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ZACHARY B. WIENHOFF

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DOPPLER RADAR ANALYSES OF TORNADIC SUPERCELLS
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SCHOOL OF METEOROLOGY

BY

Dr. Howard B. Bluestein, Chair

Dr. Alan Shapiro

Dr. Louis J. Wicker

Dr. Jeffrey C. Snyder

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As an incoming graduate student from the University of Illinois, attending the University of Oklahoma meant leaving the place where I grew up for the first time. Not only was the academic transition difficult, but the added stress of being away from my home and family meant additional stress. Without the numerous people that I have met in Norman, this journey would have been much more difficult.

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Abstract

On 19 May 2013, two supercells moved through central Oklahoma and produced seven tornadoes, including one of EF-3 and one of EF-4 intensity. High temporal and spatial resolution observations were collected in both strong tornadoes by a mobile, rapid-scan, X-band, polarimetric radar (RaXPol). In conjunction with the nearby WSR-88D Twin Lakes radar, dual-Doppler analyses were synthesized to produce the three-dimensional wind field and vertical vorticity throughout the tornadoes' lifecycles. In order to account for significant differences in the temporal resolutions of the two radars, a reflectivity-tracking scheme was employed to interpolate linearly (in a Lagrangian sense) between two radar volumes through a variational algorithm. Previous studies have performed dual-Doppler syntheses with an S-band radar and a non-rapid-scanning mobile radar (e.g., Markowski et al., 2012a; Tanamachi et al., 2012; Kurdzo et al., 2015); however, the analyses in this study are unique because they are one of the first, to the authors' knowledge, to make use of data from both an operational, S-band radar and a mobile, *rapid-scan*, X-band radar.

This work focuses on the methodology used to combine both RaXPol and KTLX datasets, including discussion regarding the strengths and weaknesses of this technique. Analyses are presented on the both the Edmond-Carney and Norman-Shawnee tornadoes, examining the characteristics of the wind field associated with the intensification, maturity, and demise of the tornado. The 3-D wind fields of the tornado and mesocyclone are examined to see how they evolved in time, and especially how vertical vorticity and vertical velocity changed as the tornado

intensified. Differential reflectivity (Z_{DR}) from the WSR-88D is compared to the vertical velocity field estimated from dual-Doppler analyses to examine how the characteristics of a Z_{DR} column varied with respect to changes in the updraft intensity. Results from both cases are compared, and some thoughts on future improvement to this technique will be presented.

Chapter 1

Introduction

Prior to advances in numerical modeling, dual-Doppler analyses of tornadic thunderstorms served as one of the earliest techniques used to further our understanding of these phenomena. Dual-Doppler analyses allowed us to take one-dimensional radial velocity data and, for the first time, resolve three-dimensional features in tornadic supercells. In the late 1960s, the National Severe Storms Laboratory (NSSL) began using multiple Doppler radars to resolve three-dimensional features of thunderstorms, first beginning with the use of 3 cm Doppler radars (Brown and Peace Jr, 1968). In 1975, NSSL published one of the first dual-Doppler studies of a tornadic supercell using two S-band radars (10 cm wavelength) in which the mesocyclonic circulation, as well as regions of strong updrafts and downdrafts, were observed (Doviak et al., 1976). Despite the ability to resolve storm-scale features in three dimensions now, the lack of fixed-site radar coverage at this time made this technique nearly impossible use, and furthermore, any tornado or supercell data collected within the dual-Doppler domain occurred simply as a matter of luck.

Over the past two decades, mobile-Doppler radars have become increasingly common for atmospheric research. The mobile capabilities of these radars made studying mesoscale phenomena, such as tornadoes and supercells, more feasible as scientists were able to plan and strategically place units in locations to maximize the potential for data collection in a relatively safe manner (Bluestein, 1999).

