Use of Four-Dimensional Data Assimilation by Newtonian Relaxation and Latent-Heat Forcing to Improve a Mesoscale-Model Precipitation Forecast: A Case Study

WEI WANG AND THOMAS T. WARNER

Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

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ABSTRACT

The Penn State/NCAR mesoscale model has been used in a study of special static- and dynamic-initialization techniques that improve a very-short-range forecast of the heavy convective rainfall that occurred in Texas, Oklahoma and Kansas during 9-10 May 1979, the SESAME IV study period. In this study, the model is initialized during the precipitation event. Two types of four-dimensional data assimilation (FDDA) procedures are used in the dynamic-initialization experiments in order to incorporate data during a 12-hour preforecast period. With the first type, FDDA by Newtonian relaxation is used to incorporate sounding data during the preforecast period. With the second FDDA procedure, radar-based precipitation-rate estimates and hourly raingage data are used to define a three-dimensional latent-heating rate field that contributes to the diabatic heating term in the model's thermodynamic equation during the preforecast period. This latent-heating specification procedure is also employed in static-initialization experiments, where the observed rainfall rate and radar echo pattern near the initial time of the forecast are used to infer a latent-heating rate that is specified in the mesoscale model's thermodynamic equation during the early part of the actual forecast. The precipitation forecasts from these static- and dynamic-initialization experiments are compared with the forecast produced when only operational radiosonde data are used in a conventional static initialization.

The conventional (control) initialization procedure that used only operational radiosonde data produced a precipitation prediction for the first three to four hours of the forecast period that would have been inadequate in an operational setting. Whereas at the initial time of the forecast there was substantial convective precipitation observed in a band near the edge of an elevated mixed layer, the model did not initiate the heavy rainfall until about the fourth hour of the forecast.

The use of the experimental static initialization with prescribed latent heating during the first forecast hour produced greatly improved rainfall rates during the first three to four hours. The success of the technique was shown to be not especially sensitive to moderate variations in the duration, intensity and vertical distribution of the imposed heating. Applications of the Newtonian-relaxation procedure during the preforecast period, that relaxed the model solution toward the initial large-scale analysis, produced a better precipitation forecast than did the control, with a maximum in approximately the correct position, but the intensities were too small. Combined use of either the preforecast or in-forecast latent-heat forcing with the Newtonian relaxation produced an improved forecast of rainfall intensity compared to use of the Newtonian relaxation alone. Even though both the experimental static- and dynamic-initialization procedures produced considerably improved very-short-range precipitation forecasts, compared to the control, the experimental static-initialization procedure that used latent-heat forcing during the first forecast hour did slightly better for this case.

1. Introduction

Numerous real-data case studies have been performed with mesoscale meteorological models, and in some cases the models have demonstrated an encouraging amount of skill in predicting convective-scale precipitation events (Zhang and Fritsch 1986a; Lakhtakia and Warner 1987). However, there are at least two reasons why our optimism should be cautious about how straightforwardly these successes can be translated into improved operational very-short-range (0-12 h) and short-range (12-48 h) quantitative precipitation forecasts (QPF). First of all, the model initial conditions generally receive a great deal of special attention in these studies. Either special data are utilized to supplement the operational dataset, or very careful subjective analyses are laboriously performed and then digitized for use by the model (Zhang and Fritsch 1986a,b). In those cases where good mesoscale precipitation forecasts have been produced with the use of only smooth synoptic-scale data, the model dynamics and physics-parameterizations have had sufficient time to develop mesoscale structures before the onset of precipitation. However, operational forecasts may need to be initialized during a precipitation event. Second, even though these experimental mesoscale QPFs are very encouraging by historical standards, they often still do not verify well objectively. That is, the model may predict the convective event during the correct
12-h period and in the correct state, but it generally does not provide the more-exact timing and position information that would be needed by operational meteorologists and by river forecasters concerned about precipitation rates in different watersheds.

It is therefore reasonable to inquire whether there exist objective methods of enhancing the model initial conditions with operationally available data in order to provide mesoscale very-short-range QPFs of sufficient quality to serve as operational products. Clearly the final answer to such a question must be based on experience gained from modeling a large number of cases. However, studies of specific techniques using fewer cases are valuable because they can expose the existence of numerical problems and demonstrate the dynamic response of a model atmosphere to certain data types.

Numerous techniques have been developed that employ routinely available mesoscale data to define the initial conditions for numerical mesoscale QPFs. Both static initialization (SI) and dynamic-initialization (DI) procedures are utilized. For example, Tarbell et al. (1981), Salmon and Warner (1986) and Wolcott and Warner (1981) use a diabatic omega equation, while Benoit and Roch (1987) employ a normal-mode initialization to incorporate rainfall data using SI procedures. Ninomiya et al. (1987), Ninomiya and Kurihara (1987), and Fiorino and Warner (1981) employ DI procedures that use inferred latent heating rates during a preforecast period or during the early part of the actual forecast.

With an objective similar to that of these earlier studies, this study investigates the use of special DI and SI techniques to improve a very-short-range QPF of the heavy convective rainfall that occurred in Texas, Oklahoma and Kansas during 9–10 May 1979, the SESAME IV study period. In this case, the model is initialized after the onset of the precipitation event. Two types of four-dimensional data assimilation (FDDA) procedures are used in the DI experiments to incorporate data during a 12-h preforecast period. With the first type, FDDA by Newtonian relaxation (Hoke and Anthes 1976; Ramamurthy and Carr 1987; Stauffer et al. 1985) is used to incorporate sounding data during the preforecast period. With the second FDDA procedure, radar-based precipitation-rate estimates and hourly raingage data are used to define a three-dimensional latent-heating rate field that contributes to the diabatic heating term in the model's thermodynamic equation during the preforecast period. During the preforecast assimilation period, the large-scale solution of the model is able to develop a dynamic balance while the local forcing effects and nonlinear processes contribute to the development of a mesoscale component to the solution. In principle, the initial conditions will then be in better dynamic balance and contain more mesoscale structure than if they had been based on a static initialization. This latent-heating specification procedure is also employed in SI experiments, where the observed rainfall rate near the initial time of the forecast is used to infer a latent-heating rate that is specified in the mesoscale model's thermodynamic equation during the early part of the actual forecast. This latter technique was used recently with success by Ninomiya and Kurihara (1987) for a convective system in a Baiu frontal zone. The QPFs from these SI and DI experiments are compared with the QPF produced when only operational radiosonde data are used in a conventional SI.

The mesoscale modeling system used in this study is described in section 2 and the meteorological aspects of the case study period are discussed in section 3. The experimental design is described in section 4, while section 5 discusses the results in terms of the sensitivity of the QPF to the initialization procedures. The summary and conclusions are found in section 6.

