A Numerical Simulation of the 7 May 1985 Mesoscale Convective System

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ABSTRACT

On 7 May 1985 a mesoscale convective system (MCS) developed within a moderately forced large-scale environment for upward motion and moved into the observing network of the Oklahoma–Kansas Preliminary Regional Experiment for STORM (PRE-STORM). The initial region of convective development occurred outside the PRE-STORM network in a data-sparse area. Simulations using The Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model are produced using initial conditions from two different initialization techniques, static initialization and improved dynamic initialization, to evaluate the ability of the mesoscale model to reproduce the complex evolution and structure of this MCS. The results indicate that, even when including the special PRE-STORM data in the model initialization process, the numerical simulation that uses the initial condition from the static initialization fails to simulate the observed evolution of the 7 May 1985 MCS. This is attributed to both the relatively weak large-scale forcing for upward motion and the lack of adequate mesoscale observations of the low-level moisture distribution and wind field in the Texas panhandle and western Oklahoma. In contrast, the initial condition from the dynamic initialization approach that uses the results of a continuous four-dimensional data-assimilation technique (nudging) as a first guess for a static initialization (both of which include the special PRE-STORM data) produces a successful simulation of the MCS. This simulation captures remarkably well many of the observed and analyzed mesoscale features determined from the high-resolution PRE-STORM observing network data. Threat scores for precipitation amounts and root-mean-square errors of sea level pressure are calculated to provide an objective measure of the quality of the simulations.

1. Introduction

A mesoscale convective system (MCS) is a distinct meteorological phenomenon that is comprised of groups of thunderstorms, frequently organized into an arc-shaped leading convective line and trailed by a stratiform precipitation region (Houze et al. 1990). MCSs produce a significant proportion of warm season precipitation in the central United States and often are associated with severe weather (Newton 1967; Fritsch et al. 1981; Fritsch et al. 1986), making the prediction of MCS development and evolution an important forecast concern. Large-scale models do not predict MCSs explicitly (Schwartz et al. 1987; Junker et al. 1989), although numerous studies show that it is possible to simulate the development and evolution of MCSs with mesoscale numerical weather prediction models if the appropriate parameterization schemes are used (Zhang and Fritsch 1986, 1988b; Zhang et al. 1989). For example, Zhang and Fritsch (1986) successfully simulate the size, propagation rate, and orientation of a squall line and mesoscale convective complex (MCC) that produced heavy rainfall and flooding over Pennsylvania. Observed mesoscale features such as mesohighs, mesolows, and presquall lows are produced by the model and compare well with observations. Further numerical simulations by Zhang and Fritsch (1988c) and Zhang et al. (1989) suggest that it is possible, using a sufficiently realistic mesoscale model initialized with only standard upper-air observations, to predict the genesis and evolution of MCSs. However, a recent study of MCS development in a weakly forced large-scale environment by Stensrud and Fritsch (1994a,b) suggests that, when the large-scale forcing is weak, it is not possible to simulate the genesis and evolution of MCSs correctly from initial conditions created using only rawinsonde data and conventional model initialization techniques.

Part of the difficulty in simulating MCSs is that convective development often occurs between the sparsely distributed observing stations. In addition, the lack of mesoscale details in the initial state is a major source of error for short-range numerical weather prediction (Haltiner and Williams 1980), especially since meso-
scale features play an important role in the initiation of deep convection (Doswell 1987; Stensrud and Fritsch 1993). When the large-scale forcing for upward motion is weak, this problem becomes even more acute. As a result, numerical products often are not very useful for the short-range (0–12 h) prediction of quantitative precipitation amounts (Funk 1991). This is a serious concern of operational weather prediction, since the modernized National Weather Service is placing even more emphasis on the 6–12-h forecasts of hazardous mesoscale weather phenomena (McPherson 1994).

One approach to overcome the problems associated with the lack of mesoscale data is to use our physical understanding to maximize the use of available observations. Stensrud and Fritsch (1994a,b) subjectively analyze a model initial condition and create artificially constructed soundings to augment the observational datasets to include some important, yet poorly observed, mesoscale features that are indicated by surface, radar, and satellite observations in the model initial condition. The initial condition that includes these observed mesoscale features leads to a more accurate evolution of convection in the model than the initial condition that lacks these mesoscale details. Unfortunately, this methodology is time consuming and not easily adaptable to automation, although they argue it may be necessary in some situations.

A more practical approach to including mesoscale features in a model initial condition is to use four-dimensional data assimilation (FDDA). One method that has shown consistent success in mesoscale model simulations is Newtonian relaxation, or “nudging” (Hoke and Anthes 1976). By using nudging, Kuo and Guo (1989) assimilate time-continuous wind profiler observations into a mesoscale model and find that this assimilation method is effective in recovering mesoscale circulations that are not properly resolved by the rawinsonde observing network. Stauffer and Seaman (1990) and Stauffer et al. (1991) found that continuous assimilation of standard-resolution rawinsonde observations throughout a model integration, rather than at only the initial time, also can generate realistic mesoscale structures not resolved well by the data.

For the case studied in this paper, the large-scale forcing for upward motion is relatively weak and the initial convective development occurs between sparsely distributed observing stations. Results from numerical simulations, discussed in section 4, show that even when including special observations from the Oklahoma–Kansas Preliminary Regional Experiment for STORM (PRE-STORM), the initial condition created by a static initialization fails to lead to a successful simulation. This is attributed to the initial convective development occurring outside the PRE-STORM network, since the operational network is too coarse to detect the important mesoscale features associated with the initiation of convection on this day. This deficiency leads us to resort to FDDA, which uses observations obtained not only at the initial time, but also during the assimilation period prior to the initial time, to determine a more accurate initial condition. The particular method used in this study involves two steps: 1) the nudging method is used to produce a more accurate mesoscale initial condition, and 2) the end products from this nudging assimilation are used as the first-guess fields for a static initialization, which blends in the conventional and special PRE-STORM data using an objective analysis technique. The second step significantly improves the model’s initial fields and may be viewed as a “suboptimal interpolation.” Further development along a path toward the optimal interpolation may provide additional improvements.

