Assimilating Radar, Surface, and Profiler Data for the Sydney 2000 Forecast Demonstration Project

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ABSTRACT

A variational scheme for the analysis of low-level wind data is presented, and its performance during a recent field experiment is described. The analysis scheme finds an optimal fit to the data and a background field under the constraints of a dry boundary layer model. The scheme was run at the Sydney 2000 Forecast Demonstration Project and assimilated data from two Doppler radars, a surface mesonet, and a boundary layer profiler. With a few exceptions, the analysis system ran reliably over the 6-month period of the project, providing wind fields every 10 min. Described herein is the performance of the system in tracking a number of different phenomena, including sea breezes, a southerly change, and the low-level convergence associated with a severe tornadic hailstorm. Finally, the analyzed wind fields are verified using independent aircraft data and compared with winds calculated by the more traditional dual-Doppler approach.

1. Introduction

Over the last decade there has been a significant increase in the coverage of the lower atmosphere by operational Doppler radars. This is particularly true in the United States where the 1988 Doppler Weather Surveillance Radar (WSR-88D) network now provides extensive coverage over the contiguous United States. In addition, Terminal Doppler Weather Radars (TDWRs) have been installed at approximately 45 major airports across the United States and provide high-resolution radial velocity and reflectivity measurements of the lower troposphere. Doppler radar networks are also being installed in Europe as part of the Cooperation in the Field of Scientific and Technical Research Advanced Radar Systems project (COST75) (Collier 1998), while single operational Doppler radars have been installed in many other countries.

Although the observational capability has rapidly increased over the last decade, the ability to analyze this wealth of data and assimilate it into numerical models has not kept pace. There has been a number of research studies in the area of radar data assimilation (see e.g., Sun et al. 1991; Sun and Crook 1994; Weygandt et al. 2002; Montmerle et al. 2001). Furthermore, groups at research laboratories such as the National Oceanic and Atmospheric Administration (NOAA) Forecast Systems Laboratory (McGinley et al. 2000) and the Center for the Analysis and Prediction of Storms at the University of Oklahoma (Xue et al. 2000; Brewster 1996) have been examining methods to assimilate radar data into their prototype models. However, so far, the assimilation of radar data in operational models has generally been limited to bulk, integrated products such as velocity azimuth display (VAD) and reflectivity-derived cloud and precipitation fields.

As a step toward the operational assimilation of full-volume radar datasets, we have been developing a real-time system, the Variational Doppler Radar Analysis System (VDRAS), that assimilates low-level radar data. The system has been under development for approximately 5 yr and has been run operationally for the past 2 yr at the Weather Forecast Office in Sterling, Virginia, using data from a single WSR-88D radar. The single-Doppler assimilation system is described in detail in Sun and Crook (2001). Recently, we enhanced the system to assimilate data from multiple Doppler radars. This system was run at the Bureau of Meteorology in Sydney, Australia, as part of a Forecast Demonstration Project (FDP). The Sydney 2000 FDP was a pilot project of the World Weather Research Program (WWRP) aimed at demonstrating and comparing a number of different short-term forecast (or nowcast) schemes in support of the 2000 Summer Olympics. For the Sydney 2000 FDP, the variational analysis system assimilated data from two Doppler radars spaced approximately 50 km apart as well as observations from an upper-level profiler and a surface mesonet.

Herein, we describe the system for variational assimilation of multiple Doppler observations (section 2). In section 3 we describe the data sources that were available...
for the Sydney 2000 FDP, and present four case studies from the FDP in section 4. We verify the analysis results in section 5 and give some conclusions in section 6.

2. Analysis method

The real-time system for assimilation of single-Doppler information is described in detail in Sun and Crook (2001). In this paper we will briefly summarize the system and discuss the enhancements made for the Sydney 2000 FDP, for which data from two Doppler radars were available.