2. The numerical modeling system

a. General description of the model

The numerical model employed in this study is the Penn State/NCAR mesoscale model described by Anthes et al. (1987). It is a three-dimensional, hydrostatic, primitive-equation model with the \( \sigma \) terrain-following vertical coordinate, where \( \sigma = (p - p_0)/(p_s - p_0) \), \( p \) is the pressure at the top of the model and \( p_0 \) is the surface pressure. Physical parameterizations include a subgrid-scale precipitation parameterization to be discussed in the next subsection, a grid-scale (resolvable) precipitation parameterization, and a medium-resolution parameterization of the planetary boundary layer and surface fluxes, as described by Blackadar (1976, 1978) and Zhang and Anthes (1982).

The SI experiments employ a vertical-mode initialization procedure (Errico 1986) to minimize the intensity of unrealistic gravity-wave activity during the simulations. Three vertical modes are initialized. Details of the experimental DI procedures will be provided in section 4. The lateral boundary conditions are defined using the relaxation technique described by Davies (1976) and employ sounding observations available every 6 hours during the SESAME-IV period.

b. Case-specific characteristics of the model

The specific characteristics of the model as it is applied in this study are provided in Table 1. Figure 1 shows the boundaries of the computation domain and the objective analysis domain, as well as the locations of the standard radiosonde data used in the initializations.

Because this study focuses on the quality of the precipitation forecast, additional information is provided about the specific characteristics of the precipitation parameterization. Grid-scale precipitation is produced when the relative humidity exceeds 100%, and the re-
Table 1. Case-specific model description.

<table>
<thead>
<tr>
<th>Characteristics</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>top of model</td>
<td>100 mb</td>
</tr>
<tr>
<td>number of computational levels</td>
<td>15</td>
</tr>
<tr>
<td>computational sigma levels</td>
<td>0.9950, 0.9850, 0.9650, 0.9300, 0.8850, 0.8325, 0.7725, 0.7050, 0.6300, 0.5450, 0.4500, 0.3500, 0.2500, 0.1500, 0.0500</td>
</tr>
<tr>
<td>grid increment</td>
<td>30 km</td>
</tr>
<tr>
<td>horizontal mesh size</td>
<td>49 × 49</td>
</tr>
<tr>
<td>grid center</td>
<td>34°N, −99°W</td>
</tr>
</tbody>
</table>

resulting latent heat is included in the thermodynamic equation at the appropriate model levels. The subgrid-scale precipitation parameterization is based on the procedures described by Kuo (1974) and Anthes (1977), with the specific properties being discussed in Anthes et al. (1987). Specifically, when the model sounding is convectively unstable and the vertically integrated horizontal moisture convergence exceeds a pre-defined critical value, the parameterization scheme is triggered. This grid–column moisture convergence is partitioned into two components; one that represents subgrid-scale precipitation and one that represents a moistening of the column according to a calculated profile. The vertical distributions of latent heat and moisture that result from the moist convection are prescribed by two parabolic profiles that have time- and space-dependent upper and lower bounds. These parabolic profiles are based on the diagnostic heat and moisture-budget study of Kuo and Anthes (1984). The upper and lower limits of the profile are calculated by diagnosing the cloud top and cloud base from the sounding.

The ground-surface temperature is calculated using an energy equation described by Anthes et al. (1987), which requires the specification of the surface albedo, emissivity, roughness, and wetness in addition to the soil thermal inertia. Lanicci et al. (1987) and Lakhtakia and Warner (1987) show that the correct specification of the spatial distribution of these parameters, especially soil wetness, is essential to the production of a reasonable simulation of the mesoscale structures in this case. This study uses the surface properties described in Lanicci et al. (1987). The effects of clouds on the short- and long-wave radiation at the surface are also parameterized.

c. Data analysis

Four types of meteorological data are employed in the static and dynamic initializations; conventional surface data and radiosonde data, hourly rain amounts from the operational gage network, and almost-continuous radar data from Garden City, Kansas; Oklahoma City, Oklahoma; and Amarillo, Texas.

Gridded fields of the predicted variables at 6-h intervals between 1200 UTC 9 May and 1200 UTC 10 May are obtained by analyzing the surface data and the SESAME radiosonde data with the successive-correction type of objective analysis procedure described by Benjamin and Seaman (1985). The operational analysis of the National Meteorological Center is used as a first-guess field.

The gridded fields of hourly precipitation amount, used for the initialization and QPF verification, could not generally be determined entirely from the gage network because of the small spatial scale of the observed rainfall compared to the gage density. Thus, radar images that reflect the area coverage of the hydrometeors in the atmosphere and also imply rainfall intensity in terms of the reflectivity levels, are used to provide additional insight into the probable distribution of precipitation at the surface during each hour. The reflectivity data are used to determine instantaneous rain rates based on the standard National Weather Service relationship for convective-rainfall observations with the WSR-57 radar. Probably because converting these instantaneous radar estimates of rain rate to hourly totals is not straightforward, there seems to be a general trend for the radar-inferred hourly totals to be greater than those observed by the gage network. This apparent overestimation by the radar or by the hourly averaging procedure is compensated for by scaling the radar-based hourly rain-rate field so that the radar estimates better match the gage data, which are admittedly sparse. The spatial distribution of rainfall implied by the radar is, however, generally used directly without modification. The hourly precipitation fields thus derived are then digitized to define the gridded fields. Because of ques-

![Fig. 1. The objective analysis domain (A) and the computational domain (B). A* indicates the location of a NWS radiosonde observation. CS1 and CS2 show the orientation of cross-section plots.](image-url)
tions about the grid-scale representativeness of the gage data and limitations in the use of radar data to imply the distribution of precipitation at the surface, there is naturally some uncertainty about the exact spatial distribution and the magnitude of the fields. Time continuity of patterns is used when possible. Nevertheless, these fields are used in the initialization since they do result from a reasonable synthesis of operational data.

3. The meteorological case

The SESAME IV case of 9–10 May 1979 is used as the basis for this study. Because much has been written about this case (July and Turner 1980; Ogura et al. 1982; Carlson et al. 1983; Lakhtakia and Warner 1987), the synoptic-scale situation will not be described here. However, the observed precipitation will be discussed since it is the quality of the QPF that is the focus of this research.

The severe weather started shortly after 2100 UTC 9 May in the northeastern Texas panhandle and in northwestern Oklahoma. A number of tornadoes, as well as hail and severe thunderstorms, were observed in this area from 2130 UTC 9 May to 0430 UTC 10 May (see Fig. 49 in July and Turner 1980). Heavy rains, associated with the intense convection at the northwestern edge of an elevated mixed layer, were confined to a small area in northwestern Oklahoma and in the extreme northeastern Texas panhandle. Very high radar-echo tops of 13 000–18 000 m were measured during the period at Amarillo, Texas and Oklahoma City, Oklahoma. Various mechanisms have been proposed to explain the initiation of the flow of buoyant air to the northwest, away from the capping inversion. For example, Ogura et al. (1982) and Lakhtakia and Warner (1987) discuss the possibility of a circulation driven by differential surface heating (an inland sea breeze), while Benjamin and Carlson (1986) evaluate the importance of a low-level transverse circulation in the entrance region of a midlevel jet streak.