The following section describes the development and evolution of the 7 May 1985 MCS and the large-scale environmental conditions at the model initial time. The model simulation system and the generation of model initial conditions are described in section 3, where two types of initialization schemes are used to generate the two different versions of the model initial conditions. Section 4 compares the model simulation results produced from the different initial conditions with observations, while summary and concluding remarks are given in section 5.

2. The 7 May 1985 MCS and synoptic setting

a. The 7 May 1985 MCS

Early on, and before 0000 UTC 7 May 1985, thunderstorms developed in northeastern New Mexico and far northwestern Oklahoma (Fig. 1a). Over the next several hours the storms in New Mexico moved into the Texas panhandle and organized into an MCS with an arc-shaped leading convective line oriented southwest-to-northeast trailed by a stratiform precipitation region (Brandes 1990). Thunderstorms also continued to persist near the frontal boundary in Oklahoma and southern Kansas, although these storms failed to organize. Infrared satellite images (Fig. 1) show that the MCS in Texas moved east-southeastward into Oklahoma after 0300 UTC 7 May. Cloud-top temperatures indicate that as the convective line moved into Oklahoma, it became more intense, as indicated by cloud-top temperatures of less than −65°C. The maximum areal extent of the −52°C infrared cloud-top temperatures occurred at 0900 UTC, after which time the stratiform rain region began to erode on its western boundary (Brandes 1990). The convective line began to decay as it moved into eastern Oklahoma.

This event produced numerous reports of large hail, some approaching baseball size, in western Oklahoma and the Texas panhandle (NOAA 1985). In addition, rainfall totals exceeding 14 cm occurred in far northwestern Oklahoma, near Woodward (WWR in Fig. 2), causing damage to crops, roadways, bridges, and farm equipment estimated at over $750,000. Forecasters ac-
Fig. 1. Infrared satellite pictures from the GOES-West earth satellite on 7 May 1985 at (a) 0000 UTC, (b) 0300 UTC, (c) 0600 UTC, (d) 0900 UTC, and (e) 1200 UTC.
accurately predicted that severe storms would develop in the Texas and Oklahoma panhandles, southwest Kansas, and southeastern Colorado, and that these storms might organize into an MCS after sunset (Meitin and Cunningham 1985). However, forecasts of MCS development and evolution are very difficult (Schwartz et al. 1987), and it is not surprising that the development of the MCS in the Texas panhandle was not entirely as expected. Therefore, although severe storms did develop as forecast, more accurate numerical guidance on MCS development and evolution, including total rainfall, would have been helpful to the forecasters.

This MCS was analyzed in detail by Brandes (1990) and Zheng (1993) using both the standard surface and upper-air data and the special PRE-STORM data (Fig. 2). Features exhibited by the MCS include a leading squall line, a trailing stratiform rain region, mesoscale updraft and downdraft regions, rear inflow, wake low, mesohigh, presquall low, and a mesovortex. As Brandes (1990) indicates, this MCS is particularly interesting for several reasons: (a) it never reached a steady-state condition, since MCS decline began immediately after growth finished; (b) it had a well-defined mesovortex that persisted for most of the MCS lifetime and was a significant feature in the MCS organization; and (c) it was sampled by many special observational systems used during PRE-STORM. These special datasets allow for a much more detailed comparison of the model output with observations than typically is available. While Brandes (1990) provides documentation of the large-scale environment at 0000 UTC 7 May that was associated with this MCS, several analyses are presented here as well. The reader is referred to Brandes (1990) for additional information.

b. Synoptic setting

Surface observations at 0000 UTC 7 May (Fig. 3a) indicated that a weak, large-scale stationary front extended west-to-east through the northern Texas panhandle and northern Oklahoma. A weak pressure gradient was present across the MCS genesis region, with associated light surface winds. Temperatures were highest, and mixing ratios lowest, to the south and southwest of the genesis region. Warm advection occurred across the Texas panhandle and most of Oklahoma. The western edge of the region of mixing ratios greater than 10 g kg

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Winds at 700 mb (Fig. 3c) over the genesis region were weaker than at the 850-mb level and the wind directions veered with height, as expected, in a region of warm temperature advection. Values of warm advection have decreased substantially from 850 mb, with values near 3°C (12 h)

The winds continued to veer up to 500 mb (Fig. 3d), where they became west to west-southwesterly over the genesis region. The wind and height fields clearly define a weak short-wave trough near the eastern boundaries of New Mexico and Colorado, while a weak ridge exists over Oklahoma and Kansas. The 500-mb vertical velocity field calculated
from the standard upper-air data using the kinematic method shows that the short-wave trough over the Texas panhandle was associated with values of upward motion of no more than 1.6 μb s⁻¹ (Fig. 3d). This value of upward vertical velocity is smaller than the values frequently found in association with organized mesoscale convective weather systems, where 5 μb s⁻¹ appears to be a small value and 500-mb vertical motions may exceed 12 μb s⁻¹ (Maddox 1983; Hoxit and Chappell 1975). However, this value of vertical velocity is not as small as the values (<1.0 μb s⁻¹) associated with very weak large-scale forcing by Stensrud and Fritsch (1993).