The assimilation scheme finds a solution to the equations of motion that fits the data and a background field as closely as possible. The numerical model used in the real-time system consists of the dry Boussinesq equations of motion (Sun et al. 1991). [A moist version of the assimilation system exists (see Sun and Crook 1997), but it is computationally too expensive to run in real time.] The equations are discretized on a Cartesian grid (i.e., with flat terrain) that has constant grid spacing in the vertical and horizontal.

The optimal fit is achieved by reduction of the following cost function:

$$ J = J_o + J_a + J_p, $$

where $J_o$ is the discrepancy from the radar data (radial velocity and reflectivity), $J_a$ the discrepancy from a background field, and $J_p$ the penalty terms.

The term $J_o$ is defined by

$$ J_o = \sum_{\text{radar time}} \sum_{\text{data grid}} (Hx - y_o)^T \mathbf{O}^{-1} (Hx - y_o), $$

where $y_o$ is the data, $x$ the model variables, $\mathbf{O}$ the observational error matrix (assumed diagonal in this work), and $H$ the forward operator that maps the model variables on the model grid to the data variables on the data grid. (The structure of the data grid is described in more detail in section 3.) In our radar data assimilation system, $H$ has two components. The first component maps model variables onto the data grid by taking a weighted average of the model variables in a radar beam; that is,

$$ v_{r,e} = \frac{\sum Gv \Delta z}{\sum G}, $$

where $G$ is the power gain in the radar beam, $v_{r,e}$ is the observed radial velocity, $v$ is the model radial velocity, and $\Delta z$ is the model’s vertical grid spacing. The second component of $H$ maps the model Cartesian velocity components ($u$, $v$, $w$) onto the model radial velocity $v$, and is given by

$$ v_r = \frac{x - x_o}{r} u + \frac{y - y_o}{r} v + \frac{z - z_o}{r} w, $$

where $r$ denotes the distance between a grid point ($x$, $y$, $z$) and the radar location ($x_o$, $y_o$, $z_o$).

The second term in the cost function, $J_a$, is the discrepancy from a background field. The background field is important for providing the lateral boundary conditions, filling regions of missing radar data, and providing a first guess for the minimization problem. Typically, the background field is some form of large-scale field (analysis or forecast) in which the errors between grid points are correlated. For the Sydney 2000 FDP a number of options for the background field presented themselves. These included 1) using a forecast field from an operational model such as the Australian Local Analysis and Prediction System (LAPS) model, 2) using a forecast initialized from the previous VDRAS analysis, 3) using the previous VDRAS analysis, and 4) performing our own analysis of large-scale data. Option 1 would have required a significant amount of setup time that our short deadline did not allow, and options 2 and 3 were not possible because of problems with data quality (discussed in the next section). We thus chose to perform our own analysis of some of the larger-scale datasets that were available during the Sydney 2000 FDP. The analysis was obtained by combining mesonet observations at the surface with profiler data above. The surface observations were first interpolated to the model grid using a Barnes analysis scheme (Barnes 1964) and then merged with the profiler data above by using a local linear least squares fit. The background error covariance model that is used is pseudo-Gaussian in the horizontal and is described in Sun and Crook (2001). As will be shown, by treating the larger-scale data (surface and profiler) separately from the higher-resolution radar data we are able to maintain the details contained in the radar data while fitting to the background field in regions without radar data.

The last term $J_p$ in the cost function is the temporal and spatial smoothness term and is described in Sun and Crook (2001). The weighting coefficients in the penalty terms are determined by trial and error and depend to a certain degree on the amount of smoothing that is required.

The cost function in (1) is minimized using a limited-memory quasi-Newton iterative procedure (Liu and Nocedal 1988). For each iteration, the numerical model is integrated forward and the cost function computed. The adjoint model is then integrated backward and the gradient of the cost function is obtained. We have found that when using real data the cost function reduction typically levels out after 30 iterations. We thus decided to terminate the minimization procedure after 40 iterations.