In spite of the occurrence of severe weather as early as 2130 UTC 9 May, substantial rainfall was not observed until about 0000 UTC 10 May. Fort Supply Dam, northeast of Gage, Oklahoma, reported the maximum rainfall of 16.8 cm for the region, in the 6-h period from 0000 to 0600 UTC 10 May. More than two-thirds of that amount fell in the first 3 h of the period. The 6-h precipitation observed by the hourly gage network is shown in Fig. 2a for the period 0000–0600 UTC 10 May. Using more-numerous observations, the NMC Heavy Precipitation Branch produced the analysis in Fig. 2b of the total 24-h precipitation from 1200 UTC 9 May to 1200 UTC 10 May. The rainfall intensity in the region began to diminish about 0530 UTC 10 May. By 0600 UTC 10 May, a stationary front to the northwest had become a cold front and was advancing eastward. The northern precipitation area shown in Fig. 2 was located behind this front. However, the heavier precipitation to the southeast, associated with the edge of the elevated mixed layer and perhaps the front, is the primary focus of this research.

4. Experimental design

Two series of experiments are conducted, with 0000 UTC 10 May always defined as the initial time of the forecast. In the first series, the model is initialized at 0000 UTC 10 May and is integrated for 12 h. This series includes the SI experiments and the control forecast. In the second series, consisting of the DI experiments, a preforecast period of 12 h is used, during which data are assimilated before the start of the actual forecast at 0000 UTC 10 May. Figure 3 shows schematically how the data are assimilated during the integration cycle, while Table 2 summarizes the design

![Fig. 2. Observed 6-hour (a) and 24-hour (b) precipitation for the periods beginning at 0000 UTC 10 May 1979 and 1200 UTC 9 May 1979, respectively. The 6-hour analysis shows the data from the hourly gage measurements, whereas the 24-hour analysis was produced by the NMC Heavy Precipitation Branch based on more numerous measurements from cooperative observers.]
of each experiment. A brief description of each experiment is found in the following paragraphs.

a. Control forecast

The control forecast (CTL) is a 12-h simulation that is statically initialized at 0000 UTC 10 May with surface and radiosonde data. It is this forecast that is used as a benchmark against which the forecasts with experimental initializations are compared because it employs a relatively conventional initialization technique and data set. The numerical characteristics and physical-process parameterizations of the model are not exceptional compared to some operational models, so this control forecast can be interpreted as characteristic of what can be currently provided by some operational forecast centers. Note that the use of observed data for lateral boundary conditions (unlike operational models) should not strongly influence the model solution during the very-short integration times used in this study.

b. Static-initialization experiments

In this series of experiments, hourly raingage data for the period 2300 UTC 9 May–0000 UTC 10 May and radar data for times near 0000 UTC are used to construct an estimate of the precipitation-rate field that is appropriate for the period 0000–0100 UTC 10 May, the first hour after the initial time of the forecast. This two-dimensional field is then used to produce an estimate of the three-dimensional latent-heating field, which entirely determines the latent-heating contribution to the diabatic-heating term in the model’s thermodynamic equation for a specified period of time after the static initialization. This procedure is a way of employing one of the few sources of mesoscale data currently available (i.e., the radar and raingage data) and will force vertical circulations and associated divergence patterns that will hopefully induce precipitation with a realistic spatial and temporal distribution. The practical implementation of this technique, however, requires that the following problems be addressed or at least recognized.

1) The procedure is only useful when the model is initialized during a precipitation event.
2) It is impossible to exactly determine the vertical distribution of the latent-heating function from the raingage and radar data.
3) The duration of the latent-heat forcing must be relatively short because a temporally constant function is being used.
4) Inaccurate large-scale fields in the initial conditions, or some types of model-produced errors on the large scale, may prevent the forced mesoscale circulations from being retained after the assimilation period.
5) An acceptably short (see item 3) assimilation time period may be insufficient to allow the model dynamics to produce mesoscale circulations of realistic amplitude.
6) Significant uncertainties can exist in the estimates of the hourly rain-rate averages for a grid box because of difficulties analyzing convective rainfall patterns from sparse data.

Regarding the first point, it is worth noting that forecasts initialized prior to the onset of precipitation may produce more-accurate QPFs than forecasts initialized during the event because, in the former case, the model dynamics may have sufficient time to develop the synoptic-scale and mesoscale vertical motion patterns. Thus, this initialization procedure can be utilized in those situations where there is the most need for extraordinary methods. In the study described here, the major precipitation event occurred after the ini-

<table>
<thead>
<tr>
<th>Simulation code</th>
<th>Preforecast latent-heat forcing (21–00 UTC)</th>
<th>During-forecast latent-heat forcing</th>
<th>Newtonian relaxation of V and T (coefficient)</th>
<th>Comment (Fig. 3 reference)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>conventional initialization</td>
</tr>
<tr>
<td>S1</td>
<td>No</td>
<td>0–1 h</td>
<td>No</td>
<td>LH2</td>
</tr>
<tr>
<td>S2</td>
<td>No</td>
<td>0–1 h (half strength)</td>
<td>No</td>
<td>LH2</td>
</tr>
<tr>
<td>S3</td>
<td>No</td>
<td>0–1 h (modified vertical distribution)</td>
<td>No</td>
<td>LH2</td>
</tr>
<tr>
<td>S4</td>
<td>No</td>
<td>0–2 h</td>
<td>No</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>No</td>
<td>No</td>
<td>3 x 10^-4</td>
<td>NR</td>
</tr>
<tr>
<td>D2</td>
<td>No</td>
<td>No</td>
<td>5 x 10^-4</td>
<td>NR</td>
</tr>
<tr>
<td>D3</td>
<td>Yes</td>
<td>No</td>
<td>3 x 10^-4</td>
<td>NR, LH1</td>
</tr>
<tr>
<td>D4</td>
<td>Yes</td>
<td>No</td>
<td>No</td>
<td>LH1</td>
</tr>
<tr>
<td>D5</td>
<td>No</td>
<td>0–1 h</td>
<td>3 x 10^-4</td>
<td>NR, LH2</td>
</tr>
</tbody>
</table>
tialization, but a significant precipitation rate existed at the initial time.

The second point, about the uncertainty in estimating the appropriate vertical distribution function for the latent-heat forcing, is an important one. The problem is especially difficult because the latent-heat forcing profile that will produce the best forecast after the forcing ceases is not just related to the actual distribution in the atmosphere. Rather, the forcing profile also should be consistent with the distribution that will be determined internally by the model’s precipitation parameterizations during the forecast after the latent-heat forcing is terminated. Ideally of course, the model-determined distribution should be consistent with that in the real atmosphere. In this study, an estimated distribution function is used during the forcing period. This parabolic profile with bounds at 100 and 850 mb and a maximum at 400 mb was determined from previous experience with this parameterization and is used during the forcing period in experiment S1. Checks during the forecast show it to be generally consistent with the space- and time-dependent distribution determined by the model. To test the sensitivity of the success of this procedure to the estimate of the vertical heating function used during the forcing period, another experiment, S3, uses a parabolic distribution function with the same upper and lower bounds as S1, but the maximum is at 600 mb. The precipitation parameterizations in the model, however, are kept the same for S1 and S3.