3. Model simulation system and generation of model initial conditions

a. Model simulation system

The numerical model employed in this study is the Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (Anthes and Warner 1978; Anthes et al. 1987). It is a hydrostatic primitive equation mesoscale model that uses the σ terrain-following vertical coordinate, where \( \sigma = (p - p_c)(p_t - p_s)^{-1} \), \( p \) is the pressure of the model layer, \( p_t \) is the pressure at the model top (100 mb), and \( p_s \) is the surface pressure. The simulations reported in
this study used 23 $\sigma$ layers defined as 0.025, 0.075, 0.125, 0.175, 0.225, 0.275, 0.325, 0.375, 0.425, 0.475, 0.525, 0.575, 0.625, 0.675, 0.725, 0.775, 0.820, 0.865, 0.910, 0.945, 0.970, 0.985, and 0.995. The model also included the following:

- a two-way interactive nested-grid procedure (Zhang et al. 1986). The coarse-grid mesh (CGM) covers $46 \times 61$ grid elements with 75-km grid spacing (Fig. 4) and the fine-grid mesh (FGM) covers $49 \times 61$ grid elements with 25-km grid spacing. The coarse- and nested-grid terrain fields are obtained by using a Cressman-type objective analysis scheme applied to 30- and 5-min resolution terrain datasets, respectively, archived at NCAR.
- an improved implicit Fritsch-Chappell convective parameterization scheme (Fritsch and Chappell 1980; Zhang and Fritsch 1986) for the FGM, and an improved Kuo convective parameterization scheme (Anthes et al. 1987) for the CGM.
- an explicit bulk convective scheme containing equations for cloud water, rainwater, ice, and snow that follows the parameterizations of Lin et al. (1983), Rutledge and Hobbs (1985), and Hsie et al. (1984). This scheme is described in detail by Zhang (1989).
- a modified version of the Blackadar high-resolution planetary boundary layer scheme as described by Blackadar (1976, 1979), Zhang and Anthes (1982), and Zhang and Fritsch (1986). These physical parameterization schemes are similar to those used to study other MCSs by Zhang et al. (1989) and Stensrud and Fritsch (1994).

### b. Generation of model initial conditions

Two basic types of initialization techniques are used in this study: static initialization (SI) and dynamic initialization (DI). The SI is the standard initialization option in use with the PSU–NCAR mesoscale modeling system and is similar to those used in operational weather prediction centers. The initialization package also includes an option to incorporate any special datasets, such as the PRE-STORM data, into the input surface and upper-air datasets used to create the model initial condition. First-guess fields used by the standard initialization package are produced by interpolating the National Meteorological Center (NMC) 2.5° latitude–longitude global analyses to the coarse-mesh grid points. These fields include analyses of temperature, relative humidity, and the horizontal wind components at standard pressure levels, as well as sea level pressure and ground temperature. The first-guess fields are modified by blending in standard surface and rawinsonde data, plus any data from special experiments or other nonconventional sources, through a successive-correction type of objective analysis procedure (Benjamin and Seaman 1985). The analyzed fields then are interpolated to the model sigma levels, the integrated mean divergence in a column is removed to minimize noise (Washington and Baumhefner 1975), and the coarse-mesh variables are interpolated to the nested-mesh grid points (Zhang et al. 1986). The SI valid at 0000 UTC 7 May is created by using both the conventional surface and upper-air observations plus the special PRE-STORM surface and upper-air observations.

The DI procedure uses the nudging technique to perform FDDA during a 12-h preforecast (assimilation) period defined from 1200 UTC 6 May to 0000 UTC 7 May. During the preforecast integration, the nudging technique is employed to move the model state toward the analyzed state by adding, to one or more of the prognostic equations, artificial tendency terms based on the difference between the two states (Anthes 1974; Hoke and Anthes 1976). These artificial tendency terms appear in the model equations as

$$\frac{\partial p* \alpha}{\partial t} = F + G_\alpha W_t p^*(\alpha_0 - \alpha),$$

where $\alpha$ is the nudged variable, $p^* = p - p_r$, $F$ is the associated model forcing term, $G_\alpha$ is a nudging coefficient that determines the relative weight of the nudging term (with respect to $F$), $W_t$ is a four-dimensional weighting function, and $\alpha_0$ is a gridded field of $\alpha$ obtained from observations.

The nudging coefficient $G_\alpha$ is selected based on the following considerations (Stauffer and Seaman 1990): (a) the nudging timescale should be as close as possible to the timescale of the slowest physical adjustment process and (b) the computational stability criterion must be satisfied, that is, $G_\alpha \leq 1/\Delta t$ (Hoke 1976). Typical values of $G_\alpha$ for wind and temperature range from $10^{-4} \text{ s}^{-1}$ to $10^{-3} \text{ s}^{-1}$ (Stauffer and Seaman 1990). For the simulations in this paper, a small value of $G_\alpha = 10^{-5} \text{ s}^{-1}$ is chosen for the mixing ratio, because the
analyzed mixing ratio field has relatively poor resolution in comparison to the mixing ratio field suggested by the surface data [as also seen in Stauffer and Seaman (1990)]. The values of the nudging coefficients are listed in Table 1.

The DI procedure used in this study is the same as that discussed in Stauffer and Seaman (1990) and Stauffer et al. (1991) in which the model fields are nudged toward analyzed upper-air fields at 12-h intervals (here defined as 1200 UTC 6 May and 0000 UTC 7 May) while also being nudged toward analyzed surface fields at 3-h intervals. The analyzed fields used by the DI procedure are created using the conventional surface and upper-air observations plus the special PRE-STORM surface and upper-air observations. However, owing to the temporally discontinuous nature of the special upper-air data, the PRE-STORM data are included in these analyses only at a single time (0000 UTC 7 May). Similar applications of nudging have been used successfully by Wang and Warner (1988) and Kuo and Guo (1989).

Model results using the initial condition created by the DI approach (not shown) indicate that several of the observed features of the MCS are not reproduced in the model simulation, including the wake low and the mesohigh, while the model underproduces rainfall in southwestern Oklahoma and Texas and overproduces rainfall in Kansas. This result necessitates the development of an improved DI procedure, denoted as ADI. In ADI, the model fields generated by the DI at 0000 UTC 7 May are used as first-guess fields in the static initialization. These first-guess fields then are modified using the objective analysis technique of Benjamin and Seaman (1985) by blending in both the conventional and special PRE-STORM surface and upper-air data at 0000 UTC 7 May. The use of DI allows the model to generate realistic mass–momentum imbalances and to improve the low-level moisture distribution during the 12-h assimilation period. Model results show that these features, in particular, are necessary to trigger convection in the proper location and at the correct time during the subsequent numerical simulation.