Radars transmitting at W-band (Bluestein and Pazmany, 2000), X-band (Alexander and Wurman, 2005; Wurman et al., 1997), and C-band (Biggerstaff et al., 2005) have all been used to document tornadoes and supercells at higher resolution than achieved at S-band. The idea of utilizing such mobile platforms to perform a dual-Doppler analysis was first proposed by Bluestein et al. (1995) using 3 mm wavelength radars.

As radar technology slowly advanced, dual-polarization radars became more common in radar studies as it offered a great deal of additional information to meteorologists. Information about hydrometeor sizes, shapes, and cloud composition allowed scientists to better understand the makeup of clouds, a topic on which little was known prior to this time. As the importance behind this technology was more widely understood, dual-polarization radar began to become more common on mobile platforms, and was added to the WSR-88D network in 2011. In conjunction with dual-polarization variables, such as differential reflectivity, and dual-Doppler observations, dynamic processes observed within thunderstorms could be compared to the microphysical aspects to better understand how each affects the other.

As the number of mobile radars multiplied and the technology became increasingly complex, an increase in the temporal resolution of these instruments began to be strongly desired. Although phased-array radars were in existence, a more cost-effective option with dual-polarization capabilities became increasingly appealing. In 2011, RaXPol, a mobile, rapid-scan, X-band, dual-polarization radar became operational (Pazmany et al., 2013). RaXPol allowed for a much higher temporal resolution data collection than traditional mechanical scanning radars through the use of a heavy-duty, high-speed pedestal that can withstand an antenna scanning rate of 180° s^{-1} . At this rate, RaXPol can collect 10° volumes in approximately

20 seconds, a rate unmatched by any other atmosphere-observing, mechanical-scanning radar. In order to collect a sufficient number of independent samples, RaXPol employs a frequency hopping technique which allows for relatively equal numbers of samples to be collected as a traditional radar while maintaining 180° s^{-1} . RaXPol will be discussed in greater detail in the instrumentation Chapter 2.

One of the greatest challenges behind dual-Doppler wind syntheses deals with the collection of the data itself. Dual-Doppler analyses are rarely available because doing so requires careful collaboration and synchronization of participating radars beforehand. Because the behavior of supercells and tornadoes are often times unpredictable, the task of setting up and collecting useful dual-Doppler data can be daunting. Furthermore, the increased advances in certain types of radar technology, and no change in other areas, provides additional challenges. For example, because RaXPol is one of the only radars of its kind (mobile, rapid-scan, polarimetric), synced data collection with another radar of similar temporal resolution is nearly impossible. With technology like RaXPol sweeping the Great Plains at a rate of 180° s^{-1} , the use of preexisting radars such as the WSR-88Ds in concert with mobile radars seems pointless given that the analyses could only be valid over a small period of time. However, provided that a method be developed to match the temporal resolutions of relatively high scanning rate antennas to those that scan much slower, these data sets could still complement one another. With several different sources of data viewing these storms from multiple angles, is it possible that the data be adapted so that fixed-site, WSR-88D data can be combined with rapid-scan, mobile radar data to create a dual-Doppler analysis despite the drastic differences in scanning rates? This work aims to tackle, and hopefully answer, this question.

On 19 May 2013, two supercells moved through central Oklahoma producing seven tornadoes, including one of EF-3 and one of EF-4 intensity. The purpose of

