Proceedings of ERAD (2004): 487–493 © Copernicus GmbH 2004

ERAD 2004

On the assimilation of Doppler radial winds into a high resolution NWP model

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Abstract. An approach to the assimilation of Doppler radar radial winds into the UK Met Office operational forecast model is described. We discuss the types of errors which might occur in radar radial winds. The construction of simulated high resolution radar data containing such errors is described. Examples of these data are presented. Such a data set will be used within an Observing System Simulation Experiment (OSSE) to investigate the impact of various assumptions made in the assimilation procedure. Examples of radial winds derived from the Chilbolton S-band radar located in central southern England and the Met Office operational model implemented at a 1 km resolution are discussed.

1 Introduction

Assimilation of Doppler radar wind data into atmospheric models has recently received increasing interest. This is because of the increasing use of limited area high resolution numerical models for weather prediction. The models require observations with high spatial and temporal resolution for determining the initial conditions, for which purpose radar data are particularly appealing.

Over the last thirty years or so networks of weather radars, providing measurements of radar reflectivity from which rainfall has been estimated, have been established within operational observing systems. Initially the radars, operating at s-band (10 cm) or c-band (5–6 cm) wavelengths, did not have the capability to measure the motion of the targets (mainly hydrometeors but also insects and birds, and, for high power systems, refractive index inhomogeneities) towards or away from the radar site. During the last twenty years weather radars having Doppler capability measuring radial motion of the targets have become standard such that now in Europe well over half of the operational radars are Doppler systems (Collier, 2001).

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Considerable effort has been, and continues to be, put into the development of nowcasting techniques based upon the extrapolation of radar reflectivity fields aimed at generating forecasts of precipitation up to 3–6 h ahead (for review; see Collier, 1996; Krzysztofowicz and Collier, 2004). Whilst such systems have met with some success, particularly when incorporating wind fields from mesoscale numerical models (Golding, 1999), they are not appropriate for forecasting to longer lead times. Improvements to forecasts for these lead times are now being sought through the assimilation into mesoscale models of radar reflectivity using latent heat nudging methods (Macpherson et al., 1996) and variational techniques in which model “reflectivity” is compared with actual measured reflectivity (Sun and Crook, 2001). More recently Doppler radar radial winds have also been assimilated into NWP models as vertical wind profiles derived from Velocity Azimuth Display (VAD) analysis (Lindskog et al., 2002; Benjamin et al., 2004) and using variational techniques (Sun and Crook, 1997, 1998).

In this paper we outline the likely errors in estimates of Doppler radar radial winds. We describe their representation as part of a system for generating simulated data for use in Observing System Simulation Experiments (OSSEs) to be carried out using the Met Office operational Unified Model (UM) forecast system.

2 Errors in the determination of radial wind speed by Doppler radar

Targets moving away from or towards a radar produce a Doppler shift between the frequency of the transmitted signal (pulse), and the signal reflected from the targets and received back at the radar. However, ambiguities may arise in these measurements due to range folding and velocity aliasing (see Doviak and Zrnic, 1993). Fortunately procedures have been developed to remove these problems (see for example Gong et al., 2003).

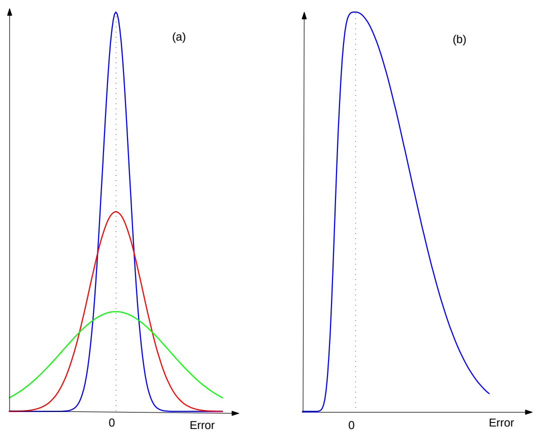

Fig. 1. Distribution of wind speed error due to (a) instrumental noise, and (b) strong velocity gradient across the pulse volume.