The additional step of interpolation incorporates any observed features not reproduced well by the model simulation during the assimilation period. This two-step procedure significantly improves the analysis of the initial fields at 0000 UTC 7 May, since there remain subtle, yet significant, differences between the model nudged fields from DI and the observations at the end of the assimilation period. Conventional DI is not sufficient to reproduce the observations in detail (see the next subsection), since nudging only applies a gradual correction to the model equations and cannot produce a precise fit between the model fields and observations at the end of the assimilation period. Symbolically, the above step of interpolation can be expressed as

\[ v^a = v^f + K(v^{ob} - Hv^f), \]

where \( v \) is a state vector representing all the model variables at all the grid points, \( H \) is the observation matrix, \( K \) is the interpolation matrix, superscript \( f \) stands for the forecast of the nudging integration (i.e., the end product of DI), superscript \( ob \) stands for observations, and superscript \( a \) stands for the analysis. If both the observation error and (nudging) forecast error covariance matrices are known, then the optimal \( K \) can be found through the "optimal interpolation assimilation procedure"—a procedure that minimizes the analysis error (Ghil and Malanotte-Rizzioli 1991). If the forecast error covariance matrix is computed with the assimilation integration, then \( K \) can be determined similarly to the "gain matrix" in the Kalman filter. Although these advanced methods are beyond the scope of this paper, they are worthy of consideration for future studies, since they will optimize the analysis (though their computational costs will be very high).

c. Comparison of model initial conditions with observations

Since the representation of the planetary boundary layer is extremely important to any numerical simulation of deep convection, we focus our attention upon a comparison between the observed and model system created initial conditions at the 850-mb level. The observed 850-mb mixing ratio (cf. Fig. 3b) at 0000 UTC 7 May shows a well-defined moist tongue. The shape and position of the axis of the moist tongue can be identified reasonably well from the standard upper-air data and PRE-STORM soundings, with the axis passing through CSM and WWR in western Oklahoma. However, the precise location of the western boundary of the analyzed moist tongue is difficult to determine owing to the lack of special PRE-STORM data extending westward into Texas. The 850-mb mixing ratio fields produced by the ADI and the SI procedures show subtle, yet significant, differences between each other and the observations. For example, the gradients of mixing ratio in west Texas produced by the ADI procedure (Fig. 5a) are much stronger than those produced by SI (Fig. 5b), and even stronger than suggested by the available observations (likely owing to the lack of data with mesoscale resolution in west Texas). Note that the region of mixing ratio greater than 12 g kg\(^{-1}\) in southwestern Oklahoma is reproduced by both ADI and SI (as shown in Figs. 5a and 5b) but not by the conventional DI owing to the gradual correction process of nudging being unable to overcome errors in either the model physics, the model ini-
tial condition at 1200 UTC 6 May, or both (also see Zheng 1993). This region of high mixing ratio is found to be very important for the successful simulation of the MCS and justifies the additional step of interpolation used in the ADI approach.

The ADI procedure also creates a region of mixing ratio greater than 10 g kg⁻¹ in northeastern New Mexico that is not reproduced by the SI. Cross sections of relative humidity from the ADI and the SI initial conditions (Fig. 6) indicate that the ADI produces high relative humidities over the northeast corner of New Mexico, which exceed 80% from the near surface to 500 mb, whereas the relative humidities from the SI are less than 50% throughout the troposphere. Infrared satellite imagery (Fig. 1a) and radar data (not shown) at 0000 UTC 7 May indicate a developing area of deep convection over the northeast corner of New Mexico. Examination of the model output during the assimilation period indicates that this region of high mixing ratio values was produced by the initiation of deep convection in the model. This feature is not reflected in the initial mixing ratio fields generated by the SI since no upper-air observations are available in this region. It is important to note that as the model is started at 0000 UTC 7 May, the convective parameterization schemes are inactive in all of the simulations presented in this study. Thus, even though convection was active during the assimilation period of the DI approach, the model must activate convection anew after the initialization at 0000 UTC 7 May.

Differences also are seen in the wind fields of the two initialization procedures. The initial wind field at 850 mb from ADI (Fig. 7a) captures both the low-level jet stream (as indicated by the 850-mb wind speeds in excess of 10 m s⁻¹) and the wind-shift line along the border of northwest Oklahoma and Kansas (see Fig. 3b). In contrast, the wind field from SI (Fig. 7b) shows that this wind-shift line is not reproduced well, although the low-level jet across central Oklahoma and central Kansas is captured. The low-level jet stream from ADI stretches southward from southern Kansas into far west-central Texas (Fig. 7a), in contrast to the jet stream from SI that only extends from southern Kansas into far northern Texas (Fig. 7b). The wind field from the conventional DI shows that the jet stream is located only in the data void south of the Oklahoma border in north-central Texas (not shown). Thus, it is the blending of the conventional DI output, with the jet stream located in north-central Texas, and the observations, with the jet stream located in central Oklahoma and southern Kansas, that produces the low-level jet stream that stretches from central Texas into far southern Kansas. Clearly, the ADI approach, by combining the dynamically consistent first-guess field at the same resolution as the mesoscale model with the observations, provides an improved initialization on this day. The associated wind shift and convergence to the northwest of the jet core area, plus a moister boundary layer (as shown in Fig. 6a), provide a very favorable condition for convective initiation and development.