The duration of the latent-heat forcing period in the SI experiments is arbitrarily chosen as 1 h for all except S4. In S4, a 2-h period of latent-heat forcing with the same heating rate as in S1 is used to determine the sensitivity of the QPF to the duration of the forcing. It is felt that a period longer than 1 h is not generally reasonable because precipitation patterns are often highly variable in time. On the other hand, too short a period will not allow sufficient time for vertical motion to develop in order that precipitation be maintained after the forcing period.

It was noted above that the success of this procedure will no doubt depend on whether the large-scale initial conditions and forecast represent an environment conducive to the development of the precipitation system. In this case, the initial conditions reflected a reasonably accurate elevated mixed layer in terms of the position of its northwest edge where the precipitation developed and in terms of the intensity of the inversion below the elevated mixed layer. In other cases where the conventional data misrepresent the large-scale environment or forcing, an application of this technique may be less successful.

Obviously there is also some uncertainty in using irregularly spaced measurements of a potentially highly variable field such as precipitation in order to estimate grid-box averages that are converted to the latent-heating rate. Thus, in order to test how sensitive the success of the procedure is to the accuracy of the amplitude of the heating function, experiment S2 is performed in which the latent-heat forcing at each grid point is divided by two. The same vertical distribution function is used as in S1.

c. Dynamic-initialization experiments

Two techniques are used to incorporate meteorological information during the 12-h preforecast period from 1200 UTC 9 May to 0000 UTC 10 May. One involves specification of the latent-heating function in a way identical to that described in the last subsection, but the forcing is applied only during the preforecast period and is based on observed hourly precipitation during this period. No forcing is applied after the start of the forecast. Note that problems 2, 4 and 6 from the list that applied to the use of latent-heat forcing in the SI procedure, are also relevant to this DI application. That is, limitations also exist here in terms of estimating the vertical distribution of the latent heating, the importance of defining a reasonable large-scale meteorological solution, and analyzing the rain rates. Figure 4 shows the hourly precipitation-rate analyses for the last 3 h of the preforecast period. From each of these fields, the three-dimensional latent-heating function is defined, and applied for the appropriate one-hour period during the preforecast integration. The other DI technique uses the Newtonian-relaxation procedure described by Hoke and Anthes (1976). In this procedure, one or more of the time-dependent variables are relaxed toward analyses valid at the end of the preforecast period. An artificial tendency term in the forecast equation for a variable \( \alpha \), has the form

\[
\frac{\partial \alpha}{\partial t} = F + G_\alpha W(x, y, \sigma, t) \cdot (\alpha_0 - \alpha),
\]

where \( F \) represents all the model’s physical processes, \( G_\alpha \) is a positive relaxation coefficient that determines the relative weight of the relaxation term to \( F \), \( W \) is a four-dimensional weighting function, and \( \alpha_0 \) is a gridded field of \( \alpha \) obtained from an objective analysis of observations. For these experiments, both the wind and temperature are relaxed toward the analyses at 0000 UTC 10 May, the end of the preforecast integration period. The weighting function \( W \) in the relaxation term is zero from 1200–1800 UTC, increases linearly from zero to one during the period 1800–2100 UTC and remains at one from 2100 to 0000 UTC. For this application there is no spatial variation of the weighting function. It should be kept in mind that there are many different specific techniques for implementing a Newtonian-relaxation procedure. For example, the model solution can be relaxed toward observations rather than toward an analysis, different combinations of predicted variables can be relaxed, different temporal weighting functions can be used, etc. Therefore, even though a
reasonable procedure is employed in this study, there are many alternatives.

The first DI experiment (D1) employs a relaxation coefficient of $3 \times 10^{-4}$ while the second (D2) uses a coefficient of $5 \times 10^{-4}$. In the third experiment (D3) the Newtonian-relaxation of wind and temperature with the coefficient of $3 \times 10^{-4}$ is used in conjunction with the latent-heat forcing that is specified during the 3 h following 2100 UTC (see Fig. 3). Regarding the latter experiment, it is worth pointing out that the simultaneous use of the two techniques does not represent a duplicate, and therefore erroneously large, forcing of the same component of the solution. Only synoptic-scale wind- and temperature-field features are resolved in the analyses used in the relaxation terms whereas the latent-heat forcing in this case is on the mesoscale. Even if the mesoscale fields are available for use in the relaxation procedure, the fact that the magnitude of the relaxation term is proportional to the difference between the model produced field and the field toward which the solution is being relaxed, means that the effect of the relaxation term will diminish as the latent-heat forcing begins to produce correct mesoscale structures in the model solution. A fourth DI experiment (D4) employs the same latent-heat forcing as in D3, but no Newtonian relaxation is used. That is, the first 9 h of the preforecast period contain no forcing. This experiment is useful in that it illustrates the influence of preforecast latent-heat forcing when the large-scale error growth has not been controlled by Newtonian relaxation. In order to facilitate comparison of the experiments with latent heat-forcing during the preforecast period (e.g. D3) and during the forecast itself (e.g. S1), experiment D5 is performed. In D5, one hour of latent-heat forcing is used during the first forecast hour, after the Newtonian relaxation is applied during the preforecast period. Essentially D5 is a combination of the procedures used in S1 and D1 and will be used to determine whether the S1 procedure benefits from use of a better balanced initial state as well as one with some mesoscale structure. Alternatively, D5 is comparable to D3 except that the latent-heat forcing is applied during the first hour of the forecast rather than during the preforecast period.

5. Results

Because this study involves the testing of techniques for improvement of the QPF, the primary measure of forecast skill will be the hourly precipitation fields. In view of the subjective way in which the radar and rain-gage data were combined in order to produce the verification fields, there is concern that objective measures of skill such as threat scores or correlation matrices (Anthes 1983) cannot be relied on to compare the observed precipitation with the control QPF and the experimental QPFs. Therefore, even though objective scores will be provided, subjective comparisons will be used also.

a. The control simulation

As would be expected with any mesoscale numerical model simulation that is statically initialized with synoptic-scale data, the control experiment does not generate the appropriate mesoscale circulation to support the precipitation until about the fourth hour of the forecast. Figures 5a and 5b show that the QPF for the first three forecast hours (0000–0300 UTC) of CTL is totally inadequate. Even though some precipitation is predicted near the front in Kansas, none is produced near the edge of the elevated mixed layer in Oklahoma. Figures 6 and 7 show the observed hourly precipitation and the predicted hourly precipitation from CTL, respectively. By the fourth hour, CTL has developed a fairly realistic intensity to the precipitation near where the maximum is located in northwestern Oklahoma. However, the maximum is predicted to be about 70 km too far to the west. Even though the predicted pattern does develop a reasonable northeast–southwest orientation by the fifth hour, the predicted rainfall does not extend into the Texas panhandle, as observed es-
FIG. 5. Observed and forecast 3-h precipitation (cm) for 0000–0300 UTC 10 May 1979: observations (a), CTL (b), SI (c), D1 (d), D3 (e) and D5 (f).

