Phase-Correcting Data Assimilation and Application to Storm-Scale Numerical Weather Prediction. Part II: Application to a Severe Storm Outbreak

KEITH A. BREWSTER
Center for Analysis and Prediction of Storms, University of Oklahoma, Norman, Oklahoma

(Manuscript received 8 May 2001, in final form 27 July 2002)

ABSTRACT

A scheme to correct phase errors in numerical model forecasts using Doppler radar, radiosonde, profiler, and surface data is demonstrated to improve forecasts in a complex severe thunderstorm situation. The technique is designed to directly address forecast phase errors or initial position errors as part of a data assimilation strategy. In the demonstration the phase error correction is applied near the time of initial cell development and the forecast results are compared to the uncorrected forecast and forecasts made using an analysis created at the time of the observations. Forecasts are verified qualitatively for the position of thunderstorm cells and quantitatively for accumulated precipitation. It is shown that the scheme can successfully correct errors in thunderstorm locations and it has a positive influence on the subsequent forecast. The advantage of the phase correction over the control lasts for about 3 h despite storm dissipation and regeneration, and interactions among multiple storms.

1. Introduction

Phase errors of waves, propagation speed errors in mesoscale features, and errors in the initiation location of individual storms are common in meso- and storm-scale forecasts, and such errors can be vexing to correct in analysis schemes, especially those using data from a single time. It is of interest to investigate whether such errors can be addressed directly in an analysis or data assimilation system.

Brewster (2003, hereafter Part I) summarizes some recent work in the correction of phase errors, describes a method for identifying and correcting such errors in thunderstorm forecasts using radar and mesoscale datasets, and presents encouraging results for an observing system simulation experiment. The phase-correction method described therein seeks a local translation, described by a field of translation vectors, $\delta x$, to apply to the forecast field in order to shift and distort it to best match the observed data. A minimization of the mean-square difference from the observations is used to find the phase error for each of several test volumes within the forecast domain. A functional based on squared differences is formed:

$$J[\delta x] = \frac{s(|\delta x||l^{-1})}{\sum_{j=1}^{\alpha} \sum_{i=1}^{N} \frac{[H(F(x, + \delta x)) - o_j(x)]^2}{\sigma_{ij}^2}},$$

where $o_{ij}$ is an observation, $x_i$ is the observation location, $\delta x$ is the horizontal displacement vector, and $\sigma_{ij}^2$ is the expected observation variance, which, in general, is a function of variable, data source, and height. The symbol $F$ represents the forecast field smoothed by a nine-point filter in two dimensions; the smoothing is done to avoid fitting small-scale noise to the observations. Here, $H$ represents a transformation, if necessary, from the forecast variables to the observed quantity. Each variable is weighted according to $\alpha$, which may

<table>
<thead>
<tr>
<th>Pass</th>
<th>$i, j$ vol width</th>
<th>$i, j$ overlap</th>
<th>$k$ vol height</th>
<th>$k$ overlap</th>
<th>$N_i$ vol</th>
<th>U/A</th>
<th>Surface</th>
<th>Radar</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>61</td>
<td>31</td>
<td>3</td>
<td>10</td>
<td>2</td>
<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>2</td>
<td>35</td>
<td>18</td>
<td>6</td>
<td>10</td>
<td>2</td>
<td>6</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>3</td>
<td>16</td>
<td>15</td>
<td>12</td>
<td>15</td>
<td>No</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>4</td>
<td>10</td>
<td>5</td>
<td>25</td>
<td>12</td>
<td>2</td>
<td>25</td>
<td>Yes</td>
<td>Yes</td>
</tr>
</tbody>
</table>
TABLE 2. Variable weighting assignments.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Weight, $\alpha$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u$ wind</td>
<td>5.0</td>
</tr>
<tr>
<td>$v$ wind</td>
<td>5.0</td>
</tr>
<tr>
<td>Pressure</td>
<td>0.0</td>
</tr>
<tr>
<td>Potential temp</td>
<td>5.0</td>
</tr>
<tr>
<td>Specific humidity</td>
<td>3.0</td>
</tr>
<tr>
<td>Reflectivity</td>
<td>5.0</td>
</tr>
<tr>
<td>Radial velocity</td>
<td>2.0</td>
</tr>
</tbody>
</table>

account for the total number of observations of each type or the anticipated usefulness of a particular variable in determining the displacement error.