this thesis is to perform a dual-Doppler analysis on both tornadic supercells to examine the evolution of both storm with respect to each other and previous studies. Evolution of the near-storm inflow is examined with respect to the tornado, and more specifically the roles the tornado played on its own environment are hypothesized. Finally, analyses of the Z_{DR} column with respect to the updraft are analyzed as we aim to learn more about the relationship between the Z_{DR} column behavior and tornadogenesis. Section 1.1 will include a brief review of dual-Doppler analyses, including the techniques through which the three-dimensional wind field is obtained, potential sources of error in the analyses, and previous studies of tornadoes using this analysis method. Section 1.2 will also briefly review differential reflectivity and polarimetric observations of supercells. Chapter 2 will review the instrumentation used in this study, and chapter 3 will cover the methodology behind how the data were obtained, processed, and analyzed, including the use of a spatially variable advection correction technique (Shapiro et al., 2010). Chapter 4 will examine the results of dual-Doppler syntheses from both the Carney-Edmond and Norman-Shawnee supercells, and Chapter 5 will look at the evolution of observed ZDR columns with respect to dual-Doppler derived updraft speeds. Finally, Chapter 6 will provide thoughtful conclusions to this thesis, discussing specifically how future work can help verify results from this experiment and make this technique a useful tool for future studies. An appendix section will discuss the use of reflectivity-tracking techniques on high spatiotemporal resolution data in an attempt to retrieve the three-dimensional wind field from single-Doppler data.

1.1 An Overview of Dual-Doppler Analyses

1.1.1 An Iterative Technique for Dual-Doppler Synthesis

Dual-Doppler analyses have been frequently used to observe the three-dimensional wind field in a way not capable of being accomplished through single-Doppler observations alone. Over the past 40 years, this technique has been employed to observe tornadic supercells on numerous occasions (e.g. Ray et al., 1975; Brandes, 1977a; Dowell and Bluestein, 1997). In the past 20 years, the increase in the number of mobile, ground-based radars has increased not only our ability to obtain dual-Doppler data due to the ability to move and preposition for a storm, but also the ability to see more fine-scale features as resolution capabilities increased. After the construction of the Doppler-on-Wheels (DOW) 2, the original DOW (DOW1) was retired, paving the way for DOW3 to be born (Wurman, 2001). The conception of DOW3 led to the construction of dual-Doppler analyses through mobile radar operations alone. While the hardware to accomplish these tasks was quite young, the theory behind dual-Doppler had already been in the works since the late 1960s and continued to make advances as well.

Dual-Doppler analyses are possible when two (or more) radars have different viewing angles of the same point of interest. That is, if the cross-beam angle of the two radars is not parallel, it is possible to observe the three-dimensional wind field. Obtained from Dowell and Shapiro (2003), the equations employed in a standard dual-Doppler synthesis to obtain the three-dimensional wind field are:

$$\mathbf{r}_1 \cdot \mathbf{v} = \sin \alpha_1 \cos \phi_1 u + \cos \alpha_1 \cos \phi_1 v + \sin \phi_1 w = v_r^{(1)} - \sin \phi_1 w_t \equiv V_1 \quad (1.1)$$

$$\mathbf{r}_2 \cdot \mathbf{v} = \sin \alpha_2 \cos \phi_2 u + \cos \alpha_2 \cos \phi_2 v + \sin \phi_2 w = v_r^{(2)} - \sin \phi_2 w_t \equiv V_2 \quad (1.2)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} + \Gamma w = 0. \quad (1.3)$$

$$\Gamma(z) = \left(\frac{1}{\rho}\right) \left(\frac{d\rho}{dz}\right) \quad (1.4)$$

In (1.1) and (1.2), $\mathbf{r} = \sin \alpha \cos \phi \mathbf{i} + \cos \alpha \cos \phi \mathbf{j} + \sin \phi \mathbf{k}$ represents a unit vector which changes with the direction of the radar beam. The equation $\mathbf{v}(x, y, z) = u\mathbf{i} + v\mathbf{j} + w\mathbf{k}$ represents the three-dimensional wind component at a particular Cartesian grid point. The angle ϕ represents the elevation angle of the radar beam with respect to the horizontal surface, and α represents the azimuth angle of the radar beam with respect to cardinal north. v_r is the radial velocity obtained from the Doppler radar, and w_t represents the terminal fall velocity of particles. For the sake of this paper, the particle fall speeds are computed empirically as a function of both density (ρ) and reflectivity according to Atlas et al. (1973). In equations (1.1) and (1.2), subscripts and superscripts (1) and (2) represent the two different radars. In these equations, V_1 and V_2 represent the radial velocity components as seen by each radar with the particle fall speeds removed from each. Finally, in equations (1.3) and (1.4), we consider the anelastic form of the mass continuity equation (Dowell and Shapiro, 2003).

