 m above the terrain (Fig 13a). At 1800 LST May 6 the wind speed near the surface was overpredicted by as much as 5 m s⁻¹, although the wind directions were in excellent agreement with the observations. At 0600 LST May 7, the profile of wind speed was similar in structure to the observed profile, except that the wind speeds were 5 to 8 m s⁻¹ slower throughout the column. The mesoscale model produced a much stronger low-level jet at this location during the second nocturnal period (Fig. 5b). A maximum jet speed of 19 m s⁻¹ 324 m above the terrain was predicted for 2100 LST May 7. Although the prefrontal low-level jet during the late afternoon hours was probably due to synoptic forcing, the nocturnal increase in wind speed was most likely due to frictional decoupling of the wind above the stable boundary layer. This jet behaved more like a classical nocturnal jet as the evening progressed, and by 0600 May 9 the observed wind increased to 24 m s⁻¹ about 800 m above the ground. The model predicted that the jet speed continually decreased during the evening to reach a quasi-steady wind speed of 15 m s⁻¹ through 0600 LST May 8.

The vertical profiles for the boundary-layer model near Monett are remarkably similar to those of the March 22 case. The initial wind profile at 1800 LST overpredicted the wind speed because of NGM 850-mb heights at that time; however, barotropic initial conditions may have been a good assumption for this simulation (Fig 14a) based on this profile. The boundary-layer model had a larger error in the wind direction than did the mesoscale model. By 0600 LST May 7, the model predicted the wind speed and direction in
excellent agreement with the observations. The profiles of wind speed and direction for the second nocturnal period in Fig 14b show that the model overpredicted the wind speed at most levels, except near the low-level jet maximum of 23 m s⁻¹ about 700 m above the terrain on 0600 May 8. The boundary-layer model did not predict the observed strong vertical gradients in wind speed above and below 700 m.

2. Relative concentration and trajectory forecasts

The initial relative concentration region for the mesoscale model at 1800 May 6 (simulation 2) and 1800 LST May 7 (simulation 3) was located in central and southern Texas. The predicted position of the insects for the 24-h forecast valid at 0600 LST May 7 for simulation 2 after the first night of migration is shown in Fig 15a. The maximum concentration was located in southwestern Iowa, and transport could have occurred into Minnesota and the eastern Dakotas. The predicted position of the insects at the 42-h forecast from this simulation valid at 0000 LST May 8 showed that black cutworm moths could have traveled as far as northern Minnesota (Fig. 15b). By the end of the second nocturnal period, the maximum concentration was advected outside of the northern border of the model. The nearly stationary cold front in the western plains permitted southerly flow from the Gulf of Mexico into Canada and resulted in a significant transport of moths into the Corn Belt. In simulation 3, the concentrations forecasted for 0600 LST May 8 and 0000 LST May 9 moved in a northeasterly direction as depicted in Figs. 15c-d because a surface front advanced from the northwestern plains on May 8. During the second evening of this simulation, the maximum concentration moved into central Illinois.

A summation of the concentrations at the 881 m level for the 18, 24, 42, and 48-h forecasts for simulations 2 and 3 is shown in Figure 16a-b. Simulation 2 clearly indicated transport from the south, and simulation 3 showed that transport was from the southwest. These two regions were summed and compared to observed trapping data in Fig. 16c. The numerical mesoscale model results are supported by significant captures in southwest Minnesota, eastern Nebraska, northwest Iowa, and northern Illinois. Final trajectory positions depicted in Fig 16d for both simulations 2 and 3 depict similar possible locations of transport. As with the advection-diffusion forecasts, simulation 3 predicted a more southerly track for the trajectories.

The boundary-layer model produced maximum concentrations over central Missouri after the two night period ending on 0600 LST May 8 (Fig. 17a). The trajectory forecasts shown in Fig. 17b demonstrated that transport was possible into Northern Missouri, southern
Iowa, and central Illinois. Neither the concentrations, nor the trajectory forecasts made by this model predicted the significant trapping data in western Minnesota and eastern Nebraska. The main reason for this discrepancy was that the boundary-layer wind speeds during the first night that were too slow. The mesoscale model was able to transport the insects into Kansas and Nebraska by the end of the first night, so that on the second night the prefrontal southerly winds carried them even further north. The initialization of the boundary-layer model at 1800 LST resulted in significant errors in the flow field in the central plains because several simulation hours were required to generate significant baroclinicity.