Other problems remain, namely the existence of data holes (where there are no targets), and irregular coverage, instrumental noise and sampling errors. Various types of interpolation schemes have been used to fill in data holes and poor coverage (see for example Lin et al., 1993), although such schemes are unnecessary when three dimensional assimilation schemes are implemented. However, the impacts of instrumental noise and sampling are more problematic.

May et al. (1989) discuss, and assess, a number of techniques used to estimate the Doppler shift in the received signals. The Doppler shift is proportional to the slope of the phase of the autocorrelation function (at zero lag) of the returned signals. An estimator of the shift is the phase at the first lag divided by the value of the lag in time units. This is known as pulse pair processing, and may be improved by averaging more than one value of the phase divided by the lag (poly pulse pair).

An alternative approach is to estimate the Doppler shift directly from the first moment of the Doppler spectra (Doviak and Zrnic, 1993) perhaps using a maximum likelihood estimator (similar to a least squares fit) of the logarithmic spectral signal. A further technique is possible based upon the analysis of the power spectrum, its circular convolution and Fast Fourier Transform (FFT) of the same. Interestingly, it was concluded by May et al. (1989) that the major limitation to the radar performance is the small-scale variability of the wind across the pulse volume. Therefore there is little to be gained by using complicated algorithms to estimate the Doppler shift. The width of the Doppler spectrum, usually assumed to be Gaussian, determines the correlation time of the signal, which is inversely proportional to the spectral width. Hence the spectrum width is directly related to the error in the measurements.

Instrumental errors may be reduced by selecting measurements at range intervals somewhat longer than the radar range gate resolution (Keeler and Ellis, 2000). For example, Xu and Gong (2003) selected data every 1 km along each radar beam for a radar having a range gate resolution of 250 m.

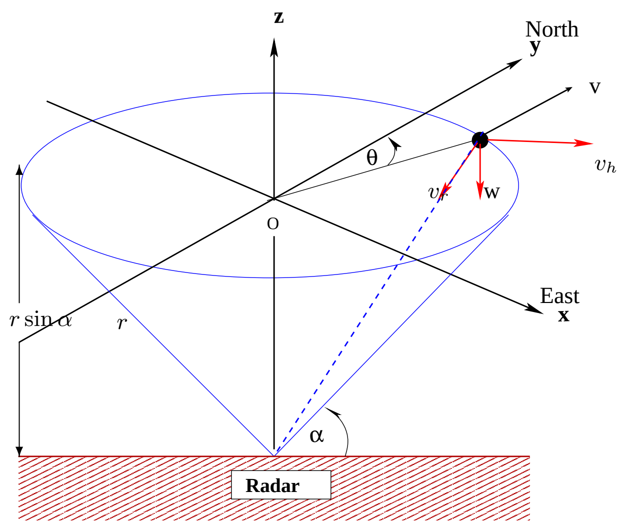

Fig. 2. Geometry for scan of velocities on a VAD circle

Sampling errors depend upon the size of the pulse volume corresponding to each data point. The Chilbolton radar in central South England has a pulse volume of about 300 m by 0.25 degree, and, even though this is small, even smaller scale wind variability may introduce different sampling errors from measurement point to measurement point. In practice the sampling errors could be weakly correlated from point to point, but only a very small additional error will be introduced if this is ignored. Practically, sampling errors dominate since instrumental errors are usually minimized in operational systems. However, in what follows we outline a system for creating artificial radar radial wind data sets within which different types of error may be included. Figure 1 shows schematics of the impact upon a Gaussian Doppler spectrum of various effects for example strong wind shear across the pulse volume, and instrumentally-induced effects. Several of these effects upon the Doppler spectrum may be present in the same radar image, and, in the case of geophysically-induced effects, their magnitude may vary with range and azimuth. The height and size of the pulse volumes will increase with increasing distance from the radar.

3 Simulation model of Doppler radar radial wind fields

The construction of artificial radar data sets has been carried out for several studies over the last twenty years or so (see for example Saarikivi, 1987; May et al., 1989; Xu and Gong, 2003). Consider a conical radar scan (Fig. 2) in a Cartesian coordinate system (x, y, z). The components of the wind field corresponding to these coordinates are u, v and w respectively.