The leading term on the right-hand side, $s$, is a distance-dependent function that increases $J$ with increasing distance. This distance penalty term is designed to prevent aliasing and to prevent the erroneous identification of position errors that might otherwise occur due to random observation error. The function used here follows from Thiebaux et al. (1990), the inverse of the second-order autoregressive (SOAR) function:

$$s(|\delta x|^{-1}) = \frac{\exp(|\delta x|^{-1})}{(1 + |\delta x|^{-1})},$$

where $l$ is a length scale parameter. Here $l$ is set as

$$l = 0.5\sqrt{L_x^2 + L_y^2},$$

where $L_x$ and $L_y$ are the lengths of the sides of the test volumes (discussed below) in the $x$ and $y$ directions, respectively.

The sum is normalized by a factor, $N_a$:

$$N_a = \sum_{j=1}^{n_{var}} \sum_{i=1}^{n_{obs}} \alpha_j,$$

The normalization factor is included to account for the
fact that observations may “drop out” of the calculation of $J$ in the special case where the region is near the domain boundary and the test shift vector takes the observation outside the forecast domain. This could otherwise decrease $J$, because the total number of observations in the sum is decreasing, and potentially lead to a false minimum in the functional.

Table 1 shows the dimensions and data usage for the phase correction applied to the data in this work. Volume dimensions are given in grid lengths, with the grid length being 3 km. In the table, “overlap” is the overlap of each test volume with its neighbor; “$N_x$,” “$N_y$,” and “$N_z$” refers to the number of volumes in each of the $x$, $y$, and $z$ directions, and $N_z$, the number of volumes counting in the vertical (a total of $N_x \times N_y \times N_z$ volumes in the domain).

Four iterations of the phase-correcting scheme are used in the work presented here. The first pass is designed to seek the synoptic-scale error using large test volumes. The process continues using sequentially smaller test volumes, and includes the more dense radar data in the final two passes. In the second and subsequent passes the incremental phase corrections are summed with the result from the previous iteration(s).

In the case presented here, all observed variables are used in evaluating the error functional except pressure, which would be complicated by the slope of the model surfaces (when the shift is applied, the shift is applied to the perturbation pressure to avoid affecting the mean
vertical pressure distribution due to the slope of the model surfaces). The radar data used here consist of radar radial wind and reflectivity remapped to the 3-km resolution grid by averaging all data within each grid cell. The radial velocities were calculated using the local slope of the radial from the four-thirds-earth model (detailed in Part I). The hydrometeor terminal velocity was removed from the observations using a simple parameterization of the terminal velocity from the reflectivity following Ziegler (1978). The weight assigned to each variable, $a$, is detailed in Table 2.

The case of 8 June 1995 is used to examine the ability of the assimilation techniques to correct errors in forecasts of storms at the time of early convective development on a day of severe weather. The synoptic setting for this case is described in section 2. A description of the mesoscale spinup forecast used to create the mesoscale features used as background forecasts for the storm-scale (3-km grid scale) experiments follows in section 3. A qualitative comparison among data assimilation schemes for the storm-scale forecasts is presented in section 4. Section 5 focuses on the verification of the quantitative precipitation forecasts. Discussion of the results and future plans follow in sections 6 and 7, respectively.

2. The 8 June 1995 synoptic setting

Eight June, 1995 is a major case for the Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX; Rasmussen et al., 1994) as several damaging tornadoes were produced by storms in the eastern Texas Panhandle. This section details the synoptic and mesoscale setting for this case.

At 1800 UTC on 8 June, the surface map in the southern plains region (Fig. 1) featured a very slow-moving cold front from east of Wichita, Kansas, extending into the northern Texas Panhandle and continuing into the foothills of the Rockies in north-eastern New Mexico. There was a dryline in the west-

---

Fig. 8. Surface temperature (°C) and wind (barbs). Observations and model forecast at 2010 UTC 8 Jun 1995. Full barb is 5 m s$^{-1}$.