Based on the relative concentration and trajectory forecasts in Figs. 15 - 17, the mesoscale model clearly produced superior forecasts of the transport of insects based on trapping data for this case.

C. Effect of Inhomogeneous Soil-Moisture Conditions

McCorcle (1988) and Fast and McCorcle (1990a) demonstrated that inhomogeneous soil-moisture and soil-type distributions could affect the diurnally varying buoyancy forces over the slope of the Great Plains. Fast and McCorcle (1990b) used a mesoscale model to demonstrate that heterogeneous soil-moisture distributions could affect the baroclinic mesoscale systems such as fronts and low-level jets. In that study, soil-moisture distributions were found to cool and moisten the boundary layer. The cooler thermal structure affected the low-level pressure distribution so that a cold front was weakened considerably near the surface. The intensity and structure of the low-level jet that transports insect pests may be significantly affected surface heterogeneities, such as distributions of soil moisture.

Some of the errors in the horizontal and vertical structure of the dynamic flow field may be the result of the dry-soil assumption in simulations 1 - 5. This assumption could affect the horizontal distribution of the surface sensible heat fluxes. In order to explore this possibility, both the March 22 and May 6 cases (simulations 1 to 3) were repeated, except a plausible soil-moisture distribution for the spring of 1988 was employed in the initial conditions (simulations 6 to 8 of Table 1). This soil-moisture distribution is based a composite of the spring 1988 Crop Moisture Index maps. During the spring of 1988, persistent dry regions were located over central Texas, northern Louisiana, Kansas, southern Iowa, northern Missouri, and the northern plains. Relatively moist soil-moisture conditions existed in the Mississippi river valley and in a two bands that stretched across the Great Plains.
Results of the simulations 6 to 8 indicated the thermal and moisture structure of the boundary-layer was significantly affected and the surface wind speeds were altered by as much as 3.5 m s⁻¹. For example, in simulation 7, the presence of soil-moisture ultimately reduced the wind speeds by as much as 2.5 m s⁻¹ in central Oklahoma and western Nebraska 2 m above the terrain at 1800 LST May 6 as shown in Fig. 18a. In Fig. 18b the wind speeds 881 m above the ground at this time were reduced by 2 m s⁻¹ compared to those of the dry-soil simulation in Fig. 10c. While the overall wind speed difference was rather modest, the individual velocity components were changed by as much as 4.2 m s⁻¹. This resulted in a shift of the flow at 881 m to a more westerly direction. These flow field changes could result in a 86 to 130 km difference (about one grid point for the mesoscale model) in the forecasted regions of pest infestation if this difference persists throughout the night; however, the differences between the dry-soil and wet-soil simulations diminished during the nocturnal periods. For simulations 7 and 8, the final forecasted concentration was very similar to Fig. 16c, except that the concentrations are significantly reduced in the Dakotas and central Iowa (not shown). The presence of soil-moisture produced favorable regions of convergence that increased the predicted transport of moths into Minnesota (not shown).
V. CONCLUSIONS

Two numerical model formulations are employed to predict the transport of insect pests to the Corn Belt. The boundary-layer model described by McCorcle and Fast (1990b) used barotropic initial conditions and was forced at the top by time-dependent 850 mb height fields. The mesoscale model described by Fast and McCorcle (1990b) incorporated baroclinic initial conditions and was forced at the top by a time-dependent 300 mb height field and by observed temperature and humidity fields. The prognostic variables at the lateral boundaries of the mesoscale model domain were determined from the observed momentum and thermal fields.

Observational entomological studies have indicated that transport of insect pests, such as the black cutworm moth, are highly dependent upon the meteorological conditions; therefore, conventional atmospheric numerical models may be able to predict the movement of these pests. Forecasts of insect transport are sensitive to the initial and boundary conditions used by advection-diffusion routines. These forecasts also depend upon the forecasted wind speed and direction. Many numerical models reported in the literature could be used to simulate the transport of insect pests in the lower atmosphere. These models have their own representation of the atmospheric physics, initial and boundary conditions, and grid resolution; therefore, the errors in the forecast variables will be significantly different. It is important to test different model formulations to evaluate the predicted wind field and to find any systematic errors in the model results.