It is assumed that velocity-range folding has been removed. The wind is assumed to vary with height according to an Ekman spiral with variable surface friction. The wind direction at the top of the boundary layer is parallel to the isobars, whilst the wind direction at surface is in the direction of the lower pressure due to the surface friction, the coriolis force and pressure gradient force. Here

$$u = U_g(1 - e^{-az}\cos az), \quad (1)$$

$$v = U_g e^{-az}\sin az, \quad (2)$$

where U_g is the geostophic wind, $a = \sqrt{f/2k}$, f is the coriolis parameter and k is the eddy-exchange coefficients ($\approx 5 \times 10^4\,\mathrm{cm}^2\,\mathrm{s}^{-1}$) in middle latitudes.

The simulated data are assumed to be available on the measurement points. The radial velocity is calculated from

$$v_r = u\cos\alpha\sin\theta + v\cos\alpha\cos\theta + w\sin\alpha \quad (3)$$

where α is the elevation angle and θ is the azimuth angle of the radar beam.

At each measurement point a Gaussian (or a modification of a Gaussian) distribution is introduced, the magnitude and spatial variation of which may represent the different error types (Fig. 1). Examples of the type of radial velocity field so-produced are displayed in Fig. 3. Such data can be used to test variational analysis schemes.

4 The Met Office UM and 3D-Var assimilation system

We intend to assimilate radial doppler wind data into versions of the limited area form of the Met Office Unified Model (nonhydrostatic formulation derived from Cullen et al., 1997) which is currently run operationally at 12 km with a 3D-Var analysis system based on Lorenc et al. (2000).

3D-Var has been used operationally to assimilate radar wind information in the form of VAD wind profiles (see Parrish and Purser, 1998). Recently, 3D-Var, for the Swedish Meteorological and Hydrological Institute (SMHI) has been extended for assimilating Doppler radar wind data either as radial super-observations[1] or as VAD wind profiles (see Gustafsson et al., 2001; Lindskog et al., 2002).

The Met Office 3D-Var system uses an incremental formulation. Under the assumption that the background and observation errors are Gaussian, random and independent of each other, the optimal estimate of the Cartesian wind $\mathbf{x}_a = \mathbf{x}_b + \delta\mathbf{x}$ in the analysis space is given by the incremental cost function,

$$\mathcal{J}[\delta\mathbf{x}] = \frac{1}{2}\delta\mathbf{x}^T\mathbf{B}^{-1}\delta\mathbf{x} + \frac{1}{2}[\mathbf{H}\delta\mathbf{x} - \mathbf{y} + \mathcal{H}\mathbf{x}_b]^T\mathbf{E}^{-1}[\mathbf{H}\delta\mathbf{x} - \mathbf{y} + \mathcal{H}\mathbf{x}_b], \quad (4)$$

where $\delta\mathbf{x} \equiv \mathbf{x}_a - \mathbf{x}_b$ is the state vector of the analysis increments (the estimated radial winds is given by $\mathcal{H}\delta\mathbf{x} + \mathcal{H}\mathbf{x}_b$), $\mathbf{x}_b$ is the state variable of the background Cartesian winds, and $\mathbf{y}$ denotes the observed radial winds in the observation space. $\mathcal{H}$ is the nonlinear observation operator that relates the model variables to the observation variable and a transformation between the different grid meshes, and $\mathbf{H}$ is the *linear* observation operator with elements $h_{ij} = \partial\mathcal{H}_i/\partial x_j$. Some constructions of the background and observation error covariance matrices **B** and **E** are given in Lorenc (1997).

[1] *Super-observations* are spatial averages of raw measurements with different resolutions.