3. Data sources

Figure 1 shows the observational network that was available for the Sydney 2000 FDP. The data sources that were used in the analysis scheme were the following:
Fig. 1. Observational network for the Sydney 2000 FDP. Shown are the locations of the Doppler radars (C-Pol and Kurnell), the associated dual-Doppler lobes, the 54.1-MHz profiler at the Sydney airport, and the surface mesonet stations. The western and southern boundaries of the analysis domain are shown by the dashed lines. The eastern boundary is 10 km to the right of the figure’s edge, while the northern boundary is 10 km above the top of the figure.

1) C-band Doppler radar at Kurnell near the Sydney airport: beamwidth 1.0°; 11 elevation angles (0.7°, 1.5°, 2.5°, 3.5°, 4.5°, 5.5°, 6.9°, 9.2°, 12.0°, 15.6°, and 20.0°); volume scan rate 5 min.

2) C-band polarimetric Doppler radar (C-Pol), located approximately 48 km west-northwest of Kurnell: beamwidth 1.0°; 11 elevation angles (0.6°, 1.5°, 2.4°, 3.3°, 4.2°, 5.1°, 6.3°, 7.8°, 9.2°, 10.6°, and 13.4°); volume scan rate 10 min.

3) 29 mesonet stations within a domain of 150 km × 150 km (locations shown in Fig. 1): wind measurement height 10 m; update rate between 1 and 30 min.

4) A 54.1-MHz profiler located at the Sydney airport for specification of the upper-level flow: update rate 15 min.

The analysis procedure is conducted in the following steps:

1) Data collection: All data sources that are valid within the assimilation time window of 10 min are collected.

2) Data preparation: Some of the quality control (QC) that is performed within VDRAS is described in Sun and Crook (2001). During the Sydney 2000 FDP, two additional problems arose with data quality.

   (i) Unfolding of C-Pol radial velocity: C-Pol was run with a Nyquist velocity of 12.5 m s⁻¹. Folded velocities occurred often during the FDP. If the region of aliased velocities was small (less than ~10 km in width), then the adjoint system was relatively unaffected by the contamination. However, larger areas of folding did cause significant problems. An unfolding algorithm was run on both radar datasets; however, there were times when it failed to unfold aliased velocities and other times when it unfolded correct velocities. In order to prevent the effects of folded velocity data from propagating from one analysis cycle to the next, we decided to turn off the cycling procedure and use the mesoscale analysis as a background field (rather than the analysis from the previous cycle).

   (ii) Sea clutter: The Kurnell radar, which is located close to the coast, was often contaminated by sea clutter. One of the signatures of sea clutter is a rapid decrease with height of the reflectivity field. To remove sea clutter contamination, data over the
ocean that showed a decrease of 10 dBZ between the first and second tilts were removed.

After QC, the data from both radars are interpolated to the model Cartesian grid in the horizontal but left on their original constant elevation surfaces in the vertical. The justification for this processing procedure is that the poorest sampling resolution of the radar data (approximately 2 km at 120 km, which is the farthest range distance in the analysis domain) is still better than the horizontal resolution of the model grid (2.5 km). In the vertical, however, the data resolution varies from being much finer than the model grid (375 m) near the radar to much coarser far from the radar near the domain boundaries. Figure 2 is a schematic diagram showing a vertical cross section through the analysis domain, along with the various data sources available (radar, mesonet, and profiler). Close to the radar, the beam (which typically has a 1\° width) may encompass one or no analysis grid points. Far from the radar, three or four grid points may fall within a radar beam and hence get averaged through Eq. (3) to provide the model estimate of the radar observation.

The grid points labeled A and B in Fig. 2 indicate a region where particular care must be taken during the assimilation procedure. Both grid points are on the same horizontal level, but gridpoint values at B are compared with data from the lowest radar beam, whereas values at A are compared with data from the radar beam above. If significant vertical wind shear exists, this can result in the generation of anomalous horizontal wind shear (i.e., convergence/divergence) located at a fixed distance from the radar. This problem can be mitigated by improving the error statistics of the observations and background fields. It can also be reduced by artificially broadening the beams so that they overlap to a certain extent. However, the problem cannot be entirely eliminated and is a natural result of converting from a spherical to a Cartesian coordinate system.