The mesoscale model was found to simulate the surface frontal positions with higher baroclinicity in those regions in the lower atmosphere than the boundary-layer model. Both models produced excellent forecasts of the wind direction between 500 and 1000 m above the terrain. The mesoscale model systematically overpredicted the wind speeds 100 to 500 m above the surface in the late afternoon and underpredicted the wind speed during the nocturnal periods when important low-level jets occurred. The boundary-layer model overpredicted the wind speed near the surface throughout the forecast period for both cases examined in this study, except during the first 6 h of the May 6 case because initialization began at 1800 LST May 6. Both models predicted the observed locations of the maximum wind speed regions. The vertical structure of the wind speed and direction predicted by these models was also examined. The low-level jets simulated by the mesoscale model was shallower and occurred closer to the surface than those produced by the boundary-layer model.
Both models qualitatively simulated the transport of black cutworm moths to the Corn belt for the two periods examined in this study. The mesoscale model appeared to forecast the transport of moths better than the boundary-layer model in the March 22 case based on the trajectory forecasts. This was due to the mesoscale model's ability to produce larger vertical shears in the wind speed and direction. The mesoscale model predicted the transport of moths better than the boundary-layer model in the May 6 case based on both the concentration and trajectory forecasts. The initialization of the boundary-layer model at 1800 LST May 6 was the primary reason for the large forecast errors in the transport of insects to the Corn Belt for this case. Even though the low-level winds produced by the boundary-layer model were stronger than those of the mesoscale model, the concentration regions were not advected significantly further. This may be due, in part, to the finite-difference scheme used for the advection terms, which should be better approximated by the fourth-order scheme used by the mesoscale model.
VI. ACKNOWLEDGMENTS

This research was supported by the Iowa State University Agricultural and Home Economics Experiment Station under project 2804.
VII. REFERENCES


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*M = mesoscale model described in Fast and McCorcle (1990b).
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*yes = soil-moisture distribution based on composite Crop Moisture Index maps for the spring of 1988 - see text for details.
Table 2. Differences in wind speed and wind direction between model results and observations at approximately 881 m above the terrain for March 22, 1988 case

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*aWDD = absolute difference between observed wind direction and model results at station location.

*bWSD = observed wind speed subtracted from model results at station location.
Table 3. Differences in wind speed and wind direction between model results and observations at approximately 881 m above the terrain for May 6, 1988 case

<table>
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<th>Station</th>
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<th>WSD*</th>
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\*
\*WDD = absolute difference between observed wind direction and model results at station location.

\^WSD = observed wind speed subtracted from model results at station location.
1. Selected RAOB stations used to compare model results with observed wind fields. Station locations are overlaid on the domain employed by the mesoscale model.
Fig. 2. Wind speed and direction predicted by the mesoscale model and the observed values 881 m above the terrain. (a) Model results for 00 UTC March 23, contour interval of 2 m s$^{-1}$. (b) Observed wind speeds in m s$^{-1}$ (top) and wind direction (below) for 00 UTC March 23. (c) As in (a), but for 12 UTC March 23. (d) As in (b), but for 12 UTC March 23.
Fig. 3. Wind speed and direction predicted by the mesoscale model and the observed values 881 m above the terrain. (a) Model results for 00 UTC March 24, contour interval of 2 m s\(^{-1}\). (b) Observed wind speeds in m s\(^{-1}\) (top) and wind direction (bottom) for 00 UTC March 24. (c) As in (a), but for 12 UTC March 24. (d) As in (b), but for 12 UTC March 24.
Wind speed and direction predicted by the boundary-layer model 529 m above the terrain for (a) 00 UTC March 23, (b) 12 UTC March 23, (c) 00 UTC March 24, and (d) 12 UTC March 24, contour interval of 2 m s$^{-1}$
Fig. 5. Profiles of wind speed (○) in m s⁻¹ and direction (△) predicted by the mesoscale model for a node near Monett, Missouri and observed wind speed (●) and direction (▲) for the Monett RAOB. (a) Selected time periods during the first nocturnal period from 1630 LST March 22 to 0600 LST March 23. (b) Selected time periods during the second nocturnal period from 1630 LST March 23 to 0600 LST March 24.
Fig. 5. (continued)
(a)