Fig. 9. Phase-shift vectors for the 10 m AGL model level at 2010 UTC. Vector scale in the lower-left corner is in unit grid lengths (one grid length = 3 km).

Fig. 10. Surface temperature (°C) and wind (barbs). Observations and ARPS fields after phase correction.
ern Texas Panhandle, with the dewpoint temperature at Childress, Texas (CDS), of 24°C contrasting with a dewpoint of 1°C at Clovis, New Mexico (CVS). Winds behind the dryline were relatively weak. A trough of low pressure was coincident with the stationary front, but there was little indication of the dryline in the sea level pressure field.

Special soundings were taken at 1800 UTC (noon, local standard time) to assess the extent of the convective instability and strength of the capping inversion. A sounding taken by a mobile research crew in northwestern Oklahoma, at Seiling (location marked by black diamond in Fig. 1), is shown in Fig. 2, and is considered representative of air east of the dryline on this day. Due to the very moist air in this region (dewpoint of 23°C, water vapor mixing ratio of 18 g kg\(^{-1}\)), this sounding has an extremely high CAPE of nearly 4000 J kg\(^{-1}\) and the capping inversion at 750 hPa is not an impediment.
for an unmixed parcel lifted from the surface. However, due to the slight superadiabatic lapse rate just above the surface, actual buoyant plumes in that area would likely be better mixed, and thus the cap could prevent convection from occurring. Nevertheless, only slight forcing would be needed to release the available convective energy, and the wind shear was sufficient to support supercell thunderstorms.

At 1908 UTC the first convective echoes appeared on the Amarillo, Texas (AMA), radar. The first cell was along the front in the northeastern corner of the Texas Panhandle, near Perryton, Texas (labeled PYX in Fig. 1). This cell quickly increased in intensity and moved northward with time, into the Oklahoma panhandle. The cell developed a low-level circulation and produced hail and a brief tornado, but as it moved farther into the cool air north of the frontal boundary, its intensity eventually decreased.

At 2008 UTC new cells formed farther south; one was near the triple point, near Stinnett, Texas (STN), but two others developed in the warm air east of Amarillo, near Pampa, Texas (PPA). This was to be the area of the most severe storms of the day. Storms in this area produced large hail and damaging tornadoes. More details of the initiation and progression of cells in this area will be discussed with the model results.

By 0000 UTC on 9 June, a cluster of intense thunderstorms covered the northern half of the eastern Texas Panhandle and gradually moved eastward into western Oklahoma, continuing to produce severe weather into the evening hours.

In summary, the synoptic background was favorable for storms as it provided significant vertical wind shear and low-level winds favorable for transporting the unstable air into the region. The quasi-stationary frontal boundary and the dryline provided convergent low-level flow to initiate convection, but the capping inversion was weak enough that cells also formed just ahead of those boundaries. There was no evidence of a strong traveling synoptic-scale short wave to drive the surface boundaries and convection, so the motion of the storms was largely due to diurnal progression of the dryline, advection of individual cells, and interaction among the storms.
3. The 12-km mesoscale spinup simulation

Although we seek to make a storm-scale simulation, the storm-scale model run will require time-dependent lateral boundary conditions and it is desirable to have an initial field that contains a representation of the mesoscale. Therefore, a 12-km mesoscale spinup simulation is made using the Advanced Regional Prediction System (ARPS) nonhydrostatic model (Xue et al. 1995, 2000, 2001). Figure 3 shows the domain of the 12-km run with the 3-km domain embedded. A schematic of the assimilation procedure is shown in Fig. 4, and is detailed in this section.

Because convection began in the early afternoon, a midmorning initialization is chosen for the spinup. A 9-h 12-km forecast run in the region of interest beginning at 1500 UTC is made using the ARPS. The National Centers for Environmental Prediction (NCEP) Rapid Update Cycle (RUC) forecast initialized at 1200 UTC is used to begin the process and for lateral boundary conditions. The RUC forecast fields are interpolated to the ARPS model grid and analysis increments based on the observed data are calculated using the analysis program of the ARPS Data Assimilation System (ADAS; Brewster 1996). The ADAS analysis program uses the successive correction technique of Bratseth (1986), which converges to the optimal interpolation solution.