1) Calculation of the mesoscale analysis: The next step in the analysis procedure is to calculate the background field from the surface observations and upper-level profiler data. The surface observations are first analyzed to the model grid using a Barnes interpolation scheme. The radius of influence in the scheme was set at the average station spacing (~30 km). Above the surface, the background field is obtained by applying a local-linear least squares fit to the surface value and the profiler data above. In Sydney, the lowest level of usable profiler data was typically at 600 m or above.

2) Assimilation of the radar data: The final step in the analysis is the assimilation of the Doppler radar data. As already noted, this is achieved by minimizing the cost function [Eq. (2)] by running the numerical model forward and the adjoint of the model backward. A relative weight of 0.1 was used for the background term compared to the observational term; in other words, the standard deviation of the background field error was assumed to be a factor of 10 larger than the observational error. This factor was obtained by comparing typical rms errors in our analyses (the background field) with typical observational errors in radial velocity. A comparison of our VDRAS analyses in Sterling, Virginia, against independent aircraft data (which admittedly has its own sources of error) indicated an average rms difference of 3–4 m s\(^{-1}\). Observational error in radial velocity for typical values of spectrum width and signal-to-noise ratio in the boundary layer are of the order of 1 m s\(^{-1}\) (see Fig. 6.6 of Doviak and Zrnić 1984). Hence, we estimate that the ratio of background variance to observational variance is of the order of 10. Clearly, more research is needed in the area of background/observational error estimation at these scales, and we are actively working in this area.

For the field test, the system was run on a DEC Alpha workstation at the Sydney Bureau of Meteorology. In order to run in real time, it is essential that the algorithm complete its calculations faster than the time window used for assimilation. For the Sydney system, we used a time window of 10 min, which typically captured two volumes from C-Pol and three volumes from Kurnell. Consequently, we tuned the size and resolution of the analysis domain so that 40 iterations could be completed in just less than 10 min of CPU time. This resulted in a domain size of 120 km × 120 km × 3 km, with a horizontal resolution of 2.5 km and vertical resolution of 375 m.

4. Case studies

With a few exceptions, the analysis system ran continuously from mid-August 2000 until February 2001.
The exceptions were primarily caused by interruptions in the data flow and problems with folded velocity data from C-Pol. In this section we will describe four cases that occurred during the FDP and show the real-time VDRAS analyses. In the following section we will perform a more quantitative verification analysis of the performance of VDRAS.

**a. 8 October southerly change event**

During the afternoon of 8 October 2000, a southerly change moved through the Sydney area. The leading edge of the southerly change is shown by the thin line of reflectivity observed by C-Pol (see Fig. 3). The analysis of surface data at 1430 local time (0330 UTC) is shown in Fig. 3a. The contours depict the horizontal convergence (contour interval \(10^{-4} \text{ s}^{-1}\); only regions of convergence are shown). The surge can be made out in the surface analysis; however, the center of the convergence maximum is approximately 20 km ahead of the radar-observed convergence line. This discrepancy is due to the coarseness of the surface observations (average spacing \(\sim 30 \text{ km}\)). Furthermore, the maximum convergence analyzed is only \(2.5 \times 10^{-4} \text{ s}^{-1}\).

Figure 3b shows the VDRAS analysis at the lowest grid level \(z = 180 \text{ m}\) using radar, surface, and profiler data. The convergence now maximizes along the leading edge of the surge. The convergence maximum is not continuous along the fine line but is concentrated in two locations, northeast and south of C-Pol. A secondary maximum occurs to the north of C-Pol and is associated with a weak surge ahead of the main southerly change. The convergence along the fine line is significantly stronger in this analysis, reaching a maximum of \(1 \times 10^{-3} \text{ s}^{-1}\), which is four times larger than in the surface analysis. Although the wind speeds behind the southerly change are enhanced somewhat (from \(\sim 7\) to \(\sim 8 \text{ m s}^{-1}\)) the convergence increase is primarily due to the higher resolution of the radar data compared to the surface network.