Fig. 6. As in Fig. 5, except results from the boundary-layer model for a node near Monett, Missouri
Fig. 6. (continued)
Fig. 7. Relative concentration 881 m above the ground predicted by the mesoscale model for (a) 0600 LST March 23, and (b) 0000 LST March 24.
Fig. 8 Final concentration region and trajectory positions predicted by the mesoscale model. (a) Summation of the relative concentrations for the 0000 and 0600 LST periods on March 22 and 23 for the 881 m level. Black cutworm moth captures (o) and significant captures (+) during March 24 to 27. (b) Initial locations of air-parcel locations (*) and the final trajectory positions (x) on 0600 LST March 24 after two nocturnal periods.
Final concentration region and trajectory positions predicted by the boundary-layer model. (a) Relative concentration 529 m above the ground for 0600 LST March 24. (b) Initial locations of air-parcel trajectories (•) and the final trajectory positions (x) predicted by the boundary-layer model on 0600 LST March 24 after two nocturnal periods.
Fig. 10. Wind speed and direction predicted by the mesoscale model and the observed values 881 m above the terrain. (a) Model results for 00 UTC May 7, contour interval of 2 m s\(^{-1}\). (b) Observed wind speeds in m s\(^{-1}\) (top) and wind direction (bottom) for 00 UTC May 7. (c) As in (a), but for 12 UTC May 7. (d) As in (b), but for 12 UTC May 7.
Fig. 11. Wind speed and direction predicted by the mesoscale model and the observed values 881 m above the terrain. (a) Model results for 00 UTC May 8, contour interval of 2 m s⁻¹. (b) Observed wind speeds in m s⁻¹ (top) and wind direction (bottom) for 00 UTC May 8. (c) As in (a), but for 12 UTC May 8. (d) As in (b), but for 12 UTC May 8.
Fig. 12. Wind speed and direction predicted by the boundary-layer model 529 m above the terrain for (a) 00 UTC May 7, (b) 12 UTC May 7, (c) 00 UTC May 8, and (d) 12 UTC May 8, contour interval of 2 m s\(^{-1}\).
Fig. 13. Profiles of wind speed (○) in m s⁻¹ and wind direction (Δ) predicted by the mesoscale model for a node near Monett, Missouri and observed wind speed (●) and direction (☆) for the Monett RAOB. (a) Selected time periods during the first nocturnal period from 1630 LST May 6 to 0600 May 7. (b) Selected time periods during the second nocturnal period from 1630 LST May 7 to 0600 May 8.
Fig. 13. (continued)
Fig. 14. As in Fig. 12, except results from the boundary-layer model for a node near Monett, Missouri.
Fig. 14. (continued)
Relative concentration 881 m above the ground predicted by the mesoscale model for (a) 0600 LST May 7 from simulation 2, (b) 0000 LST May 8 from simulation 2, (c) 0600 LST May 8 from simulation 3, and (d) 0000 LST May 9 from simulation 3
Final relative concentration region for the 881 m level and trajectory positions predicted by the mesoscale model. (a) Summation of the concentrations for the 0000 and 0600 LST periods on May 7 and 8 from simulation 2. (b) Summation of the concentrations for the 0000 and 0600 LST periods on May 8 and 9 from simulation 3. (c) Summation of relative concentration in (a) and (b) and black cutworm moth captures (○) and significant captures (+) during May 8 to 10. (d) Initial locations of air-parcel trajectories (•), the final trajectory positions (x) from simulation 2 on 0600 LST May 8, and the final trajectory positions (□) from simulation 3 on 0600 LST May 9.
Fig. 17  Final concentration region and trajectory positions predicted by the boundary-layer model.  (a) Relative concentration 529 m above the ground for 0600 LST May 8.  (b) Initial locations of air-parcel trajectories (•) and the final trajectory positions (x) predicted by the boundary-layer model on 0600 LST May 8 after two nocturnal periods.
Fig. 18. Predicted difference fields (simulation 7 - 2) for 1800 LST May 6 for wind speed (contours) and wind components (vectors) for (a) 2 m above the terrain and (b) 881 m above the terrain
SUMMARY AND CONCLUSIONS