Surface data used in ADAS include the surface airways observations, the Oklahoma Mesonet, the Colorado Agricultural Meteorological Network, and the Department of Energy’s Atmospheric Radiation Measurement (ARM) surface network. Aloft, data from the ver-
tical wind profilers are used. There were some special soundings taken at 1800 UTC for the VORTEX project that were included in the analysis for that hour. Radar data are used in the ADAS cloud analysis algorithm, but the raw velocity data were not used at this scale. Satellite infrared (IR) and visible data from the Geostationary Operational Environmental Satellite-8 (GOES-8) are also used in the ADAS cloud analysis [Zhang et al. (1998), following from the Local Analysis and Prediction System (LAPS) algorithms; Albers et al. (1996)].

The analysis increments are introduced to the model using incremental analysis updating (similar to Bloom et al. 1996) employing a constant time weighting over a 10-min window. The data analysis and analysis increment updating are repeated every hour for 1600, 1700, and 1800 UTC, with the increments calculated at 10 min before the hour (generally corresponding to the time of the surface observations) and applied during the 10 min preceding the top of each hour.

Figure 5 shows the ARPS model assimilated state at 1800 UTC, plotted with observations at the same time, for surface temperature, dewpoint temperature, and winds and pressure. Note that the assimilation system has developed tight gradients in temperature along the front in the Oklahoma panhandle and in the dewpoint fields in west Texas and southern Colorado. Although this model run did not use a convection parameterization nor did it have sufficient resolution to fully resolve thunderstorms, it was able to capture the cold air pocket, divergent winds, and pressure anomaly (not shown) of the thunderstorm outflow in northwestern Oklahoma.

The 3-h 12-km forecast valid at 1800 UTC is interpolated to the 3-km grid to provide the initial conditions for the control experiment and an analysis background for the other experiments. The 12-km run is continued until 0000 UTC 9 June in order to generate boundary conditions for the 3-km run. The boundary conditions are thus used in a one-way nesting arrangement. No additional data are provided to the 12-km run beyond 1800 UTC. The 3-km model is run from 1800 to 0000 UTC.

4. The 3-km storm-scale simulations

The phase correction procedure is applied using radar and surface data at the time of early storm growth in a 3-km forecast initialized at 1800 UTC. The four experiments are “control,” no additional corrections are applied after the 12-km spinup; “shift,” only the phase correction is applied using the single-step shift method; “ADAS only,” the ADAS analysis is run using the radar and mesoscale data; and “shift + ADAS,” ADAS is run using the phase-corrected model fields as a background.

The storms in the area of interest began about 2000 UTC, with the exception of a single storm that formed in the Oklahoma panhandle about an hour earlier. The storms grew quite rapidly, and by 2010 UTC there was sufficient radar reflectivity observed to define the principal initial cells in the northeast Texas Panhandle. Figure 6 is the Amarillo radar reflectivity for 2008 UTC. At this time, there are also thunderstorms developing in the model forecast; Fig. 7 is the simulated reflectivity at the lowest model layer (10 m AGL) for 2010 UTC. The reflectivity is derived from the model hydrometeors using relationships from Kessler (1969) and Rogers and Yau (1989):

\[
R = 10 \log_{10} \left\{ 1.73 \times 10^4 (10^3 \bar{\rho} q_1) \right\}^{3/4} + 3.8 \times 10^4 [10^3 (10^3 \bar{\rho} q_1 + q_h)]^{2/3},
\]

where \( R \) is the reflectivity in dBZ; \( q_r, q_s, \) and \( q_h \) are the forecast rainwater, snow, and hail concentrations (kg kg\(^{-1}\)), respectively; and \( \bar{\rho} \) refers to the horizontal mean atmospheric air density (in kg m\(^{-3}\)).