**b. 18 September sea-breeze case**

Figure 4 shows the low-level winds and convergence field on 18 September 2000, 1634 LT (Fig. 4a), 1734 LT (using all data) (Fig. 4b), and 1734 LT (Fig. 4c) (using just surface data). The low-level (0.6°) reflectivity from C-Pol is also plotted and shows a thin line associated with the sea breeze over Sydney. Typically the sea breeze would first become evident in C-Pol data along the coastline south of Sydney. Over time a bulge would often form in the sea breeze as it moved inland over the lower terrain of Sydney but was held back by the higher terrain north and south of Sydney. On this

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1 The maximum convergence in the full VDRAS analysis is also six times larger than the VDRAS analysis using just surface and profiler data. The maximum convergence in the latter analysis is reduced compared to the surface analysis since there is no horizontal variation in the profiler data.
The 18 Sep 2000 sea-breeze case. (a) VDRAS analysis at 1644 LT, (b) VDRAS analysis at 1744 LT, and (c) surface analysis at 1744 LT. The contour interval for horizontal convergence field is $2.0 \times 10^{-4} \text{s}^{-1}$.

day, the leading edge of the sea breeze moved inland at approximately $2.5 \text{ m s}^{-1}$.

The analysis system has captured the sea-breeze flow along with the horizontal convergence at the leading edge of the breeze. The convergence field (with a contour interval of $2 \times 10^{-4} \text{s}^{-1}$) has been plotted and closely tracks the position of the sea breeze at both analysis times. The convergence reaches a maximum of $1.5 \times 10^{-3} \text{s}^{-1}$ along the leading edge of the sea breeze at the latter analysis time.

The C-Pol reflectivity data also shows an east–west-oriented convergence line in the westerly flow ahead of the sea breeze. This convergence line, which was occasionally observed during the Sydney 2000 FDP, appears to be a result of the interaction of a westerly flow with the higher terrain to the west of Sydney. On this day, the westerly flow ahead of the sea breeze was quite strong ($6$–$7 \text{ m s at } z = 180 \text{ m}$). Note that the analysis system has picked up some of the convergence associated with the fine line ahead of the sea breeze.
c. 3 October sea-breeze case

A second sea-breeze event (3 October 2000) is shown in Fig. 5. For this case, the winds ahead of the sea breeze were light and variable. The sea breeze first becomes evident in the C-Pol data south of Sydney (Fig. 5a) at 1450 LT. Three hours later, the sea breeze has moved approximately 50 km inland and again has developed a similar bulge shape to that seen on 18 September. The speed of propagation of the leading edge of the sea breeze on this day was approximately 3 m s\(^{-1}\).

The low-level convergence analyzed along the sea breeze was less on this day compared to 18 September presumably because of the reduced flow ahead of the sea breeze. At 1450 LT, a convergence maximum of 2.5 \(\times\) 10\(^{-4}\) s\(^{-1}\) is analyzed along the convergence line. At 1740 LT, it has increased to 5.0 \(\times\) 10\(^{-4}\) s\(^{-1}\).

d. 3 November hailstorm and tornado case

On 3 November 2000, a tornadic hailstorm developed south of Sydney and moved north-northeastward. The hailstorm’s direction of motion was more northerly than most of the storms on that day, as it followed the boundary layer forcing produced at the intersection of the sea breeze with the storm’s outflow. The position of the sea breeze and outflow are shown in Figs. 6a,b at 1450 and 1540 LT, respectively. These positions were subjectively analyzed in real time by a forecaster at the Bureau of Meteorology in Sydney. The hailstorm can be seen just to the south of this intersection point in the 0.6° reflectivity from C-Pol (grayscale) in both Fig. 6a and 6b. The analyzed wind vectors and horizontal convergence are also overlaid. As can be seen, strong low-level convergence has been analyzed at the intersection point of the gust front with the sea breeze. This convergence “bull’s-eye,” which reached a maximum of 1.1 \(\times\) 10\(^{-3}\) s\(^{-1}\), successfully tracked the motion of the gust front–sea-breeze intersection point as it moved northward.