4. Numerical simulations

Numerical simulations from 0000 to 1200 UTC 7 May are produced using the two different versions of initial conditions (ADI and SI). Results from the model simulations are compared with both observations and the results of Brandes (1990), Zheng (1993), and Brandes and Ziegler (1993). These comparisons indicate that the simulation produced using the ADI initial condition is able to reproduce the observed evolution of events to a reasonable degree of accuracy. Therefore, we focus upon the results from the ADI simulation. The success of the ADI-produced initial condition to repro-
modeled convection is slightly to the west of that indicated by satellite, but over the next 3 h this convective area moves southeastward and begins to organize in agreement with observations. While the simulated movement of convection is somewhat different from the observed, in part owing to the adjustment process during the first few hours of the model simulation, the simulation maintains a large degree of accuracy in the placement of convection.

By 0600, observations indicate that the MCS was located in the northeastern portion of the Texas panhandle and was beginning to enter into the special PRE-STORM observing network. The MCS already had a distinct arc-shaped leading convective line oriented southwest to northeast with a trailing stratiform precipitation to the northwest (Brendas 1990). Thunderstorms not associated with this MCS were located along the frontal boundary in northwestern Oklahoma and southern Kansas. These storms had decreased in intensity during the past few hours and only light showers were reported at this time.

Results from the model simulation indicate that the observed evolution of events during the past 3 h is reproduced well. The simulated sea level pressure pattern is in good agreement with observations (Fig. 8), particularly with respect to the placement of the convective outflow boundaries. The lack of high-resolution surface data in the Texas panhandle preclude the verification of the mesohigh strength and precise location. Nevertheless, the simulated mesohigh is consistent with the conceptual model of squall mesosystems where the mesohigh typically is formed in the early stages of MCS development (Fujita 1963).

The accuracy and usefulness of a model simulation is often best evaluated by comparing the observed and modeled rainfall fields, since precipitation is one of the most sensitive and important model output fields. The model exhibits skill in both the pattern and magnitude of the hourly total rainfall field as observed and predictions at 0700 UTC (Fig. 9). The simulated rainfall amounts show a maximum hourly total rainfall of 9.6 mm in the Texas panhandle with a secondary rainfall maximum of 7.25 mm in northern Oklahoma, both only slightly less than the observed rainfall maximum of just over 10 mm. The simulated rainfall amounts in Kansas, however, are significantly smaller than the observed, suggesting that the model has difficulty producing deep convection to the north of the frontal boundary. Additionally, the predicted rainfall maximum in the Texas panhandle is positioned to the west of the observed location, indicating that the model squall line is moving slightly slower than the observed one.

During the ensuing hours, the simulated MCS continues to organize and strengthen as it gradually enters the PRE-STORM network. At 0900 UTC, observations show that the MCS reached its matured stage (Brendas 1990), which also is indicated in an examination of the hourly model output. The model continues to exhibit...
skill in producing the observed surface features (Fig. 10), including the well-known presquall mesolow, squall mesohigh, and wake low (Fujita 1963). The simulated pressure gradient across the leading convective line is of the same magnitude as the observed, while at the rear of the system the simulated wake low is fully developed, as observed, but slightly to the west of the observed position. This also indicates that the simulated system is moving eastward slightly slower than the observed one.

The success of the model simulation in reproducing many of the observed surface features of this MCS lead us to examine the model output more closely and compare the modeled fields with the observational results of Brandes (1990). This comparison focuses upon the significant mesoscale features documented in the observations and the mechanisms by which these features are formed. For example, model results indicate that the 500–300-mb thickness over the location of the presquall low increases by more than 10 m during the 3-h period from 0600 to 0900 UTC. This agrees well with the analysis of Brandes (1990) in which a 6-m thickness increase was observed for the same layer at Oklahoma City (OKC) during a 3-h period from 0530 to 0830 UTC. Both the model results and observations suggest that the presquall low is caused by the convectively induced subsidence warming in the mid-to-upper troposphere ahead of the squall lines as speculated by Hoxit et al. (1976).

One of the most distinct surface features of this MCS event is the surface mesohigh. Simulated vertical motions at different pressure levels over the mesohigh center (not shown) indicate downward motion below 750 mb and divergence below 800 mb. The shape of these profiles are similar to those reported in numerous studies of MCSs (Brandes 1990; Smull and Augustine 1993). The simulated hourly rainfall maximum of 10.9 mm at 0900 UTC is within the downdraft area (near CSM), coincident with the mesohigh center (see Figs. 10b and 11b). These results suggest that the simulated mesohigh is produced primarily by evaporative cooling in agreement with both observations (Fujita 1959) and previous modeling studies (Zhang and Fritsch 1986; Zhang et al. 1989). The simulated rainfall pattern, magnitude, and position are in good agreement with the observations (Fig. 11a), although the predicted maximum rainfall center is located to the north of the observed location.

The wake low is another prominent feature of a mature squall line (Brunck 1953; Fujita 1963; Johnson and Hamilton 1988). A warm, dry lower troposphere is a common feature of the rear portion of trailing stratiform regions in the Tropics and midlatitudes, and this structure typically is characterized by “onion-shaped” soundings (Zipser 1977). Both the observed sounding (Fig. 12a) from WWR and a simulated sounding (Fig. 12b) taken within the wake low at 0900 UTC 7 May show this well-known onion-shaped structure. A nearly dry-adiabatic lapse rate is located between 650 and 800 mb in both of these soundings. A cross section of the simulated vertical circulation and relative humidity (Fig. 13) indicates that a descending, midtropospheric rear-inflow jet brings drier air into the rear portion of the MCS. The rear inflow is produced in the model near 300 mb and descends to the surface. At low levels, local maximum warming (not shown) and drying appear near 800 mb (over point C in Fig. 13) just to the rear of the precipitation (>2 mm) region. The minimum pressure (surface wake low) occurs at this position. Thus, the intense surface wake low, which is formed at the back edge of the modeled squall system, can be explained hydrostatically as a direct result of the warm-
Fig. 8. (a) Analyzed sea level pressure (mb) at 0600 UTC 7 May using the convention of Young and Fritsch (1989). Pressure contoured every 1 mb. (b) As in (a) but for the simulated sea level pressure from the ADI run.