Under the above considerations, we have three equations and three unknowns, and we can solve these equations. In 1969, Armijo theorized a method of directly synthesizing a three-dimensional wind field in which a cylindrical coordinate system is used to solve for the wind field. This method chooses a coordinate system such that the x-axis lies on the radar baseline, and the y-axis such that it is both orthogonal and satisfying a right-hand coordinate system. With two radars in this coordinate system, the radial velocity components are as follows:

$$V_1(x, y, z) = \frac{1}{R_1}[xu + yv + z(w + V_t)] \quad (1.5)$$

$$V_2(x, y, z) = \frac{1}{R_2}[(x - x_2)u + yv + z(w + V_t)] \quad (1.6)$$

where

$$\left. \begin{aligned} R_1 &= \sqrt{x^2 + y^2 + z^2} \\ R_2 &= \sqrt{(x - x_2)^2 + y^2 + z^2} \end{aligned} \right\} \quad (1.7)$$

and a supplementary continuity equation

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = kw, \quad (1.8)$$

where k represents a small positive constant used to approximate the negative logarithmic gradient of air density. In the interest of brevity, the entire derivation is not included in the paper. Armijo defined the quantities

$$\left. \begin{aligned} \Gamma(x, y, z) &= \frac{1}{x_2}(R_1 V_1 - R_2 V_2) \\ \Psi(x, y, z) &= \frac{1}{yx_2}[(x_2 - x)R_1 V_1 + xR_2 V_2] \end{aligned} \right\}, \quad (1.9)$$

which can be subsequently related to u and v by the following equations:

$$\left. \begin{aligned} u &= \Gamma \\ v &= \Psi - \frac{z}{y}w \end{aligned} \right\}. \quad (1.10)$$

In order to satisfy the partial differential equation which arises from combining equations (1.8) and (1.10), we can convert to a cylindrical coordinate system $(s, r, \theta;$ Fig. 1.1) where the central axis of the cylinder is oriented horizontally along the radar baseline. In this cylindrical coordinate system, where the cylinder is oriented with the central axis along the x-axis, s represents the x distance (or projection on the x-axis) of a scatterer viewed by the radars (Fig. 1.1a), r represents the radial distance from the baseline to the scatterer, and θ represents the angle between a plane with the radar beams and $z = 0$ (Fig. 1.1b). Under this coordinate transformation, the solution to the partial differential equation satisfied by the lower boundary condition $w = 0$ results in

$$w(s, r, \theta) = \int_0^\theta \beta(s, r, \phi) r \cos \theta e^{kr(\sin \theta - \sin \phi)} d\phi, \quad (1.11)$$

where

$$\beta(s, r, \theta) = -\left(\frac{\partial \Gamma}{\partial s} + \cos \theta \frac{\partial \Psi}{\partial r} - \frac{\sin \theta}{r} \frac{\partial \Psi}{\partial \theta}\right). \quad (1.12)$$