5. Verification

a. Fit to the data

In the previous section we presented four case studies to illustrate the performance of the analysis system under various flow situations. In this section, we examine more quantitatively the performance of the system. We first examine the fit of the analysis to the input radar data. The average rms difference between the analysis and input radial velocity for both radars is given in Table.
1. The average was calculated over four consecutive days (29 October–1 November) when there was no contamination from folded velocities. The average rms difference from Kurnell (C-Pol) was 1.54 (1.91) m s$^{-1}$. The slightly better fit to the Kurnell data is most likely due to the fact that Kurnell was operated at a higher signal-to-noise ratio.

b. Comparison with aircraft data

Although Table 1 indicates that the analysis fits quite closely to the input data, it is not an independent test of the accuracy of the analysis. As an independent test, we use wind observations from aircraft taking off and landing at Sydney airport. Table 2 shows the mean vector difference (MVD) between the VDRAS and aircraft-observed wind vector for flights on the 4 days described in section 4 (18 September, 3 October, 8 October, and 3 November). (The MVD is the average magnitude of the difference vector between the analyzed and observed winds.) Observations from aircraft both taking off and landing at Sydney airport were used in the comparison. For takeoffs, the aircraft typically pass through the top of the analysis domain (3 km) at a short horizontal distance (~20 km) from the airport and thus does not provide a great deal of information about the horizontal variability in the wind field. However, landing aircraft enter the domain at distances of 60 km from the airport and hence can sample some of the horizontal variability. If an aircraft observation is available at the analysis level shown and within 10 min of the valid time, it is shown as an overlay on the figure (by way of a vector and text giving the wind speed and direction).

Table 2 shows that the mean vector difference ranges from 2.1 to 3.5 m s$^{-1}$ while the mean aircraft-observed wind speed ranges from 4.5 to 9.0 m s$^{-1}$. Calculated over the 4 days, the MVD is 2.6 m s$^{-1}$, while the mean wind speed is 6.3 m s$^{-1}$. We consider this to be a reasonable agreement, given that the aircraft-observed wind also contains observational error and is a line average along the aircraft’s flight track, whereas the analysis wind is an average over a model grid volume.

c. Comparison with dual-Doppler winds

Finally, we compare the analyzed fields with dual-Doppler calculated winds. If we assume that the elevation angle of the radar beam is small, then the dual-Doppler winds are calculated. The dual-Doppler winds are used to verify the analysis fields.

<table>
<thead>
<tr>
<th>Date</th>
<th>Mean vector difference (m s$^{-1}$)</th>
<th>Mean wind speed (m s$^{-1}$)</th>
</tr>
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<tbody>
<tr>
<td>18 Sep 2000</td>
<td>2.1</td>
<td>6.2</td>
</tr>
<tr>
<td>3 Oct 2000</td>
<td>3.5</td>
<td>9.4</td>
</tr>
<tr>
<td>8 Oct 2000</td>
<td>2.6</td>
<td>4.5</td>
</tr>
<tr>
<td>3 Nov 2000</td>
<td>2.2</td>
<td>5.0</td>
</tr>
<tr>
<td>Average</td>
<td>2.6</td>
<td>6.3</td>
</tr>
</tbody>
</table>
Doppler winds \((u_{\text{dual}}, v_{\text{dual}})\) can be calculated directly from
\[
\begin{align*}
 u_{\text{dual}} &= \frac{v_{1} \cos(a_{2}) - v_{2} \cos(a_{1})}{D} \\
 v_{\text{dual}} &= \frac{v_{2} \sin(a_{1}) - v_{1} \sin(a_{2})}{D},
\end{align*}
\]
where \(v_{1} \) (\(v_{2} \)) is the radial velocity from the first (second) radar, \(a_{1} \) (\(a_{2} \)) is the azimuth from the first (second) radar and \(D = \sin(a_{1}) \cos(a_{2}) - \cos(a_{1}) \sin(a_{2})\). The calculations were performed on a Cartesian coordinate system by first interpolating the radial velocities from constant elevation to constant height surfaces.