Thermally-induced nonclassical mesoscale circulations (NCMCs), similar to sea-breezes, may be established in the presence of horizontal gradients in soil moisture, soil type, vegetation, or snow cover. These circulations may be as important as other more thoroughly examined mesoscale phenomena, such as sea and land breezes, mountain and valley winds, and urban circulations. Several recent numerical studies have indicated that sharp gradients in soil or vegetation properties may induce mesoscale circulations in the absence of synoptic forcing. The purpose of this dissertation was to examine the effect of horizontally inhomogeneous soil moisture and soil type on the boundary-layer structure embedded in atmospheric circulations with significant synoptic flow.

A major task of this research was to modify the hydrostatic, three-dimensional, first-order closure, boundary-layer model described by McCorcle (1988) 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 anelastic, hydrostatic governing equations were transformed to a nonorthogonal grid system. The mesoscale model includes soil-layer and vegetation parameterizations. The mesoscale model consists of prognostic equations for the horizontal wind components, specific humidity, potential temperature, turbulence kinetic energy, particulate concentration, volumetric soil moisture, and soil temperature. Pressure, vertical velocity, and temperature are determined from diagnostic relationships.

For simulations that used baroclinic initial conditions in this study, the prognostic variables at the model top varied in time and were determined from an objective analysis of observed data, except for the horizontal wind components which were set equal to their geostrophic value. The lateral boundary conditions also varied in time and were based on the objective analysis of observed data. Synoptic data used for the baroclinic initial conditions and the Dirichlet lateral boundary conditions were obtained from the Unidata Scientific Data Management system.

In the first paper, two sets of soil-moisture experiments were performed to determine the magnitude and structure of the simulated NCMCs. One set of experiments consisted of several soil-moisture and soil-type distributions where no synoptic flow was imposed. The second set of experiments used the same surface characteristics, except that baroclinic initial conditions based on the observations from 12 UTC June 21, 1989 of a frontal passage in the
central United States were used to initialize the three-dimensional 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 at the time indicated that surface inhomogeneities may influence the thermal, moisture, and momentum fields associated with this front.

A control simulation of the frontal passage from June 21 to June 23, 1989 that assumed dry-soil and clear-sky conditions was performed to evaluate model performance. Numerical results from this simulation established that the mesoscale model was able to qualitatively simulate the wind and temperature field associated with the frontal passage.

As in the no-synoptic flow experiments, numerical simulations incorporating soil moisture demonstrated that evaporation 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. A mesohigh wind field was produced by the altered thermal field with wind speeds between 1.5 - 3.0 m s⁻¹ near the surface. 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. Some studies have suggested that significant synoptic flow patterns could mask or reduce the potential effects of surface inhomogeneities (Segal et al., 1989). In this study, soil-moisture and soil-type distributions were found to have an even greater effect than the no-synoptic flow experiments. While the most profound effects occurred near the surface, evaporated soil moisture was advected horizontally ahead of the cold front far from its source and transported vertically into the free atmosphere by nonlinear synoptic-scale circulations. The presence of evaporated soil moisture enhanced moisture convergence in several locations, indicating that soil moisture may play an important role in modifying the spatial distribution and intensity of precipitation.

Qualitative analysis of the results from the first paper indicated that soil moisture should not be neglected when simulating mesoscale phenomena. Spatial variations in soil moisture lead to horizontal inhomogeneities in the latent and sensible heat fluxes which affect the temperature structure near the surface. Observations of soil-moisture content and distribution in the United States are necessary because the parameterization of horizontal heterogeneous land characteristics in operational models may significantly influence short-range forecasts. Accurate soil-moisture profiles are necessary to initialize these numerical models; however, routine observations are currently not available. Remote sensing techniques, may make this possible in the future. 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.