The ARPS does an excellent job in developing thunderstorms in the Texas Panhandle at about the right time. Although similar storms are formed in the model at this time, there are some differences in the location of individual cells. The 3-km run was not as close in its forecast of the cell that had formed early in the afternoon in the Oklahoma Panhandle; it formed late, and is displaced to the northeast, and subsequently moved out of the domain. Nevertheless, it may have an effect on later storms through, for example, interaction with its outflow. The model forecast of the wind shift associated with the dryline was a little too far west (approximately 30 km, near Borger; see Fig. 8). The largest errors, however, were in the Oklahoma panhandle, where temperatures behind the stationary front were too warm—up to 5°C (excluding a larger error at a thunderstorm-influenced observation). In some locations, the winds
behind the front were more easterly than the north-easterly observed winds.

The phase-correction procedure was applied at this time. The field of phase-correction vectors at the lowest model level is shown in Fig. 9. We see a general shift to the south, likely in response to the temperature and wind errors behind the front, and the phase-corrected surface temperature field (Fig. 10) shows improvement due to that adjustment. The phase correction aloft (Fig. 11) identifies the northeast displacement of the cell in the Oklahoma panhandle as well as the displacement of two areas of convection in the northeast Texas Panhandle: one toward the north (southward correction indicated), the other toward the southwest. Figure 12 shows the reflectivity derived from the model hydrometeors after the phase correction has been applied. Due to the extent of the error in the development of the storm in the northeast corner of the domain, it has a distorted appearance compared to the cell shape in Fig. 7, but the other cells seem to have been repositioned well with some minor broadening.

Figure 13 is the ADASonly initial field at 2010. No reflectivity appears at level 2 because the cloud analysis first zeroes out the hydrometeor fields and the cells do not appear on radar at that low height, due to the beam height. The cell in the Oklahoma panhandle is rather distant from the Amarillo and Dodge City radars, so reflectivity for it first appears in the analysis at level 20 (about 2100 m AGL), as shown in Fig. 13. At that level the cells do appear well positioned. The ADAS analysis applied on the phase-corrected field, shift+ADAS at the surface, and at level 20 is shown in Fig. 14. The re-
Reflectivity fields in the shift + ADAS initialization are the same as in the ADAS-only experiment, as expected due to the hydrometeor zeroing removing any differences in the first guess hydrometeors in between ADAS-only and shift + ADAS.

Through 1 h after the data time, the experiments that include phase correction maintain a distinct advantage over the control and ADAS-only runs. Figure 15 is the observed low-level radar reflectivity from Amarillo at 2038 UTC. Figure 16 shows the model solutions for the four experiments valid at 2040 UTC. At this time, the most notable difference among the runs is the position of the southernmost cell. The control and ADAS runs have the cell too far south, consistent with the result at 2010 UTC. Although the ADAS run started with reflectivity initialized farther south and the hydrometeors were zeroed out, the lack of adjustment to the temperature fields resulted in the reappearance of the cell to the south and only a small cell remains in the position of the actual cell. The shift and shift + ADAS runs have a slightly larger cell in the Oklahoma panhandle consistent with the large observed storm there, though none seem to have an accurate depiction of its shape. All the runs have some spurious convection in the southeast corner of the domain and in the northeastern Texas Panhandle. The weak cap on this day makes the runs particularly sensitive to surface heating and gravity waves from the storms.

One hour after the data time, some storm development and interaction has occurred so that the original cells are not as distinguishable, but the net effect of phase correction on the forecasts is still positive. Figure 17 is the low-level reflectivity at 2110 UTC [note: due to a hardware problem with the level II data recorder at Amarillo, the 2110 UTC and subsequent radar images are produced from Weather Surveillance Radar-1988 Doppler (WSR-88D) level III Next-Generation Weather Radar (NEXRAD) Information Dissemination Service; (NIDS) datasets]. The main difference among the forecasts, shown in Fig. 18, is a break in the north–south line of convection down the middle of the figure in the control and ADAS-only forecasts and not in the radar image or the experiments including phase correction.