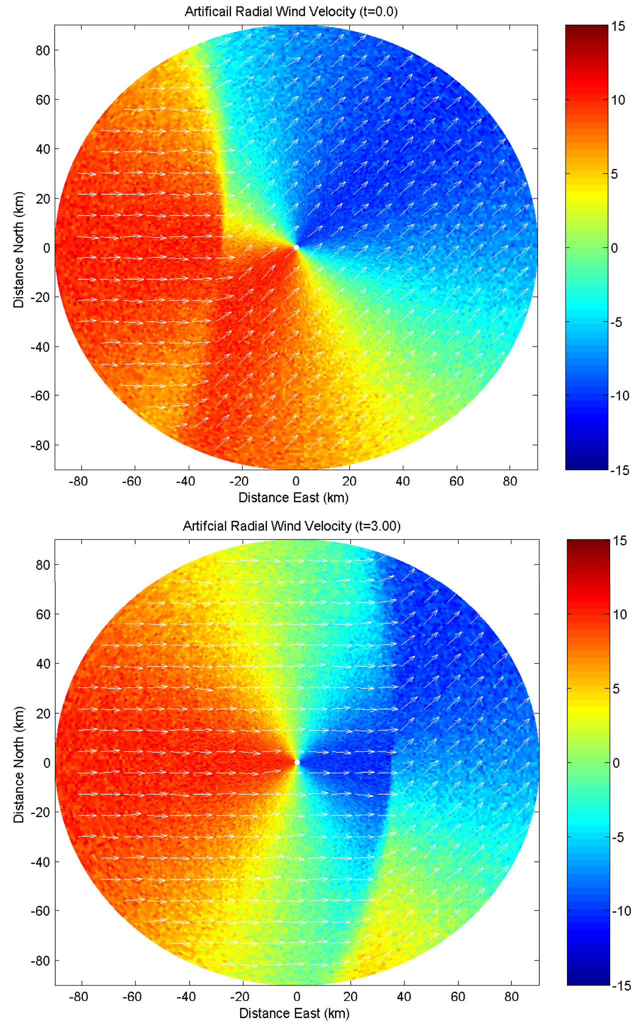

Fig. 3. A schematic representation of the definition of radial component of the wind field on the conical moving surface of a radar. The example shown is for a convergence line moving from the left to the right of the image. Gaussian noise has been inserted representing measurement error.

To avoid the computationally overwhelming problem of inverting the covariance matrix **B** in the minimization of the cost function (4) and to accelerate the convergence of the minimization algorithm, a pre-conditioning of the minimization problem is needed (see Lorenc, 1997). This can be achieved by defining a variable **U** to be applied to the assimilation increment $\delta\mathbf{x}$ ($\mathbf{U}\delta\mathbf{x} \equiv \mathcal{X}$) such that it transforms the forecast error ϵ in the model space into $\tilde{\epsilon}$, a variable of an identity covariance matrix (i.e. $<\tilde{\epsilon}, \tilde{\epsilon}^T> = \mathbf{I}$, where $< ., . >$ is an inner product). This change of variable can be written

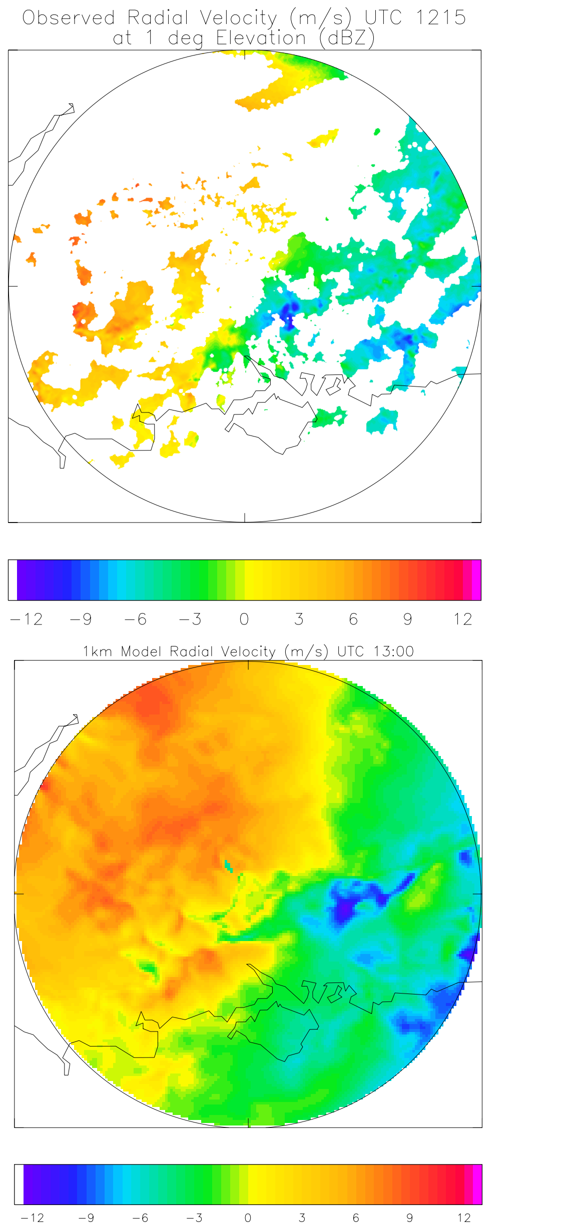

Fig. 4. Observed radial winds ($\mathrm{m\,s^{-1}}$) for 1st July 2003 compared with model-derived radial wind.

