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

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Abstract

A scheme to correct phase errors in numerical model forecast 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 hours 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 (2002) 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 data sets, 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, $\delta\vec{x}$, 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:

$$J[\delta\vec{x}] = \frac{s\left(|\delta\vec{x}|l^{-1}\right)}{N_\alpha} \sum_{j=1}^{n_v} \alpha_j \sum_{i=1}^{n_o} \frac{\left\{H\left[\overline{F}(\vec{x}_i + \delta\vec{x})\right] - o_j(\vec{x}_i)\right\}^2}{\sigma_{i,j}^2} \qquad (1)$$

where o is the observation, $\vec{x}_i$ is the observation location, $\delta\vec{x}$ is the horizontal displacement vector, and $\sigma_{i,j}^2$ is the expected observation variance, which, in general, is a function of variable, data source, and height. The symbol $\overline{F}$ represents the forecast field smoothed by a 9-point filter in two dimensions; the smoothing is done to avoid fitting small-scale noise to the observations. H represents a transformation, if necessary, from the forecast variables to the observed quantity. Each variable is weighted according to α, which may 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 SOAR function:

$$s\left(\left|\delta\vec{x}\right|l^{-1}\right)=\frac{\exp\left(\left|\delta\vec{x}\right|l^{-1}\right)}{\left(1+\left|\delta\vec{x}\right|l^{-1}\right)} \tag{2}$$

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_α:

$$N_\alpha = \sum_{j=1}^{n\,\mathrm{var}} \sum_{i=1}^{nobs} \alpha_j \tag{3}$$

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" refers to the number of volumes in each of the x and y directions, and N_z the number of volumes counting in vertical (a total of $N_x \times N_x \times N_z$ 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, α, is detailed in Table 2.

The case of June 8, 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 spin-up 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. June 8, 1995 Synoptic Setting

June 8, 1995 is a major case for the Verification of Onset 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 June 8, 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 northeastern New Mexico. There was a dryline in the western Texas Panhandle, with the dew point temperature at Childress (CDS) of 24 °C contrasting with a dew point 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 on 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 (dew point of 23 ºC, 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 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 further into the cool air north of the frontal boundary, its intensity eventually decreased.

At 2008 UTC new cells formed further south; one was near the triple point, near Stinnett, Texas (STN) but two others developed in the warm air east of Amarillo, near Pampa (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 June 9, 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 dryline, advection of individual cells and interaction among the storms.

3. 12-km Mesoscale Spin-up 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 spin-up simulation is made using the Advanced Regional Prediction System (ARPS) nonhydrostatic model (Xue et al., 2001, Xue et al., 2000, Xue et al., 1995). Figure 3 shows the domain of the 12-km run with the 3-km domain imbedded. 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 mid-morning initialization is chosen for the spin-up. A 9-hour 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 vertical 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 satellite 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 ten-minute window. The data analysis and analysis increment updating are repeated every hour for 1600, 1700 and 1800 UTC, with the increments calculated at 10 minutes before the hour (generally corresponding to the time of the surface observations) and applied during the 10 minutes 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, dew-point 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 dew point 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-hr 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. 3-km Storm-scale Simulations

The phase correction procedure is applied using radar and surface data at the time of early storm growth in 3-km forecast initialized at 1800 UTC. The forecast The four experiments are: “Control”, no additional are used after the 12-km spin-up; “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; Figure 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):

$$R = 10\log_{10}\left\{1.73\times10^{4}\left(10^{3}\,\overline{\rho}q_{r}\right)^{7/4} + 3.8\times10^{4}\left[10^{3}\,\overline{\rho}(q_{s}+q_{h})\right]^{2.2}\right\}$$

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 $\overline{\rho}$ 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 northeasterly 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 ADAS_Only 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 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 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 one hour 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. Figures 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. Although the ADAS run started with reflectivity initialized further 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 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 (Note: due to a hardware problem on the Level-II data recorder at Amarillo, the 2110 UTC and subsequent radar images are produced from WSR-88D Level-III (NIDS) datasets). The main difference among the forecasts, shown in Fig. 18, is a break in the north-south line 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 hours after data time as non-linear 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 hours after the data time (Figures 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 southern end of the line in the eastern Texas Panhandle produced the damaging tornadoes. The forecasts also correctly forecast 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. 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 x 4-km grid.

The forecasts are compared by examining the rainfall bias and the equitable threat score. The rainfall bias is defined as:

$$Bias = \frac{\sum_{i,j} R_{fcst}(i,j)}{\sum_{i,j} R_{verif}(i,j)}$$

where the summation is done over all i,j grid points (excluding a frame of 5 points along the boundaries), R_{fcst} represents the forecast hourly precipitation, and R_{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 is measure of forecast skill defined as

$$ETS = \frac{H - Ch}{F + O - H - Ch},$$

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 Ch is an estimate of the number of points which could be correctly forecasted by chance, estimated by:

$$Ch = \frac{F \times O}{N_f} \quad ,$$

where N_f is the total number of points in the forecast domain. A perfect forecast would achieve an equitable 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 June 8, 1995 are scored. Hourly periods ending at 2100 UTC through 0000 UTC on June 9 were used. It should be noted that the model-accumulated rainfall for 2100 is actually a 50-minute rainfall, but because the storms were in an early growth stage from 2000-2010, it is likely that very little precipitation was missed in that 10-minute 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 two hours. Generally one expects a shortfall in model precipitation early in the run, due to model spin-up delays, but the 3-km pre-forecast period has apparently provided the necessary spin-up to develop precipitation, and that spin-up 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 June 8. 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 is calculated:

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

The threat scores, Fig 25, are better for the experiments with phase correction than without. The threat scores generally better here than the TS reported by Zhang (1999) with the ARPS model and diabatic initialization of another severe weather case. TS 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 hours even in the face of complex thunderstorm interactions. By 3 hours, 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 non-linear thunderstorm interactions will eventually compound to overcome improvements in the initial conditions.