The advantage of the phase correction over the other experiments is less clear at 2 h after data time as nonlinear interactions among the cells continue. Figure 19 is the Amarillo reflectivity at 2159 UTC compared with the forecast fields shown in Fig. 20. The phase-corrected forecasts show a more linear organization to the eastern Texas Panhandle convection similar to the radar-observed echoes. This difference is most pronounced at the southern extent of radar echoes, though all the forecasts have too broad a coverage of simulated radar echo and a noisy appearance in the northeast corner of the domain. The slight advantage for the phase-corrected forecast is evidenced by a more solid appearance of the north–south line and a lack of spurious convection in northwest Oklahoma.

By 2300 UTC, about 3 h after the data time (Figs. 21 and 22 for the radar and model fields, respectively) forecast errors have accumulated so there are few differences among the forecast runs. It is of interest to note, however, that these forecasts are all fairly accurate in that they all contain an indication of the most significant cells. The cells at the southern end of the line in the eastern Texas Panhandle produced the damaging tornadoes. The forecasts also correctly predicted the general structure of the cells. There is evidence of the strong rotation (suggestive of the tornadic potential of these cells) in both of those storms, including the strong reflectivity gradients on the south side, surface vorticity, and apparent reflectivity appendages on the southwest flanks of the southernmost cells. The structural features are most clear in the shift experiment. Those cells are only about 10–20 km too far north in the model forecasts compared to the actual tornadic cells.

5. The 3-km precipitation verification

For a quantitative verification of the forecast experiments, the precipitation forecasts are compared to the hourly stage III precipitation fields (Fulton et al. 1998) computed by the Arkansas-Red Basin River Forecast Center (ABRFC). The observed precipitation fields are produced by estimating the rainfall rate from the radar, correcting the observations for biases based on rain gauge observations (one bias coefficient computed per radar), and merging the bias-corrected rainfall data from different radar sites onto a 4 km × 4 km grid.

The forecasts are compared by examining the rainfall bias and the equitable threat score. The rainfall bias is defined as
Fig. 20. ARPS forecast fields of 10 m AGL winds (barbs) and reflectivity (dBZ) at 2200 UTC 8 Jun: (a) control, (b) shift, (c) ADAS only, and (d) shift+ADAS.

Bias = \( \sum_{i,j} \frac{R_{\text{est}}(i, j)}{R_{\text{verif}}(i, j)} \),

where the summation is done over all \( i, j \) grid points (excluding a frame of five points along the boundaries), \( R_{\text{est}} \) represents the forecast hourly precipitation, and \( R_{\text{verif}} \) is the observed, or verification, precipitation. In this case, the verification precipitation is provided by the ABRFC 4-km rainfall analyses interpolated to the forecast grid.

The areal overlap of the model and verification precipitation is measured by the equitable threat score (Schaefer 1990; Rogers et al. 1996). The equitable threat score (ETS) is measure of forecast skill defined as

\[
\text{ETS} = \frac{H - \text{Ch}}{F + O - H - \text{Ch}},
\]

where \( H \) is the number of points where the model correctly forecasted precipitation over a specified threshold (number of “hits”), \( F \) is the number of grid points with forecast precipitation above the threshold, \( O \) is the number of points with observed precipitation above the threshold, and \( \text{Ch} \) is an estimate of the number of points that could be correctly forecasted by chance, estimated by

\[
\text{Ch} = \frac{F \times O}{N_f},
\]

where \( N_f \) is the total number of points in the forecast domain. A perfect forecast would achieve an equitable
The threat score of 1.0; a forecast with no skill, just based on chance, would have an ETS of 0.0. It is possible for ETS to be negative, if the forecast is worse than that expected for chance. ETS has an advantage over the common threat score in that the forecast cannot score higher simply by producing more precipitation. In this work, 1 mm is chosen as the threshold for computing ETS.

The four forecast experiments for 8 June 1995 are scored. Hourly periods ending at 2100–0000 UTC on 9 June were used. It should be noted that the model-accumulated rainfall for 2100 UTC is actually a 50-min rainfall, but because the storms were in an early growth stage from 2000 to 2010 UTC, it is likely that very little precipitation was missed in that 10-min window.