as $\epsilon = \mathbf{U}^{-1}\tilde{\epsilon}$. Thus

$$\mathbf{B} = <\epsilon, \epsilon> = \mathbf{U}^{-1} < \tilde{\epsilon}, \tilde{\epsilon}^{T} > \mathbf{U}^{-T}, \text{ or } \mathbf{B}^{-1} = \mathbf{U}^{T}\mathbf{U}. \quad (5)$$

This leads to a new representation of the incremental cost function of the form

$$\mathcal{J}[\mathcal{X}] = \frac{1}{2}\mathcal{X}^{T}\mathcal{X} + \frac{1}{2}[\mathbf{HU}^{-1}\mathcal{X} - \mathbf{y} + \mathcal{H}\mathbf{x}_b]^{T}\mathbf{E}^{-1}[\mathbf{HU}^{-1}\mathcal{X} - \mathbf{y} + \mathcal{H}\mathbf{x}_b] \quad (6)$$

With this cost function (6), no inversion of **B** is needed. The control variables $\mathcal{X}$ are velocity potential, streamfunction, unbalanced pressure and relative humidity.

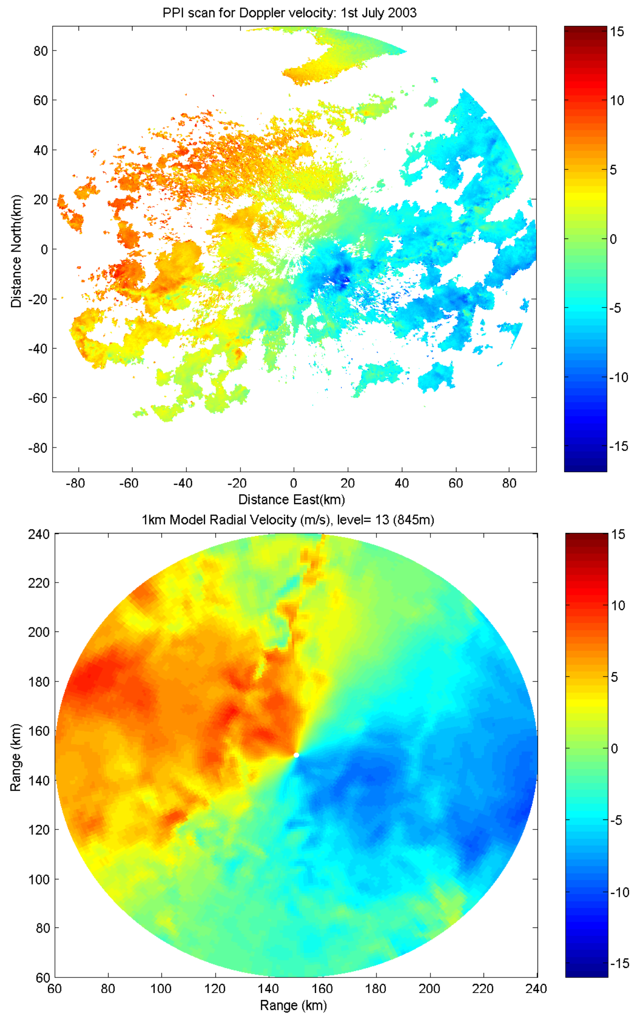

Fig. 5. Observed Doppler radial winds compared with the simulated model radial velocity, at 1 km resolution and model level 13, using Eq. (3).

4.1 Direct assimilation of PPI data (Sun and Crook, 2001)

Due to the poor vertical resolution of radar data, a vertical interpolation of radar data from constant elevation levels to model Cartesian levels can result in large errors. For this reason a direct assimilation of PPI data with no vertical interpolation was recommended in Sun and Crook (1998, 2001). However, radar data have better horizontal resolution than that of the model (the poorest polar radar data is approximately 0.5-km at the farthest range distance). An observation operator must be formulated to map the model variables from model grid such that the distance between the observations and model solution is estimated in the cost function.