FIG. 6. Observed hourly precipitation (cm) for the periods 0000–0100 UTC (a), 0100–0200 UTC (b), 0200–0300 UTC (c), 0300–0400 UTC (d), 0400–0500 UTC (e), and 0500–0600 UTC (f) 10 May 1979.
especially during forecast hours 4 and 5. Both the observed and predicted patterns translate very little during the six hours. The observed band of precipitation moves slightly to the east during hours 5 and 6, whereas the predicted pattern moves to the northeast a short distance. Figures 8a and 8b show the CTL forecast precipitation and the observed precipitation for the 6-h period 0000–0600 UTC. Certainly the CTL forecast is able to reproduce a significant precipitation event in approximately the right location. The 6-h totals are, however, diminished because of the poor forecast from 0000–0300 UTC. For example, the area covered by the 5 cm isohyett is observed to be about three times greater than it is in the forecast. Figure 9b shows the 12-h total precipitation predicted by CTL, which can be compared with Fig. 9a (1200 UTC 9 May–1200 UTC 10 May accumulation) because of the relatively small amounts that accumulated prior to 0000 UTC. The errors in this 12-h prediction are basically of the same nature as those in the 6-h prediction from 0000 UTC—displacement of the heaviest precipitation too far to the north.

Table 3 shows three types of precipitation verification scores for the 3-h and 6-h totals for experiments CTL, S1, D1, D3 and D5. All the scores are calculated for the lid-edge precipitation area only, except for the 1-cm threshold in the 6-h totals where the frontal and lid-edge precipitation areas are contiguous. All the scores are related to the model's skill in predicting the area of precipitation amounts above a specified threshold. The threat score is the most challenging measure of skill because of the penalizing effect of predicting precipitation outside of the observed area. The bias score reflects only the tendency for the model to overpredict or underpredict the area of the precipitation above a certain threshold, there being no verification of position. The skill score (as defined here) reflects the percentage of area with observed precipitation above a threshold, for which the model also predicts precipitation above the threshold. There is no penalty for overforecasting the area, as with the threat score. A meaningful comparison of these scores with those from other simulations is difficult because of the special nature of this study—i.e., a very short-range forecast, a relatively fine mesh, and a single small-scale convective event. Nevertheless, a few statistics will be noted for the National Weather Service's NGM and for other simulations with the Penn State/NCAR model. Two years of NGM threat scores, reported by NMC (1988) for 1986 and 1987, provide examples of the operational model performance. For this period, the monthly-average NGM threat scores vary between about 0.04 and 0.40 for the 0.5 inch threshold for the 24- to 48-h forecast period. For the 1.0 inch threshold for the 12- to 36-h forecast period, the NGM threat scores range between about 0.03 and 0.30. After the implementation

![Fig. 7. As in Fig. 6 but for forecast hourly precipitation (cm) from CTL.](image-url)
Fig. 8. Observed and forecast 6-h precipitation (cm) for 0000–0600 UTC 10 May 1979: observations (a), CTL (b), S1 (c), D1 (d), D3 (e) and D5 (f).

Fig. 9. Observed 24-hour precipitation (cm) for 1200 UTC 9 May–1200 UTC 10 May and forecast 12-hour precipitation (cm) for 0000–1200 UTC 10 May 1979: observation (a), CTL (b), S1 (c) and D1 (d).
of enhanced physics in the NGM in July 1986, the 0.5
inch threat scores range between 0.30 and 0.40 for the
1986/87 winter months and average near 0.18 for the
summer 1987. The 1.0 inch scores are not significantly
different. For April and May, the period of this study,
the 1987 NGM scores for the 0.5 inch threshold are
0.30 and 0.17, respectively. For the 1.0 inch threshold,
the scores are 0.20 and 0.10, respectively. The May
scores are typical of those for the remaining warm-
season months dominated by convective events. Pre-
cipitation-forecast skill scores for various versions of
the Penn State/NCAR model are reported by Anthes
et al. (1988). For 72-h forecasts of four summertime
cases, the 1.27 cm threat scores range from about 0.42
to 0.70 for the model version with physics most similar
to the one used in this study. The grid increment is 80
km. The average 24-h threat score for the 12 cases
studied, including wintertime cases, is 0.54 for the 0.254
cm threshold and 0.31 for the 2.54 cm threshold.

For the CTL simulation in this study, none of the
3-h scores show any skill because it is assumed that the
predicted precipitation in the northern part of the
domain is not associated with the event being verified
near the lid edge. The 6-h scores are better, reflecting
the fact that the model does develop some skill late in
the period. The 1-cm threat score is low because of
the large area in west Kansas for which greater than 1-cm
is erroneously forecast (bias score = 1.82). The 3-cm
threat score is reasonably good, as are the bias and skill
scores for this threshold. The larger thresholds (5 and
7 cm) show poorer scores because the finer-scale of the
areas of heavy rainfall causes modest position errors
to have a greater influence on the scores. Also, the
model shows a dry bias for the larger amounts, at least
partly because the areas to be verified are not much
larger than the grid-box area in some cases.

In summary, CTL is able to produce a reasonably
good mesoscale very-short-range precipitation forecast
in spite of the fact that only synoptic-scale data are
used in the initialization. A precipitation event of major
proportions is predicted in approximately the right po-
ton, with approximately the correct orientation and
aspect ratio, and during the correct 12-h period. How-
ever, the shortcomings in the CTL forecast could have
been of possibly great significance to operational fore-
casters. For example, the maximum is positioned too
far north in CTL and the transition in the intensity of
the hourly precipitation is poorly handled by the model.
An excessively slow start to the heavy rainfall is pre-
dicted by CTL and this could have led to misleading
warnings to the public. Certainly the hourly total from
2300–0000 UTC (Fig. 4) is sufficiently innocuous so
that it would not have caused the forecaster to question
the dry model prediction for the first few hours.

b. Static-initialization experiments

As described in Table 2, the S1 experiments differ
from CTL only in terms of the replacement of the in-
ternally determined latent-heating field by the pre-
scribed latent-heating field during a short period at the
start of the forecast. Experiment S1 is the basic ex-
periment of this series, with S2, S3 and S4 serving to
determine the sensitivity of the QPF skill to the strength,
vertical distribution and duration of the forcing, re-
spectively.