Fig. 9. (a) Observed hourly rainfall total (mm) from 0600 to 0700 UTC 7 May. Dark regions represent rainfall amounts that are larger than 10 mm. (b) As in (a) but for simulated hourly rainfall totals (mm) from the ADI run.
mesoscale updraft prevails. Discrepancies between the observed and modeled profiles are related, in part, to differences in the regions over which the calculations were computed. Brandes (1990) calculations are valid over the northwest portion of the mesovortex region and used only three soundings, whereas the model calculations are valid over the entire region of the mesovortex core (125 km × 150 km) (see Fig. 14).

The simulated mesovortex at the mature stage (0900 UTC) can be further compared with the vorticity budget analysis of Brandes and Ziegler (1993) based on the following vorticity equation for frictionless flow:

$$\frac{\partial \zeta}{\partial t} = -v_h \cdot \nabla \zeta - w \frac{\partial \zeta}{\partial z} - \zeta (\nabla \cdot v_h) - f(\nabla \cdot v_h)$$

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$$- k \cdot (\nabla w \times \frac{\partial w}{\partial z}) + k \cdot \left( \frac{1}{\rho} \right) \nabla \rho \times \nabla p, \quad (4.1)$$

TT S

where $\zeta \equiv k \cdot (\nabla \times v)$ is the vertical vorticity, and $v_h$ is the horizontal velocity. Terms on the right-hand side of (4.1) are the horizontal advection of relative vertical vorticity (HAD), the vertical advection of relative vertical vorticity (VAD), the convergence of relative vertical vorticity (CT), the convergence of planetary vorticity (CF), the tilting of horizontal vorticity (TT), and the solenoidal term (S). To make an accurate comparison with the calculations of Brandes and Ziegler (1993), the model sigma-level data are interpolated to height surfaces before the calculations are made. Since the purpose here is to compare with their analysis, rather than to make a thorough examination of the vorticity budget, the solenoidal term [neglected in Brandes and Ziegler (1993)] and the frictional terms (caused by parameterized clouds and turbulent mixing and by computational diffusion) will not be considered.

The simulated vertical profiles of CT, CF, and TT qualitatively compare well with the analyzed profiles at 0900 UTC up to 8 km, although their magnitudes are about four times as large as the analyzed (Fig. 16). These large magnitudes are attributed to the simulated profiles being averaged over a relatively small area (125 km × 150 km). When averaged over a larger area of 175 km × 175 km (not shown), the magnitudes are much closer to those calculated by Brandes and Ziegler (1993). The simulated profiles suggest that the mesovortex is produced by the convergence of both relative and planetary vorticity in the layer between 2 km (~800 mb) and 8 km (~350 mb) where the intruding rear inflow produces strong convergence in the mesovortex region. Above 8 km, analyzed results were not available owing to the lack of data. The simulated CF becomes negative above 8 km, where the flow is divergent, whereas the simulated CT remains positive, suggesting an anticyclonic flow in upper levels. [Note that CT is an area-averaged value, so it does not have to change sign at the level (around 8 km) where the area-averaged divergence or area-averaged relative vorticity changes sign.] The TT profile also is qualitatively comparable with the analyzed one, both profiles showing negative values through most of the troposphere and large negative values around 7–8 km.

While advection does not create new vorticity, it plays an important role in the local vorticity budget by redistributing vorticity. As shown in Fig. 17, the analyzed and simulated HAD profiles are both negative in the middle and upper levels. However, differences exist between the analyzed and the simulated VAD profiles. The analyzed VAD is negative between 5.5 km (~510 mb) and 6.5 km (~450 mb), while the simulated VAD is positive up to 11 km.

By 1100 UTC, both the simulated and the observed squall line moved into central Oklahoma and began to

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**Fig. 10.** (a) As in Fig. 8a but for 0900 UTC 7 May. The cross indicates the location of the station WWR. (b) As in Fig. 8b but for 0900 UTC 7 May. Line AB shows the orientation of cross-sectional plot in Fig. 13.
deep convection initiated by the model does not develop in the correct location initially and does not evolve in agreement with observations. In particular, the simulation fails to reproduce the initial convective development in northeastern New Mexico that later develops into the MCS in the Texas panhandle. As this simulation progresses, the results deteriorate further, as indicated by the sea level pressure and rainfall patterns that are consistently in poor agreement with observations. The simulated sea level pressure (Fig. 18) at 0900 UTC is significantly different from the analyzed (cf. Fig. 10a), and the presquall low, wake low, and

decline. The pressure gradients across the squall line had decreased, although the wake low and mesohigh persisted and moved southeastward with the MCS. At this time, the simulated presquall mesolow changed to an inverted pressure trough and the simulated rainfall amounts decrease, in agreement with observations. The simulated profiles of vertical velocity, relative vertical vorticity, and horizontal divergence in the mesovortex core region also start to decline, but the relative vertical vorticity (not shown) continues to concentrate at midlevels and the downdraft continues to strengthen and develops vertically (not shown). The simulated downdraft level is displaced upward from earlier times to 5 km (~550 mb), due to sublimation, which also is consistent with observations (Brandes and Ziegler 1993).

b. Simulations with SI

Although both the conventional data and special PRE-STORM data are used in the SI initialization, the

![Figure 12](image-url)  
Fig. 12. (a) Observed sounding from WWR at 0900 UTC 7 May plotted on a skew \( T - \log p \) diagram. Winds are in meters per second. Location of WWR indicated in Fig. 10a. (b) Model grid point (point C in Fig. 10b) sounding taken within the wake low at 0900 UTC 7 May plotted on a skew \( T - \log p \) diagram. Winds are in meters per second.
mesohigh are not well reproduced. While the observed wake low and mesohigh persist at later times, the simulated wake low and mesohigh dissipate after this time. The SI-simulated rainfall position, pattern, and amounts (Fig. 19) at 0900 UTC differ substantially from the observed (Fig. 11a). While the observed rainfall is distributed over the entire west-central Oklahoma where convection is strongest, the SI-simulated maximum rainfall (5.44 mm) is shifted westward into the Texas panhandle and is much smaller than the observed amount. Detailed comparisons (not shown) indicate that the SI simulation cannot capture even the most
was not significantly strong and the important mesoscale features were not observed well.