It is also worth noting that the use of a 3-hour assimilation spin-up 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 and modifications to the application of latent heating in the model are being pursued at the time of this publication; these changes may improve the impact of the radar data in ADAS at this scale.

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 error functional, Eq. 1, from the initial value to its minimum could be used as a guide to the utility of the phase correction in any particular situation.

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 non-precipitating 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 non-precipitating 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.

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9. Figure Captions

FIG. 1 Station model and sea-level pressure for 1800 UTC 8 June 1995. Station model and sea level pressure analysis (contour interval 2 hPa) from the RUC model. Station model includes temperature (upper, °C) and dew point (lower, °C), wind barbs are in ms^{-1} with a full barb representing 5 ms^{-1} and half barb 2.5 ms^{-1}.

FIG. 2 Skew-T plot of sounding taken by National Severe Storms Laboratory (NSSL) mobile crew near Seiling, Oklahoma at 1800 UTC 8 June 1995. Temperature (°C) and dew point (°C) with parcel trajectory for unmixed surface parcel. Wind barbs are in ms^{-1} with a full barb representing 5 ms^{-1} and half barb 2.5 ms^{-1}.

FIG. 3. Domains for the 8 June 1995 simulations. Entire region shown is the domain for the 12-km forecast. Dashed box is the 3-km nested domain. Model terrain in meters above sea level. Axes length scale in km.

FIG. 4. Schematic of data assimilation process used for the 8 June 1995 demonstration.

FIG. 5 12-km grid-scale assimilated state at 1800 UTC, 8 June 1995. a) Surface temperature (°C), b) Dew-point temperature (°C), c) Mean sea level pressure (hPa) and wind barbs (ms^{-1}). Full barb is 5 ms^{-1}.

FIG. 6 Amarillo (KAMA), Texas, radar reflectivity (dBZ). 2008 UTC 8 June 1995. 0.5 degree elevation angle.

FIG. 7 Model reflectivity (dBZ) forecast for 2010 UTC 8 June 1995.

FIG. 8 Surface temperature (°C) and wind (barbs). Observations and model forecast at 2010 UTC 8 June 1995. Full barb is 5 ms^{-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 (1 grid length=3 km).

FIG. 10 Surface temperature (°C) and wind (barbs). Observations and ARPS fields after phase correction.

FIG. 11 Phase shift vectors for grid level 18. Vector scale in lower-left corner is in unit grid lengths (1 grid length=3 km).

FIG. 12 Model reflectivity (dBZ) at grid level 18 after phase shift applied.

FIG. 13 Reflectivity (dBZ) and wind (ms^{-1}), from the ADAS_Only analysis at 2010 UTC. Level 2 (10-m AGL) and Level 20.

FIG. 14 Reflectivity (dBZ) and wind (ms^{-1}) for ADAS analysis on phase-corrected forecast (Shift+ADAS) at 2010 UTC. Level 2 (10-m AGL) and Level 20.

FIG. 15 Amarillo (KAMA), Texas, radar reflectivity (dBZ), 0.5 degree elevation angle, 2038 UTC 8 June 1995.

FIG. 16 ARPS forecast fields of 10 m AGL winds (barbs) and reflectivity (dBZ) at 8 June 2040 UTC. a) Control, b) Shift, c) ADAS_Only, d) Shift+ADAS.

FIG. 17 Amarillo radar reflectivity (dBZ), 0.5 degree elevation angle, 2110 UTC 8 June 1995, about 1 hour after analysis.

FIG. 18 ARPS forecast fields of 10-m AGL winds (barbs) and reflectivity (dBZ) at 8 June 2110 UTC. a) Control, b) Shift, c) ADAS_Only, d) Shift+ADAS.

FIG. 19 Amarillo radar reflectivity (dBZ), 0.5 degree elevation angle, 2159 UTC 8 June 1995, about 2 hours after analysis.

FIG. 20 ARPS forecast fields of 10-m AGL winds (barbs) and reflectivity (dBZ) at 8 June 2200 UTC. a) Control, b) Shift, c) ADAS_Only, d) Shift+ADAS.

FIG. 21 Amarillo radar reflectivity (dBZ), 0.5 degree elevation angle, 2259 UTC 8 June 1995, about 3 hours after analysis.

FIG. 22 ARPS forecast fields of 10-m AGL winds (barbs) and reflectivity (dBZ) at 8 June 2300 UTC. a) Control, b) Shift, c) ADAS_Only, d) Shift+ADAS.

FIG. 23 Rainfall bias for June 8 case. Bold line is Control experiment, short dashed line is Shift, solid line is ADAS_Only, and long dashed line is Shift+ADAS.

FIG. 24 Rainfall Equitable Threat Score, 1 mm threshold. Line textures as in Fig. 23.

FIG. 25 Rainfall Equitable Threat Score, 1 mm threshold. Line textures as in Fig. 23.

Table 1. Test volume dimensions and data use for phase error detection.

Pass	i,j vol width	i,j over lap	N_x vol	k vol hgt	k over lap	N_x vol	Use U/A	Use Sfc	Use Radar
1	61	31	3	10	2	3	yes	yes	no
2	35	18	6	10	2	6	yes	yes	no
3	16	5	15	12	2	15	no	yes	yes
4	10	5	25	12	2	25	no	yes	yes

Table 2. Variable weighting assignments.

Variable	Weight, α
u-wind	5.0
v-wind	5.0
Pressure	0.

Potential Temp	5.0
Specific Humidity	3.0
Reflectivity	5.0
Radial Velocity	2.0