Figure 7a shows the dual-Doppler winds (black) overlaid on the VDRAS winds (gray) at 1520 LT on 3 November at a height of \(z = 560\) m. Winds were only calculated in regions where the difference in azimuth angles was between 30° and 150°. The first point to note is the limited areal extent of the dual-Doppler winds, which cover less than 5% of the total analysis domain. (Note that the domain plotted in Fig. 7 represents only 25% of the total analysis domain.) This is partly due to the limited extent of the dual-Doppler lobes but also due to the limited overlap of usable data from both radars.

In comparing the two wind fields it should first be noted that the dual-Doppler winds fit the radial velocity data exactly [since there are two equations for two unknowns \((u_{\text{dual}}, v_{\text{dual}})\)] and, hence, do not account for any uncertainty in the radial wind measurements. In contrast, the VDRAS analysis is an optimal fit between the radar data and the background field and takes into account the expected error in both fields. The analysis also fits the data over a number of time levels under the constraints of a numerical model which further smooths the fields. For the 3 November case, the comparison of the two wind fields is reasonably good, particularly in terms of the wind direction. The dual-Doppler winds speeds are slightly greater than the VDRAS speeds, which is primarily due to the inclusion of the background field in the latter analysis. At low levels, the background field is dominated by the surface winds, which will typically underestimate the winds in the boundary layer above. For this case, the rms difference between the dual-Doppler and VDRAS winds is 1.4 m s\(^{-1}\) for the east–west component and 0.8 m s\(^{-1}\) for the north–south component.

Figure 7b shows the two wind fields for the 8 October case at 1425 LT. In this case the comparison is not as good, particularly in the southern dual-Doppler lobe. The rms difference between the two wind fields is 2.8 m s\(^{-1}\) in the east–west component and 2.2 m s\(^{-1}\) in the north–south component. Again, this difference is due to the fact that the background analysis typically underestimates the wind speeds in the boundary layer.

The above comparison may suggest that the inclusion of a background field can be detrimental to the wind analysis. This may be true at those grid points in the dual-Doppler lobes that have radial wind observations from both radars with small observational error. If the observational and background errors were known accurately, then the incorporation of the background field would improve the analysis. However, the true errors are not known and must be estimated. In the real-time system we set the relative weight of the background to the observational term to a factor of 0.1, which is prob-
The analysis system produced realistic wind fields with a few exceptions, the analysis system ran reasonably well during the Sydney 2000 Forecast Demonstration Project. The analysis scheme finds an optimal fit to the surface observations. Studies are being conducted to improve the estimation of the observational and background error statistics. We are also examining methods to reduce the anomalous generation of horizontal convergence in the conversion of the radar data from spherical to Cartesian coordinates.

Finally, we plan to examine the skill of the numerical model and assimilation scheme in predicting the boundary layer flow. Results from a gust front case in Sterling, Virginia, suggest that the boundary layer model is capable of producing forecasts that improve over persist-

<table>
<thead>
<tr>
<th>Radar</th>
<th>Coverage over analysis domain</th>
</tr>
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<tbody>
<tr>
<td>C-Pol</td>
<td>29%</td>
</tr>
<tr>
<td>Kurnell</td>
<td>18%</td>
</tr>
<tr>
<td>Both radars</td>
<td>14%</td>
</tr>
<tr>
<td>Both radars in dual-Doppler lobes</td>
<td>6%</td>
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</tbody>
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tence (Sun and Crook 2001). In order to reliably predict the motion of gust fronts and convergence lines it is necessary to retrieve the low-level buoyancy field that drives these phenomena. In a dry model, the retrieval of the buoyancy field depends on an accurate estimation of the convergence field and its temporal evolution. Hence, improvements in the estimation of the convergence field should lead to improvements in the skill at forecasting gust fronts and convergence lines. In a moist model, evaporative cooling can significantly affect the low-level buoyancy field. Thus, as a first step toward the implementation of a full cloud model, we are planning to include an evaporative cooling term in the current real-time system, which should lead to an improvement in the estimation of the buoyancy field.

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