c. Verification of simulation results

To quantitatively evaluate the performances of the model for the two different initial conditions, verifications of the model simulations are produced. Since rainfall is one of the most sensitive and important model output parameters, and the availability of high-resolution hourly rainfall and surface observations is good, we focus upon the verification of rainfall and sea level pressure. Two approaches can be chosen to verify randomly spaced observations with higher-resolution gridded model output: 1) interpolate the model data to the

Fig. 16. (a) Mean analyzed mesovortex vertical vorticity tendencies at 0900 UTC 7 May \((10^{-16} \text{ s}^{-2})\) from Brandes and Ziegler (1993). The CT, CF, and TT stand for the convergent amplification of relative vorticity and planetary vorticity, and the tilting of horizontal vorticity, respectively. (b) Mean model-simulated mesovortex vertical vorticity tendencies at 0900 UTC 7 May \((10^{-16} \text{ s}^{-2})\). The CT, CF, and TT are as the same as in (a).

Fig. 17. (a) Analyzed mean mesovortex vertical vorticity tendencies due to horizontal advection (HAD) and vertical advection (VAD) at 0900 UTC 7 May from Brandes and Ziegler (1993). (b) Simulated mean mesovortex vertical vorticity tendencies due to horizontal advection (HAD) and vertical advection (VAD) at 0900 UTC 7 May.
observation location, and/or 2) directly use the model gridpoint data located closest to the observation point. The latter approach is the one chosen for use here, since all observation locations are within 18 km of the model grid points. Statistics from both approaches have been calculated and show very similar results.

Verification of sea level pressure is performed over the entire nested-grid domain (cf. Fig. 4), while verification of rainfall is performed only over a part of the nested-grid domain where high-resolution rainfall data from 210 stations are available from the National Climatic Data Center and PRE-STORM observations (cf. Fig. 2). Two objective measures of forecast skill are used: the threat score for precipitation and the root-mean-square (rms) error for sea level pressure. The threat score is defined by

\[ TS = \frac{FC}{FP + OP - FC} \]  

where FC is the number of stations with correct predictions of a certain threshold amount of precipitation, FP is the number of stations predicted to have a certain threshold amount of precipitation, and OP is the number of stations where precipitation of at least the threshold amount occurred.

Table 2 shows the mean threat scores calculated for the accumulated rainfall over 3-h periods for the two initialization procedures. The threat scores with the ADI suggest that precipitation forecasts have good skill for the lower precipitation thresholds in comparison with the results of Kuo et al. (1993) who assimilated precipitable water measurements into a mesoscale model. For the highest threshold (10 mm), the threat score is low, owing to the region of deep convection in the model moving slower than indicated by the observations, which introduces a phase error into the model simulation. However, the ADI produces higher threat scores than the SI for all precipitation threshold values.

Threshold scores for 6-h accumulated rainfall amounts (Table 3) typically are higher than the mean threat score for the 3-h rainfall amounts. Indeed, for the second 6-h time interval (Table 3b) the threat score for the 12-mm (0.5 in.) threshold is 0.29 for the ADI, and the threat score increases to 0.34 for the 12-h rainfall period (not shown). These values are quite good considering that this event occurred during the warm season when the precipitation threat scores for operational models typically are very low (Junker et al. 1989). In contrast, for the second 6-h time interval the threat score for the 12-mm threshold is 0.14 for the SI (Table

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Table 2. Mean threat scores averaged over three periods of 3-h accumulated rainfall valid 0600, 0900, and 1200 UTC 7 May 1985 for the two initialization procedures.

<table>
<thead>
<tr>
<th>Threshold (mm)</th>
<th>0.25</th>
<th>1.00</th>
<th>2.50</th>
<th>5.00</th>
<th>10.00</th>
</tr>
</thead>
<tbody>
<tr>
<td>ADI</td>
<td>0.48</td>
<td>0.44</td>
<td>0.34</td>
<td>0.25</td>
<td>0.09</td>
</tr>
<tr>
<td>SI</td>
<td>0.39</td>
<td>0.35</td>
<td>0.26</td>
<td>0.20</td>
<td>0.07</td>
</tr>
</tbody>
</table>

Table 3. Threat scores calculated from 6-h accumulated rainfall fields valid 0600 UTC 7 May and 1200 UTC 7 May 1985 for the two initialization procedures.

<table>
<thead>
<tr>
<th>Threshold (mm)</th>
<th>0.25</th>
<th>1.0</th>
<th>2.5</th>
<th>5.0</th>
<th>7.0</th>
<th>10.0</th>
<th>12.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>(a) 0000–0600 UTC 7 May 1985</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ADI</td>
<td>0.38</td>
<td>0.36</td>
<td>0.31</td>
<td>0.22</td>
<td>0.16</td>
<td>0.14</td>
<td>0.18</td>
</tr>
<tr>
<td>SI</td>
<td>0.34</td>
<td>0.31</td>
<td>0.24</td>
<td>0.17</td>
<td>0.13</td>
<td>0.12</td>
<td>0.08</td>
</tr>
<tr>
<td>(b) 0700–1200 UTC 7 May 1985</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ADI</td>
<td>0.67</td>
<td>0.63</td>
<td>0.51</td>
<td>0.44</td>
<td>0.42</td>
<td>0.37</td>
<td>0.29</td>
</tr>
<tr>
<td>SI</td>
<td>0.57</td>
<td>0.54</td>
<td>0.48</td>
<td>0.41</td>
<td>0.30</td>
<td>0.24</td>
<td>0.14</td>
</tr>
</tbody>
</table>
3b). The ADI-simulated rainfall obviously is better than that simulated from the SI. In addition, the heaviest 12-h precipitation (>24 mm) predicted by the ADI simulation is in western Oklahoma in the same general location where the heaviest rainfall (>24 mm) was reported. Therefore, the model prediction also provides useful information on the likely area of heavy rainfall.