The rainfall bias is shown in Fig. 23. The model, in general, tends to overpredict rainfall in the early stages of this event, but the bias decreases after the first 2 h. Generally one expects a shortfall in model precipitation early in the run, due to model spinup delays, but the 3-km preforecast period has apparently provided the necessary spinup to develop precipitation, and that spinup occurred during a time when there was no observed precipitation.

Examination of the precipitation output from the model runs (not shown) reveals that the rainfall is also heavier in the model fields, even where it is correctly positioned. Zhang (1999) also found positive rainfall biases in ARPS forecasts with diabatic initialization and after precipitation was spun up in forecasts without diabatic initialization. Some of the excess rainfall could be due to the 3-km resolution forecasts not resolving the strength of the updrafts needed to suspend the hail and large water drops in these strong storms. While the strength of the updraft is generally scale dependent, the terminal velocities in the model are fixed. It is beyond the scope of this work to thoroughly examine precipitation efficiency in the ARPS model, though separately work is being done to identify and correct this tendency.

The biases are larger for the experiments that included ADAS. This is likely due to the addition of latent heat and moisture in the analysis. While the analysis zeros the hydrometeor fields in areas without observed cells, the wind and thermodynamic fields are not readjusted in areas where storms are “removed” by this process. Storms may then reappear where they had initially been removed. This is a shortcoming of the analysis when used with a high-resolution forecast as a background field without correspondingly high resolution thermodynamic and wind data.

Figure 24 shows the ETS measured throughout the afternoon of 8 June. The threat scores are good for the prediction of events on such small time and space scales, and actually increase with longer forecast time. This is most likely due to improvements with time in the forecast biases. In a sense the atmosphere is catching up with the overforecasted precipitation areal coverage. The best equitable threat scores were for the forecasts that included phase shifting. This advantage remains throughout the period examined, though the margin narrowed by the end of the period. The shift experiment led all others until the last period when the shift+ADAS exceeded the shift forecast.

To compare with some previous results using the same model, the threat score (TS) is calculated as

$$TS = \frac{H}{F + O - H}.$$  

The threat scores (Fig. 25) are better for the experiments with phase correction than without. The threat scores are generally better here than the TSs reported by Zhang (1999) with the ARPS model and diabatic initialization of another severe weather case. The TSs in that work ranged from about 0.16 to 0.36; ETS was not calculated in that work.

6. Discussion and future research

In this demonstration it was shown that the phase-correcting data assimilation can be very effective at improving forecasts of thunderstorms using mesonet and radar data as its primary input. Position errors in the forecasts of both mesoscale features and thunderstorms were identified and corrected. The impact on the forecast in the severe storm case persisted in time as forecast improvement was noted beyond 2 h even in the face of complex thunderstorm interactions. By 3 h, sufficient forecast errors had accumulated in both the phase-corrected and control runs that the improvement was barely discernible in simulated radar depictions. This is to be expected because there is certainly a limit to the predictability of thunderstorms given the relatively coarse
grid resolution used and incomplete observations of convection. Model errors or biases, errors in the initial conditions, and the complex nonlinear thunderstorm interactions will eventually compound to overcome improvements in the initial conditions.

It is also worth noting that the use of a 3-h assimilation spinup at 12 km served well to form a mesoscale-assimilated state from which to launch a high-resolution (3 km) forecast for prediction of thunderstorm initiation. It was also observed that the ARPS model was capable of forecasting the region of storminess with the phase correction primarily adding skill in the location of individual thunderstorms.

The relatively low impact of ADAS in the 3-km runs may be due to the lack of surface observations in this region to add to the already spun up mesoscale features, as the data over most of the domain are sparse, except the small area within the Oklahoma Mesonet. Improvements to the ADAS cloud scheme to reduce cloud water in areas of precipitation and modifications to the application of latent heating in the model are being pursued at the time of this publication; these changes show improved impact of radar data in another test case (Brewster 2002).