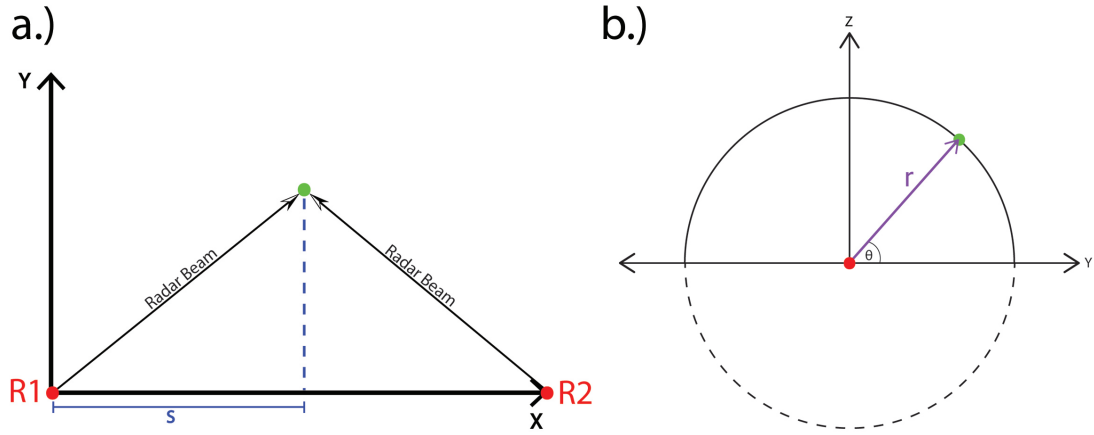


Figure 1.1: Schematic illustration of the cylindrical (s, r, θ) coordinate system proposed by Armijo (1969). In a 2D x/y view (a.), the x-axis is oriented along the baseline marked by the locations of the 2 radars (R1 and R2), with arrows representing their viewing angles. The projection of the scatterer (green dot) on the x-axis, s , is in blue. In a y/z view (b.), the orientation of the cylinder (which is oriented with the central axis along the X-axis) results in coplanar "arcs" traced out by the half of the cylinder occupying the space where $z > 0$. The baseline (x-axis) is denoted by the red dot and is into the page. The radius from the baseline to the edge of cylinder (the location of the scatterer), r , is in purple, and the angle between the radar beams and $z = 0$ is represented by θ .

Finally, substituting w into (1.10), we obtain values for u and v as well. However, one assumption that we make in this particular case is that all of the data are assumed to fall in an imaginary, three-dimensional plane S where x, y , and z are all greater than 0. If this isn't the case, which often happens in real data, the integration along the coplanar arc will fall outside of the data domain and the boundary conditions are not easily satisfied. Additionally, this method which integrates the continuity equation along a specific coplanar arc can quickly accumulate errors resulting in data that could be unreliable (Dowell and Shapiro, 2003).

The methods of direct synthesis provided by Armijo provided a way for a three-dimensional wind field to be synthesized from multiple radars; however, resulting errors in the analyses made finding more accurate methods of synthesizing these wind fields more appealing. In the late 1970s, many scientists began using a technique in which equations satisfying the three-dimensional wind field are solved iteratively with the hopes of reducing errors. Ray et al. (1980) discusses a number of techniques used to resolve the 3D winds and reduce errors in the data as much as possible.

The use of an the iterative technique first requires that data be mapped to a Cartesian coordinate system. Ray et al. (1980), for example, used a Cressman analysis technique to interpolate data to a cartesian grid using the following expression:

$$W(D_j) = \begin{cases} \frac{R^2 - D_j^2}{R^2 + D_j^2}, & D_j \leq R. \\ 0, & D_j > R,. \end{cases} \quad (1.13)$$

where $W(D_j)$ represents the Cressman weighting function, R represents the radius of influence at a particular data point, and D_j is the distance from a grid point j to the data point. In this study, data were mapped to a Cartesian coordinate system using a two-pass Barnes analysis (Barnes, 1964), which will be discussed further in section 3.2.

In Cartesian coordinates, it can be shown that the following equations can be used to satisfy the 3D wind field:

$$u = \frac{R_1 V_1 (y - y_2) - R_2 V_2 (y - y_1) - V_t [(z - z_1)(y - y_2) - (z - z_2)(y - y_1)]}{(x - x_1)(y - y_2) - (x - x_2)(y - y_1)} - w \frac{[(z - z_1)(y - y_2) - (z - z_2)(y - y_1)]}{(x - x_1)(y - y_2) - (x - x_2)(y - y_1)} = C_1 - w C_2, \quad (1.14)$$

$$v = \frac{R_2 V_2 (x - x_1) - R_1 V_1 (x - x_2) - V_t [(z - z_2)(x - x_1) - (z - z_1)(x - x_2)]}{(x - x_2)(y - y_1) - (x - x_1)(y - y_2)} - w \frac{[(z - z_2)(x - x_1) - (z - z_1)(x - x_2)]}{(x - x_2)(y - y_1) - (x - x_1)(y - y_2)} = C_3 - w C_4. \quad (1.15)$$