Figure 10 shows the hourly precipitation forecasted
by S1 and should be compared with the CTL forecast
in Fig. 7 and the observations in Fig. 6. Certainly the
greatest improvement in the hourly rates is during the
first three hours. Figure 5c shows the S1 forecast 3-h
total for the period 0000–0300 UTC and should be
compared with the CTL forecast in Fig. 5b. With the
1-h latent-heat forcing, a reasonably accurate 3-h QPF
is produced in terms of the position of the maximum,
the aspect ratio of the isohyets, the orientation of the
band, and the temporal continuity of the hourly in-
tensity. In this case, the mesoscale circulation that is
forced during the first hour of the forecast is retained.
The 6-h precipitation total from S1 seen in Fig. 8c shows a more intense band compared to CTL (Fig. 8b), and is probably more correct considering the observations in excess of 6 cm in Texas and Kansas (Fig. 8a). The larger 12-h totals from S1 also compare more favorably with the observations than do those from CTL (Fig. 9).

The skill scores in Table 3 also reflect the good quality of the S1 forecast. Even though the 3-h and 6-h bias scores are relatively large, the threat scores are good. For the 3-h forecast, the threat scores are 0.47 and 0.23 for the 1-cm and 3-cm thresholds, respectively, while for the 6-h forecast they are 0.18, 0.37, 0.33 and 0.21 for the 1 cm, 3 cm, 5 cm and 7 cm thresholds, respectively. As with CTL, the poor 1-cm threat score for the 6-h forecast in this experiment is a result of the erroneously large area in west Kansas and is not representative of forecast quality near the lid edge.

Experiment S2, with one-half the intensity of the latent-heat forcing used in S1, is still reasonably successful, even though use of the actual latent-heating estimate in S1 produces a more skillful QPF. Compared to S1, the maximum precipitation is 3.20 cm instead of 5.78 cm for the first hour and 2.41 cm instead of 3.29 cm for the second hour. The amounts for the third hour are similar. Therefore, a significant improvement compared to CTL is still evident with a moderate underestimate of the rainfall rate used to derive the latent-heat forcing.

In S3, when the level of the maximum in the vertical distribution function for the latent-heat forcing is lowered compared to that in S1, only a modest impact is observed. During the hour of forcing, S3 produces about 5% more precipitation than does S1, but thereafter the predictions are very similar. The abrupt shift at the end of the first hour from the forced to the model-determined vertical distribution function does not therefore have a significant damaging impact.

Finally, when the temporally constant latent-heat forcing is allowed to extend through hour two of the S4 forecast, slightly larger precipitation amounts are produced from hours two to four. The subsequent forecast is not significantly different than the one from S1 in this case, perhaps because the precipitation pattern was not evolving rapidly at this time.

This latent-heat forcing can impact the forecast skill in at least two ways. If the large-scale solution in the initial conditions is defined reasonably well, but contains specific errors that prevent the development of realistic precipitation by the model, the imposed latent-heat forcing may be able to sufficiently modify the large-scale conditions so that the model can sustain the precipitation in a natural way after the latent-heat forcing is removed. Also, an important mesoscale signal in the
stability, divergence or moisture fields may not have been defined well by the large-scale analysis, but the forcing could potentially develop it quickly during the early part of the forecast. In this case, the precipitation is produced along the edge of an elevated mixed layer because a low-level flow develops below the capping inversion and allows the moist, buoyant air to rise at the edge of the lid. Lakhtakia and Warner (1987) show that differential thermal forcing at the surface contributes to generating the initial flow from beneath the lid, which is then sustained by the low-level convergence associated with the latent-heat-driven circulation. In order to better understand whether the latent-heat forcing in S1 is developing the model precipitation in a way that is consistent with this proposed mechanism, figures of the forecast meteorological conditions at the lid edge are examined. Figure 11a shows the difference between the 850-mb winds from CTL and S1, after the one hour of imposed latent heating in S1. This confluence pattern prevails between this level and the surface, and thus the flow from beneath the lid is enhanced in S1. Note that the largest cross-lid-edge component in the velocity difference field is near the intersection of the lid edge (scalloped line) and the Texas–Oklahoma border. In Fig. 11b is an analogous plot of the wind field difference but at the 2-h time, 1 h after the latent-heat forcing is turned off. Local differences of over 6 m s$^{-1}$ still exist, where the greatest difference in the easterly flow from the warm, moist air has moved northward into Kansas, as did the forecast precipitation in S1 (see Fig. 10b and c). Thus, the latent-heat forcing seems to be inducing the more-rapid precipitation development because it is acting through a physically realistic mechanism—i.e., the rapid development of an easterly wind component below the lid. In addition, examination of equivalent potential temperature fields and model soundings (not presented) shows that the lower atmosphere is made more unstable as a result of the advection of moisture into the region by the forced convergence.

Another interesting contrast between CTL and S1 is seen in the vertical velocity field. In Lakhtakia and Warner (1987), the vertical motion pattern simulated by the model after a 12-h forecast for 0000 UTC 10 May has a banded structure near the lid edge. Figure 12 shows that this mesoscale structure, even though it cannot be verified with data, appears in the 3-h vertical velocity pattern for S1 but not for CTL. In CTL, a broad upward-motion area exists over the Texas panhandle at 1 h, intensifying and becoming more vertically oriented by 3 h. In contrast, the 1-h, S1, upward-motion maximum is farther to the east and narrower. At 3-h, the S1 vertical motions have developed the noted banded structure, with subsidence over the lid to the east of the lid edge, upward motion colocated with the precipitation maximum, and an upward/downward-motion couplet very closely positioned to the west of the precipitation band. Perhaps most imp-

![Figure 11](image.png)

**Fig. 11.** Vector wind difference (m s$^{-1}$) between S1 and CTL ($\mathbf{V}_{S1} - \mathbf{V}_{CTL}$) at 850 mb for 1 h (a) and 2 h (b) forecasts. The scalloped line shows the position of the lid edge at this time.

Importantly, the more focussed and intense vertical velocity pattern in S1 is shown to be retained at 3 h, two hours after the forcing is terminated.