Thus, we take advantage of the vertical resolution of the model being much better than those of radar data. The observation operation, $\mathcal{H}$, is formulated to map the data from the model vertical levels to the elevation angle levels via the formula

$$v_{r,e} = \mathcal{H}(v_r) = \frac{\sum G v_r \Delta z}{\sum G \Delta z}, \quad (7)$$

where $v_{r,e}$ is the radial velocity on an elevation angle level, v_r is the model radial velocity, and Δz is the model vertical grid spacing. The function $G = e^{-\alpha^2/2\beta^2}$ represents the power gain of the radar beam, β (in radiance) is the beam half-width and α is the distance from the center of radar beam (in radiance). The summation is over the model grid points that lie in a radar beam.

4.2 Doppler radial wind super-observations

Doppler radars produce radial wind raw data with high temporal and spatial density. The horizontal resolution of the data is around 300 m whereas the typical resolution of a mesoscale NWP model is of the order of 10 km. To reduce the representativeness error, and correspond the observations to the horizontal model resolution, one may use spatial averages of the raw data, that called super-observations. The desired resolution for the super-observations can be generated by defining parameters (which can be freely chosen) for the range spacing and the angle between the output azimuth gates (see Lindskog et al., 2002). Figure 4 shows radar radial winds from the Chilbolton radar on 1 July 2003, and the corresponding high resolution NWP radial winds calculated along the radar beam. Figure 5 displays the radar radial wind data and the corresponding NWP radial winds at a fixed model level. The impact of reducing the resolution is shown in Fig. 6.

4.2.1 Observation operator for radar radial winds

The radar wind observation operator **H** produces the model counterpart of the observed quantity that is presented to the variational assimilation. In the case of a horizontal wind observation from VAD profiles (see Fig. 2 the observation operator consists of a simple interpolation of the model wind field to the location of the observation. However, in the case of a direct assimilation of radar radial wind, which is not a model variable, the observation operator involves: (i) a bilinear interpolation of the NWP model horizontal wind components u and v to the observation location; (ii) a projection of the interpolated NWP model horizontal wind, at the point of measurement, towards the radar beam using the formula

$$v_h = u\cos\theta + v\sin\theta, \tag{8}$$

where θ is the azimuth angle of the radar beam; and (iii) the v_h is finally projected in the slanted direction of the radar beam as

$$v_r = v_h\cos(\phi+\alpha),\ \text{and}\ \phi = \arctan\left(\frac{r\cos\alpha}{r\sin\alpha + d + h}\right), \tag{9}$$

where α is the elevation angle of the radar beam (see Lindskog et al. (2002)). The formula for ϕ takes approximately into account the curvature of the Earth. In the term ϕ, r is the range, d is the radius of the Earth and h is the height of the radar above the sea level.

Some assumptions are, however, built into the standard formulation of the observation operator (9). First, the radar beam broadening is not taken into account. Second, the bending of the radar beam due to the hydrolapse in the boundary layer is not properly taken into account. Third, it is assumed that there is no mean velocity towards the radar due to the vertical motion of the precipitation, resulting in validity of measurements only for low elevation angles. This implicit assumption is embedded into (8) where only the NWP model horizontal wind is included.

One possible solution to relax the first assumption is to introduce a weighted average, using a Gaussian beam pattern, for the vertical interpolation of model horizontal wind components u, v of (8) to the observation location (see Salonen, 2002). Then to model the broadening of the radar beam in the observation operator, one can use the Gaussian weight function

$$w = \frac{1}{2\pi}\exp\left(-\frac{(z-z_0)^2}{k}\right) \tag{10}$$

in the vertical instead of linear interpolation when defining the model horizontal components u and v to the observation height. Where, in formula (10), z is the model level height