To solve the above equations iteratively, Ray et al. (1980) outlines a procedure in which w is estimated initially and equations (1.14) and (1.15) are solved for. After each subsequent computation, w is recomputed through eq. (1.8), and u and v are continually refined until the solutions converge. In this particular case, w is obtained through an upward integration of the anelastic mass continuity equation.

Perhaps more elegantly so, Dowell and Shapiro (2003) also described equations by which solving iteratively will yield solutions for u , v , and w . By first solving for the horizontal wind components in Eqns. (1.1) and (1.2), the equations are simplified into:

$$u = A + Bw, \quad (1.16)$$

and

$$v = C + Dw, \quad (1.17)$$

where

$$A = \frac{\cos \alpha_2 \cos \phi_2 V_1 - \cos \alpha_1 \cos \phi_1 V_2}{\cos \phi_1 \cos \phi_2 \sin(\alpha_1 - \alpha_2)}, \quad (1.18)$$

$$B = \frac{\cos \alpha_1 \cos \phi_1 \sin \phi_2 - \cos \alpha_2 \cos \phi_2 \sin \phi_1}{\cos \phi_1 \cos \phi_2 \sin(\alpha_1 - \alpha_2)}, \quad (1.19)$$

$$C = \frac{\sin \alpha_1 \cos \phi_1 V_2 - \sin \alpha_2 \cos \phi_2 V_1}{\cos \phi_1 \cos \phi_2 \sin(\alpha_1 - \alpha_2)}, \quad (1.20)$$

$$D = \frac{\sin \alpha_2 \cos \phi_2 \sin \phi_1 - \sin \alpha_1 \cos \phi_1 \sin \phi_2}{\cos \phi_1 \cos \phi_2 \sin(\alpha_1 - \alpha_2)}. \quad (1.21)$$

The above equations provide us with the necessary information for solving the dual-Doppler equations iteratively except for w . To solve for w in this case, the following finite-differencing scheme on the anelastic mass continuity equation is used:

$$\left(1 + \frac{\Gamma \Delta z}{2}\right) w_K = \left(1 - \frac{\Gamma \Delta z}{2}\right) w_{K-1} - \frac{\Delta z}{2} \left(\frac{\partial u}{\partial x} \Big|_K + \frac{\partial u}{\partial x} \Big|_{K-1} + \frac{\partial v}{\partial y} \Big|_K + \frac{\partial v}{\partial y} \Big|_{K-1} \right). \quad (1.22)$$

The iterative procedure, outlined in Dowell and Shapiro (2003), then begins by giving a first guess for w at our lowest grid point. In this case, the lowest grid point occurs at the surface, thus $w = 0$ at the ground due to impermeability. The wind components u and v are computed from Eqns. (1.16) and (1.17), and the procedure moves to the next vertical level where another first guess of w is made, u and v computed, and the estimate of w adjusted based on Eqn. (1.22) until the procedure converges. In this case, upward integration of the associated continuity equation is used to solve for w , whereas some cases utilize a downward integration of the same equation (Ray et al., 1980).

1.1.2 Sources of Dual-Doppler Errors

One of the major underlying issues of dual-Doppler analyses is the numerous sources of error that can result. As discussed previously, one of the inherent

problems with both direct wind synthesis (Armijo) and the iterative technique is the issue of resolving the wind field when the lower boundary condition falls outside of the data domain. The coplanar arc on which the dual-Doppler equations are solved will often intersect the lateral sides of the data coverage, but do not intersect the ground. One way to remedy this issue, as discussed in Dowell and Shapiro (2003), is to provide lateral boundary conditions (e.g., $w = 0$), resulting in unique solutions within the data domain.