c. Dynamic initialization experiments

Before the DI experiments that are listed in Table 2 are discussed, a few preliminary tests will be described. Recall that, during the preforecast period, the model should develop mesoscale structure in the fields while maintaining a reasonable large-scale solution. In these DI experiments, the model physics (differential surface
Fig. 12. Potential temperature (solid, K) and vertical velocity ($a$, $b \times 10^{-4}$ mb s$^{-1}$ and $c$, $d \times 10^{-3}$ mb s$^{-1}$) along cross-section CS2 (Fig. 1) for CTL at 1 h (a) and 3 h (b) and for S1 at 1 h (c) and 3 h (d). Isotachs in regions of upward motion are dashed.
Fig. 12. (Continued)
heating, etc.) and the external latent-heat forcing should cause mesoscale features to develop while the Newtonian relaxation limits the large-scale error growth and allows the development of a large-scale balance (i.e., the relaxation is toward the large-scale analysis at 0000 UTC 10 May). In order to test the degree to which the Newtonian relaxation affects the large-scale solution, a forecast without the relaxation was performed for the 12-h preforecast period only. Figure 13b shows the predicted precipitation from this forecast for 2100 UTC 9 May to 0000 UTC 10 May, and should be compared with Fig. 13a showing the observed precipitation for this period. Clearly the predicted distribution and intensity of the precipitation are totally incorrect, and it is these errors that the Newtonian relaxation and the latent-heat forcing must overcome in order that realistic synoptic-scale and mesoscale circulations exist at 0000 UTC. Additionally, two tests of the Newtonian relaxation are performed that differ from the basic experiment described below in which both the wind and temperature fields are relaxed toward the analyses at 0000 UTC. In one test, only the wind field is relaxed, and in the other only the temperature field is relaxed. Neither of these two experiments produce a significant improvement in the QPF compared to CTL, the forecast without relaxation. Also, experiments with both temperature and wind simultaneously relaxed toward the 0000 UTC analyses are performed using various relaxation coefficients. An experiment with a weak relaxation coefficient of $1 \times 10^{-4}$ produces a precipitation field during the preforecast period that is similar to the one produced with no relaxation (Fig. 13b). The actual forecast, after 0000 UTC, was worse than from CTL.

In experiment D1, a moderate relaxation coefficient of $3 \times 10^{-4}$ is used. The simulated precipitation during the preforecast period from 2100 UTC 9 May to 0000 UTC 10 May is seen in Fig. 13c to be much better than that shown in Fig. 13b for the same period when no relaxation was employed. The predicted precipitation totals for the 0000–0300 UTC, 0000–0600 UTC and 0000–1200 UTC forecast periods are shown in Figs. 5d, 8d and 9d. The location of the precipitation is too far to the west during the first three hours and the intensity is too weak compared to the observations. During the latter part of the period, the fields have errors similar to those from CTL. This prediction is generally inferior to that from S1 in terms of both the position and intensity of the heavy precipitation, but it is definitely better than the one from CTL. It is worth noting, however, that D1 is better than S1 in terms of the more-realistic southward extension of the precipitation into the eastern Texas panhandle after 0300 UTC (Fig. 8d). The skill scores for D1 are low because of the low intensity and westward displacement of the

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Fig. 13. Observed and model-simulated precipitation (cm) for 2100 UTC 9 May–0000 UTC 10 May 1979, the last 3 h of the preforecast period: observations (a), forecast with no Newtonian relaxation (b), D1 (c), D2 (d), D3 (e) and D4 (f).
heaviest predicted precipitation. However, for both the 3-h and 6-h totals, D1 is better than CTL for all 1-cm threshold scores. D1 is even better than S1 in terms of the 6-h, threat and bias scores with 1 cm threshold, primarily because it did not overforecast at this level in western Kansas and it correctly extended the weak precipitation farther to the south.

Use of a stronger relaxation coefficient of $5 \times 10^{-4}$ in D2 has two effects. Compared to D1, the position error is slightly less but the precipitation amounts show an even greater deficit. In D1, the 3-h maximum is 3.09 cm (Fig. 5d), but in D2 it is only 1.22 cm (not shown). The erroneously low intensity in both D1 and D2 is caused by the fact that the preforecast relaxation is toward a large-scale analysis which does not contain the mesoscale circulation needed to support the precipitation. Therefore, as the model dynamics produce mesoscale structures associated with the rainfall, differences develop compared to the 0000 UTC analysis, and the relaxation forces the model solution toward the large-scale analysis. This points out a shortcoming of this relatively simple relaxation procedure; one cannot relax toward a large-scale analysis to control the error growth on this scale without simultaneously suppressing, to some extent, the internal development of mesoscale features. A better technique would involve spectrally decomposing the model solution every time step and then forcing (relaxing) the solution based on the difference between the large-scale analysis and the large-scale component of the solution.

Because Newtonian-relaxation has been used in relatively few mesoscale modeling studies, especially ones whose focus is the quantitative precipitation forecast, it is reasonable to inquire about the mechanisms by which the procedure modified the precipitation during the preforecast period and the reasons why the QPF skill from D1 is better than that from CTL.

Figure 14 shows the 0000 UTC 10 May temperature (C) at about 50 m above the surface and the vertical velocity ($10^{-2}$ mb s$^{-1}$) at 700 mb, based on observations (kinematic vertical velocities), the 12-h preforecast without Newtonian relaxation, and D1 with Newtonian relaxation. The D1 and observed temperature fields are quite similar because the predicted temperature is relaxed toward the analysis. Without the constraint of the relaxation term, frontogenesis takes place but the position of the front is too far to the northwest. Because the observed front was located in the Texas–Oklahoma panhandle and oriented in a northeast–southwest direction during the preforecast period, this caused the model-predicted temperatures for 0000 UTC 10 May (Fig. 14c) to be too high in the northern Texas panhandle, the Oklahoma panhandle and southwest Kansas. The precipitation for the preforecast period without Newtonian relaxation, shown in Fig. 13b, is clearly associated with the erroneously placed front. With the Newtonian relaxation in D1, the maximum temperature gradient (Fig. 14b) is more correctly located and the upward motion pattern is positioned closer to the observed location of the heavy precipitation. Figure 15 shows the 0000 UTC 10 May equivalent potential temperature ($\theta_e$) and vertical velocity fields along cross section CS1 (Fig. 1) based on observations, D1, and the 12-h preforecast without Newtonian relaxation. Of the three $\theta_e$ fields, the one based on the objective analysis of the data most poorly shows the probable structure of the elevated mixed layer and the moist, unstable boundary layer below. The other two, based on the model output, show a stronger and better organized $\theta_e$ gradient between the moist boundary-layer air below the lid, the dry elevated-mixed-layer air above the lid and the dry boundary-layer air to the west. The shape of the elevated mixed layer (as subjectively defined by the heavy dashed line) also conforms better with the analysis in Carlson et al. (1983). The upward motion maximum in D1 (Fig. 15b) is located at the lid edge and is relatively deep, whereas the maximum derived from the analysis of the synoptic-scale data (Fig. 15a) is shallow and less organized. The upward motion maximum in Fig. 15c, from the preforecast without

![Fig. 14. Observed and model-simulated temperature (C, light solid) at about 50 m AGL and 700 mb vertical velocity ($10^{-2}$ mb s$^{-1}$, heavy solid and dashed) at 0000 UTC 10 May 1979: observations (kinematic vertical velocities) (a), D1 (b) and the preforecast with no Newtonian relaxation (c).](image-url)
Fig. 15. Equivalent potential temperature (K, light solid) and vertical velocity ($10^{-3}$ mb s$^{-1}$, heavy solid and dashed) at 0000 UTC 10 May 1979: observations (kinematic vertical velocities) (a), DI (b) and the reforecast with no Newtonian relaxation (c). The heavy dashed line on the east side of the cross section between 450 and 750 mb delineates the subjectively defined position of the elevated mixed layer.
relaxation, is quite strong and may be related to the lid edge in spite of its proximity to the frontal, upward-motion maximum to the north. The stronger upward motion without the relaxation (Fig. 15c) compared to D1 with the relaxation (Fig. 15b) results from the fact that the model solution in D1 is being relaxed to the observed horizontal winds which are the basis for the weak kinematic vertical velocities seen in Fig. 15a.