In the second paper, the mesoscale model was used to simulate the transport of insect
pests to the Corn Belt. Entomological studies have shown that black cutworm moths, that
overwinter near the Gulf of Mexico, can be transported to the midwest by the prevailing
winds in the spring. The larval progeny of these moths may cause serious economic damage
to a corn crop. Since the transport of these insects is highly dependent on the
meteorological conditions, convectional numerical techniques can be employed to predict
insect transport. Known insect behavior has been incorporated into atmospheric numerical
models to aid in insecticide planning and other integrated pest management decisions.

McCorcle and Fast (1990) described a boundary-layer model that was used to predict
the transport of black cutworm moths to the Corn Belt in an operational model. Similar
experiments were performed in the second paper to determined the effects of initial
conditions, boundary conditions, and numerical formulations on the predicted regions of
infestation. Two transport dates during the spring of 1988 were used to evaluate the
mesoscale model. The first period was from March 22 to 24 and probably was the first
significant transport case of the spring. The second period was from May 6 to 9 and
represented the last economic significant transport case of the spring. Results from both
models were also compared to observed wind fields to determine model performance. The
effect of horizontally heterogeneous soil moisture on the nocturnal low-level jets was also
examined.

The forecasts of insect transport by advection-diffusion and trajectory methods was
found to be sensitive to the particular initial conditions, boundary conditions, numerical
formulations, and surface inhomogeneities used by these models. Both the mesoscale and
the boundary-layer model produced excellent forecasts of the wind direction. The boundary-
layer model overpredicted the wind speed throughout the nocturnal periods, when insect
transport occurred, except during the first 6 h of the May 6 case because initialization began
near sunset at 1800 LST May 6. The mesoscale model underpredicted the wind speed during
most of the nocturnal periods when important low-level jets occurred; however, the wind
speeds 100 to 500 m above the surface were consistently overpredicted in the late afternoon.
The mesoscale model was found to simulate the surface frontal positions and wind fields
better and a higher baroclinicity in the lower atmosphere than the boundary-layer model.
Both models predicted the observed locations of the maximum wind speed regions. The
nocturnal low-level jets simulated by the mesoscale model was shallower and occurred closer to the surface than those produced by the boundary-layer model.

Based on observational trapping data, the mesoscale model forecasted the transport of the black cutworm moth to the Corn Belt better than the boundary-layer model for the two periods examined in this study. In the March 22 case, this was due to the mesoscale model's ability to produce larger vertical shears in the wind direction. The initialization of the boundary-layer model at 1800 LST May 6 was the primary reason for the large forecast errors in the transport of insects to the Corn Belt for this case.

It is anticipated that the present mesoscale model described in this dissertation will be used in the future to simulate a wider variety of mesoscale phenomena. For example, data from field experiments, such as HAPEX-MOBILHY could be employed for the initial conditions of the model. This would require executing the model with a much smaller spatial resolution; however, forecast errors due to initial conditions, boundary conditions, and surface parameterizations could be evaluated in more detail.
ACKNOWLEDGMENTS

The completion of this research would not have been possible without the contribution of several individuals. First, I would like to extend my gratitude to my advisor and co-major professor, Dr. Michael McCorcle, for providing the necessary financial support and guidance for my graduate research. He also generously provided the original version of the boundary-layer model, without which, this research would not have been possible. I would like to thank Dr. Eugene Takle, my co-major professor, for his advice and editorial assistance during the preparation of this dissertation. The head of the agronomy department, Dr. John Pesek, gave me the opportunity to conduct applied meteorological research within the agronomy department. Warren Heilman and Rob White were always there to discuss theoretical and practical meteorological problems and they provided a stimulating environment in which to conduct research. Fellow graduate students Steve Finley, Richard Turner, Cathy Hamann, and Al Dutcher were extremely helpful in providing computer and modeling assistance, and most of all, moral support.

I would also like to thank Dr. D. N. Yarger, Dr. R. E. Carlson, and Dr. S. E. Taylor for serving on my graduate committee. Mary Davis generously provided a scanning machine that made several figures in this dissertation easier to reproduce. Many of the graphical programs I have written and used in this research are based on NCAR GKS software applications obtained through the Unidata System for Scientific Data Management. These software applications were originally developed by Dan Vietor in the Department of Geoscience at Purdue University.

Special thanks must be given to my parents for their love, support, and tremendous encouragement during my graduate studies at Iowa State University.
BIBLIOGRAPHY