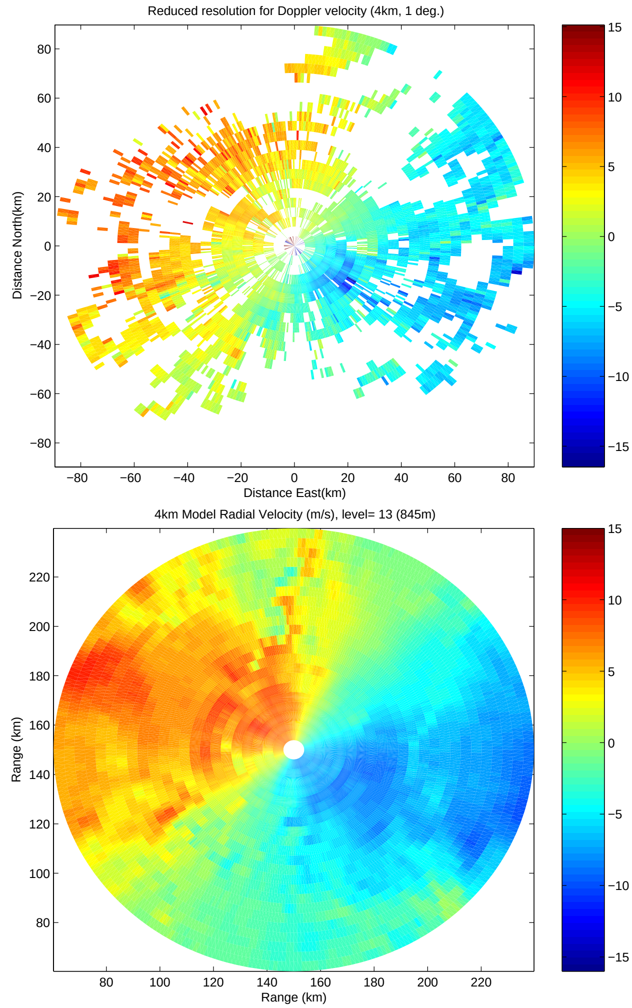

Fig. 6. The super-obbed Doppler radial winds, at 4 km resolution compared with the model radial velocity, at 4 km resolution.

and z_0 is the observation height. The k-term defines the width of the filter response function.

5 Observing System Simulation Experiments

Observing System Simulation Experiments (OSSEs) will be developed to test the ability of 3D-Var (and possibly 4D-Var) to assimilate radar radial Doppler winds to assess the impact of such observations on the analysis and forecasts and the sensitivity of the forecasts to assumed errors, super-obbing strategy, number of vertical PPIs, frequency of data etc. (Lord et al., 1997, 2002). The basic strategy is to have a control NWP forecast from the basic system (RUN1), and then use the NWP forecast fields from a different forecast run (RUN2) to simulate observations, referred to as pseudo-observations. The observations are then assimilated into an extended version of the basic system used for RUN1 to see how well the forecast from the assimilation, including the pseudo-observations, matches RUN2 both at analysis time and in the resulting forecast.

The pseudo-observations should be generated using an observation operator and a typical expected data coverage allowing scan patterns and normal data loss. The pseudo-observations should also have typical errors added. Ideally allowance should be made for model error in the system that produce RUN2 to try to make the pseudo-observations more like the real atmosphere, e.g. a boundary layer low cloud bias can be corrected in the analysis before using the forecast to simulate the data (see Lord et al., 1997). We intend to use two approaches: One is to use a pair of 12 km resolution operational mesoscale assimilation and forecast runs. The second approach is to use a higher resolution forecast to produce the pseudo-observations for assimilation into a 12 km resolution run.

6 Conclusions

The aim of this paper was to investigate issues relating to the assimilation of Doppler radar radial winds into the UK Met Office operational forecast model, using 3D-Var approach. Examples of radial winds derived from the Chilbolton radar and the Met Office operational model implemented at a 1 km resolution have been considered.

In this paper we discussed the use of simulated data from a specific day, 1 July 2003. It is intended to test the 3D-Var system using both simulated and real data. We discussed the types of errors which might occur in radar radial winds, and how these errors may be simulated. We will used the OSSE to investigate the impact of various assumptions made in the assimilation procedure.

We hope to extend the experiments 4D-Var system once a limited area version is available.

Acknowledgements. This project is sponsored by the National Environmental Research Council (NERC) through the Universities Weather Research Network (UWERN), a part of the NERC Centers for Atmospheric Science (NCAS).

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