In summary, the use of Newtonian relaxation in D1 forces the model solution for the temperature toward a more realistic frontal position than would have been produced by the internal model dynamics. This prevents the development of erroneous precipitation during the preforecast period, but does allow some precipitation to form near the Texas–Oklahoma panhandle (lid-edge). Even though the temperature field from D1 is forced to be very similar to the data analysis because of the relaxation, a model-generated mesoscale structure to the nonrelaxed moisture field does reflect a probably realistic $\theta_e$ field at low levels. This again points out the need to choose a relaxation coefficient that will allow the development of a mesoscale component to the model solution.

Even though the D1 by Newtonian relaxation toward the large-scale analyses produces a definite improvement in the 0000–0300 UTC QPF compared to the CTL, there is the noted underprediction of the mesoscale precipitation amounts as well as a position error in the location of the heaviest rainfall. These problems are partially corrected in D3 by including the latent-heat forcing in the model's thermodynamic equation during the preforecast period. The preforecast precipitation totals for D3 (Fig. 13e) are more realistic than those from D1 (Fig. 13c) in which preforecast latent-heat forcing is not used, although the intensity is probably erroneously large. The results for this experiment, shown in Fig. 5e for the 0000–0300 UTC QPF and in Fig. 8e for the 0000–0600 UTC QPF, should be compared to the analogous forecasts from D1 and S1. The use of the preforecast latent-heat forcing has increased the precipitation intensity compared to D1 as well as partially corrected the westward-displacement error in the pattern. However, D3 is still generally inferior to S1 in terms of most verification scores, except that the 1-cm and 3-cm threat scores for the 6-h rainfall are better for D3 than for S1.

In experiment D4, where the preforecast latent-heat forcing is used without the Newtonian relaxation, the QPF (not shown) is worse than from CTL. The mesoscale information provided by the 3 h of light, latent-heat forcing is not able to compensate for the modest forecast errors that develop during the preforecast period. However, the Newtonian relaxation in D3 is clearly sufficient to, at least partly, suppress their development. By comparing the preforecast precipitation from D4 and D3 in Figs. 13f and 13e, it can be seen that the Newtonian relaxation has a significant benefit when combined with the preforecast latent-heat forcing.
In experiment D5, which uses the Newtonian relaxation during the 12-h preforecast period and the 1 h of latent-heat forcing during the first forecast hour, the objective verification scores are close to those for S1, which uses only the latent-heat forcing, especially for the smaller thresholds. Bias scores are uniformly smaller and better in D5 compared to S1, where this is probably a result of the fact that the latent-heat forcing is imposed on a vertical motion pattern that is incorrectly located during the 12-h preforecast period. For example, note that the 3-h precipitation produced by use of the Newtonian relaxation alone (Fig. 5d) is too far to the west, and therefore the latent-heat forcing in D5 probably had a "competing" upward motion maximum to its west.

Combining the Newtonian relaxation with latent-heat forcing during the first forecast hour (D5) or during the preforecast period (D3), is preferable to use of the Newtonian relaxation alone (D1). The 3-h threat scores show D5 to be somewhat superior to D3, but the 6-h threat scores show D3 to be moderately more skillful than D5 for two of the four thresholds, implying that the preforecast assimilation of latent-heating information may be more long-lasting.

6. Summary and conclusions

The use of a conventional (control) technique to initialize a Penn State/NCAR mesoscale model simulation of the heavy convective precipitation during the SESAME IV study period produces considerable error in the precipitation field for the first 3-4 h of the 12-h forecast period (0000-1200 UTC 10 May). The purpose of this study is to investigate ways in which operationally available data may be used to improve this very-short-range QPF. A static-initialization procedure that employs hourly rainfall and radar data valid near the initialization time, in order to specify a latent-heat forcing function during the first hour of the forecast period, shows considerable success compared to the control, especially during the first three forecast hours. Whereas the control forecast reflects no precipitation-line formation in the Texas and Oklahoma panhandles during this 3-h period, this experimental forecast produces heavy rainfall that lasts beyond the forcing period and it reproduces the intensity, position, aspect ratio and orientation of the line reasonably well. Six-hour and 12-hour accumulations are also predicted better in this experimental forecast because there is less deficit during the first three to four hours. That is, the control experiment does not produce a compensating excess precipitation later in the forecast, that serves to balance the deficit during the early period.

Dynamic-initialization experiments are also performed. In these, a Newtonian relaxation toward analyses of winds and temperature valid at the initialization time is used independently and in conjunction with latent-heat forcing that is based on observed precipitation during the preforecast period. The forecasts from each of these experiments are better than the control forecast, illustrating that Newtonian relaxation is capable of improving a very-short-range QPF. However, for this case, the application of latent-heat forcing in the first forecast hour is more effective than in the preforecast hours.

It is encouraging that these dynamic-initialization procedures perform reasonably well because they can be viewed as less limited in one sense than the static-initialization procedure. That is, the application of a constant latent-heat forcing during the first forecast hour after the static initialization would not be as reasonable, and perhaps not as effective, for a rapidly evolving precipitation event.

Even though some of the tested latent-heat forcing and Newtonian-relaxation initialization techniques, produced an improved QPF, and therefore are successful in a technical sense at least, it is worth examining whether this improvement could have benefited the operational forecast. Because the precipitation event is in progress at the initialization time, persistence procedures must be viewed as legitimate means of producing a one- to two-hour QPF. However, the observed hourly precipitation amounts reflected in Fig. 4 for the preforecast period are not good predictors of the rates for the first three hours of the forecast, shown in Fig. 6a-c. Perhaps an experienced forecaster could have interpreted the radar charts available near the initialization time to infer the future growth in rainfall intensity during the first few hours after 0000 UTC, but only further modeling and nowcasting experience with a large number of cases will allow a better definition of the point where the numerical models' and the nowcasters' skill curves cross. Rapidly evolving precipitation events will potentially be forecast better using these experimental initialization techniques than by persistence, however, even though this also has to be demonstrated by experience with a variety of cases.

Further improvement in the QPF for this case beyond that provided by the special initialization techniques described here, may require the use of more sophisticated physical parameterizations. Certainly the Kuo-type parameterization for subgrid-scale precipitation should be viewed as a probable source of error, and may have limited the degree of improvement in the QPF achieved in these initialization experiments. Additional development effort should also be devoted to the testing of Newtonian-relaxation procedures that utilize data available for times during the relaxation period, rather than just at the end of it as in this study. Even though the investigation of the numerous alternative relaxation algorithms which employ time and space dependent relaxation coefficients is beyond the scope of this study, such capabilities would allow the three-hourly SESAME radiosonde data to be used in the preforecast relaxation period for this case.

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REFERENCES