The root-mean-square errors of sea level pressure for different time intervals for the ADI and the SI are shown in Table 4. The ADI simulation performs well, with rms errors roughly the same as those reported in other mesoscale model simulations (Anthes et al. 1989). The rms errors from the SI simulation become larger than those from the ADI simulation after 3 h when the MCS moved into the PRE-STORM observing network. In both simulations, the rms errors in sea level pressure are small, but the decreasing rms errors from the ADI simulation with time illustrate the improvement in the model simulation from the more detailed model initial condition produced by the ADI technique.

5. Summary and conclusions

Two different types of initialization schemes are used in the PSU—NCAR Mesoscale Model to simulate the observed structure and evolution of the 7 May 1985 MCS when the large-scale forcing is not strong and the initial convective development occurs in a sparse data area. The simulation that uses an initial condition created by dynamic (nudging) initialization followed by an additional step of interpolation, incorporating both the standard and special PRE-STORM upper-air and surface data, can reproduce the observed evolution and many of the mesoscale features associated with the MCS. These include 1) the generation of the important initial regions of deep convection; 2) the approximately correct timing of the development, intensification, and decay of the squall line; 3) the generation of a squall-induced mesohigh, wake low, and presquall low; 4) the simulation of onion-shaped soundings along the back edge of the trailing stratiform region; 5) the rear inflow jet; 6) the development, maintenance, and intensification of a mesovortex associated with the squall line; and 7) the dominant mechanism sustaining the mesovortex as compared with the observational study of Brandes and Ziegler (1993). Although the simulated MCS moves slightly slower than the observed, the model reproduces the hourly observational rainfall patterns quite well as indicated by the relatively high threat scores.

In the ADI scheme, the additional step of interpolation (after the 12-h nudging simulation) is crucial, since it significantly improves the representation of the initial momentum and moisture fields in the model. As shown in Zheng (1993), when nudging alone is used to produce the model initial condition at 0000 UTC 7 May, the subsequent simulation degrades substantially, even though the high-resolution PRE-STORM data are used in the nudging process. When the PRE-STORM data are not used during the assimilation period, then the simulation degrades even further. These results illustrate that nudging alone cannot effectively absorb all the information contained in the PRE-STORM data, owing to the gradual correction it applies to the model equations.

A static initialization scheme also is tested, but it is not found to be very successful. Most of the observed mesoscale features associated with the MCS cannot be reproduced by the SI simulation. The simulated hourly rainfall patterns are significantly different from observations, and the threat scores are lower and the rms errors are larger than those of the ADI simulation. The failure of the SI simulation may be partially attributed to the lack of observations at mesoscale resolution. Although the PRE-STORM data are used, the initial convective development occurred to the west of the PRE-STORM observing network and, thus, the nearby pre-convective environment is not sampled (although this environment apparently can be "retrieved" indirectly by the ADI to some degree of accuracy).

The results of this study, as well as previous studies of Fritsch and Chappell (1981) and Stensrud and Fritsch (1994b), suggest that resolving certain mesoscale details within the model initial condition often is crucial for a successful prediction of convective activity. Thus, the present observational network does not have sufficient horizontal resolution for the accurate prediction of MCS development and evolution. This poses a serious challenge to mesoscale data assimilation and initialization, especially when large-scale forcing for upward vertical motion is relatively weak within the region of convective development. In these situations, the incorporation of the poorly sampled mesoscale environmental features may be crucial to the production of a useful model forecast. It will be interesting to explore how much this problem can be reduced by using a more advanced data assimilation technique such as the adjoint method or Kalman filter. The ADI approach presented in this study should have room for further improvement in a direction toward optimal interpolation or even the more expensive Kalman filter data-assimilation scheme. While these complications are beyond the scope of this paper, they are worth consideration in future studies even though their computational costs will be very high with present computer resources.

<table>
<thead>
<tr>
<th>Forecast time</th>
<th>3 h</th>
<th>6 h</th>
<th>9 h</th>
<th>12 h</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>ADI</td>
<td>1.64</td>
<td>2.27</td>
<td>1.87</td>
<td>1.64</td>
<td>1.86</td>
</tr>
<tr>
<td>SI</td>
<td>1.46</td>
<td>2.83</td>
<td>2.12</td>
<td>1.95</td>
<td>2.09</td>
</tr>
</tbody>
</table>
Acknowledgments. This study was motivated and influenced by numerous discussions with Edward A. Brandes, Ying-Hwa Kuo, Da-Lin Zhang, J. Michael Fritsch, and Robert A. Maddox. Sue Chen, Yong-Run Guo, Wei Wang, Jian-Wen Bao, and Kun Gao provided much needed assistance and guidance in running the PSU–NCAR model and their help is greatly appreciated. PRE-STORM data were kindly provided by Jose Meitin, Edward A. Brandes, and Conrad L. Ziegler. Precipitation data were supplied by the Data Support Section at NCAR. Sue Weygangt is thanked for her assistance in producing several of the figures. Comments and suggestions provided by Brian Fielder, Peter Lamb, and two anonymous reviewers improved the presentation of the results. The numerical simulations were performed on the CRAY Y-MP at NCAR, which is sponsored by the National Science Foundation. Yong Zheng and Qin Xu acknowledge the financial support provided by the NOAA Contract NA37RJO203, the DOE-ARM Program through Battelle PNL Contract 144880-A-Q1, and the NSF Grants ATM-9113906 and ATM-9120009.

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