Though not observed to provide the best forecast in this particular case, it is generally expected that the phase correction will be applied in concert with an analysis or assimilation method that can apply amplitude corrections to the fields. Forecasts of discrete precipitation systems or atmospheric boundaries with gradients at scales near the limit of resolution of the model could benefit most from this technique. The reduction in the
The phase-correction scheme can be extended to include the use of satellite image data with suitable transformations of the model fields to satellite-observed quantities such as albedo and cloud-top temperature applied in Eq. (1). Weather forecasters often use the satellite images subjectively for the purpose of gauging position errors in models. It should be straightforward to use the IR data by computing the cloud-top temperature of the model data using a radiation model and comparing that to the observed cloud-top temperature where the IR data are not sensing the ground temperature. That information would be grouped with separate data at the height in the atmosphere corresponding to the cloud-top temperature (keeping in mind the that phase-shift field varies in three dimensions, a height assignment is needed). Assigning a valid height to visible data is more of a problem, but perhaps not insurmountable.

The phase correction could better utilize the clear-air reflectivity information if we could exploit a relationship between the nonprecipitating clear air echoes and the model variables. Currently the transform to reflectivity only uses the hydrometeors. This improvement may be mitigated by the fact that the current system can, and is, using the wind data in the clear air, which likely is somewhat redundant with the reflectivity data, though to the human eye the boundaries do seem to stand out more in the reflectivity data. Comparison of numerical model data and radar data, with thought to theoretical radar reflectivity relations, might yield a useful relationship between high values of nonprecipitating echoes and vertical velocity, refractivity gradients, horizontal wind gradients, and/or moisture.

Fiedler (1999) has also experimented with what he terms storm surgery. To date, his storm surgery lacks an objective way to identify “storms” to insert and remove. It might be possible to utilize the results of the cloud analysis within ADAS to identify such regions in the following way. The output of the cloud analysis can
be compared to the condition of the original background field. If areas are identified as having convective clouds (an intermediate step of the current system), and the phase-corrected field does not have a storm in that region (decided based on vertical velocity and hydrometeor concentration), the methods of Fielder could be used to add the vertical wind circulation of the storm (vertical velocity as well as associated convergence at low levels and divergence aloft). Or, in the case of a spurious storm, the storm circulation could be relaxed if ADAS identified no storm at that location. The current ADAS cloud analysis would have already removed or added the cloud and precipitation variables.

As with any new technique, exposure to more cases will help us learn about the technique’s strengths and weaknesses and will provide a better quantitative measure of forecast improvement that could be gained from its use in research or operations.

Acknowledgments. The author benefited greatly from guidance from Drs. Frederick Carr and Kelvin Droegemeier. Dr. Jian Zhang kindly provided software for precipitation verification. Drs. Ming Xue and Vince Wong provided valuable guidance on use of options within the ARPS model.

Some of the computations were done on the facilities of the Environmental Computing Applications System (ECAS) at the University of Oklahoma. Supplemental data from the VORTEX project were provided by the UCAR—NOAA Joint Office for Science Support, and processed precipitation files were obtained from the Arkansas–Red Basin River Forecast Center.

Support for this work was from a portion of NSF grants to the Center for Analysis and Prediction of Storms (ATM91-200009 and ATM9909007) and a grant from the Federal Aviation Administration (DOT-FAA NA17RJ1227-01).

REFERENCES


Fiedler, B., 1999: Storm surgery and phase correction in the ARPS and COAMPS coordinate system. [Available from the author at CMRP, University of Oklahoma, 100 E. Boyd, Rm 1310, Norman, OK, 73019, or http://cmrp.ou.edu/~bfiedler/cmrp/index.html.]


Xue, M., K. K. Droegemeier, V. Wong, A. Shapiro, and K. Breitweiser, 1995: ARPS version 4.0 user’s guide. Center for Analysis and Prediction of Storms, Norman, OK, 381 pp. [Available from Center for Analysis and Prediction of Storms, University of Oklahoma, Norman, OK 73019; also available online at http://www.caps.ou.edu/]


Ziegler, C. L., 1978: A dual Doppler variational objective analysis as applied to studies of convective storms. NOAA Tech Memo. ERL NSSL-85, 116 pp. [Available from NOAA/National Severe Storms Laboratory, Norman, OK 73069.]