In addition, the iterative technique provides useful results when the solutions converge, but there are a number of situations where the resulting solutions become unstable, thus diverging. Dowell and Shapiro (2003) provide a derivation for a stability condition in which the solution would diverge. The solution of the wind synthesis at each grid point is said to be stable when the following equation is satisfied:

$$|\lambda|_{max} = \frac{\Delta z}{2\Delta x \left(1 + \frac{\Gamma \Delta z}{2}\right)} \frac{|\cos \alpha_1 \tan \phi_2 - \cos \alpha_2 \tan \phi_1| + |\sin \alpha_2 \tan \phi_1 - \sin \alpha_1 \tan \phi_2|}{|\sin(\alpha_1 - \alpha_2)|} \leq 1, \quad (1.23)$$

where $|\lambda| \leq 1$ is the associated condition for stability. More specifically, λ is represented by the following:

$$\lambda \equiv \frac{F^{(n+1)}}{F^{(n)}}, \quad (1.24)$$

where F is defined as the error amplitude obtained through the iterative dual-Doppler technique. In order for the stability condition to be satisfied, $F^{(n+1)}$ must be less than or equal to $F^{(n)}$ to satisfy the condition $|\lambda| \leq 1$. That is, the error obtained by iteration $n + 1$ must be less than or equal to the error obtained by iteration n , or our errors must get smaller as we iterate for the condition to be

satisfied. In practical application, equation 1.23 suggests that we are especially susceptible to errors in our analyses (and this divergence of our iterative solutions) in areas where $|\sin(\alpha_1 - \alpha_2)|$ approaches 0 (where $|\alpha_1 - \alpha_2|$ approach 0° or 180°) or where $|\phi_{1,2}|$ approaches 90° . Additionally, we could also increase the stability of our analyses by simply decreasing our vertical grid spacing (increasing our vertical resolution) or by increasing our horizontal grid spacing (it is assumed here that $\Delta x = \Delta y$).

Doviak et al. (1976) outlines several sources of error, including the variance in analyses as a result of interpolation to a Cartesian coordinate system. Many of the issues discussed by Doviak are problems inherent to radar data collection in a single-Doppler sense as well and not due to the combination of multiple radars. Doppler radars typically make estimates of power, Doppler velocity, etc., by taking more than one sample per radar volume. An increase in the number of samples the radar takes results in more accurate measurements over a volume (i.e., the variance of the estimate is reduced). However, a number of factors still play a role in radar-data error, which can lead to spurious results. Doviak et al. (1976) suggests that the mean Doppler estimated variance is represented by

$$\sigma_1^2 = (1 - R^2)\sigma^2(\hat{V}), \quad (1.25)$$

where R^2 is a variance reduction factor determined by the interpolation volume, and $\sigma^2(\hat{V})$ is related to a number of different factors such as turbulence and shear by the equation:

$$\sigma^2(\hat{V}) = \sigma_{shear}^2 + \sigma_{turb}^2 + \sigma_{beam}^2 + \sigma_{fall}^2 + \sigma_{smear}^2, \quad (1.26)$$

where the terms represent the variances due to shear, turbulence, variations in fall velocity and a finite beam width in which variations in particle size and speed occur across the radar volume (Nathanson, 1969). In this case, a fifth term

– beam smearing – is also considered in (1.26) to account for the rotation of the radar antenna.

Variance in the Doppler velocity due to variations in fall speeds are accounted for in (1.26). In order to satisfy the dual-Doppler equations, fall speed must be known (otherwise a fourth unknown variable exists in our original set of equations). To solve this, empirical relationships are used to determine the approximate fall speeds of hydrometeors based on a relationship with radar reflectivity. Because this is purely an estimate used to satisfy our equations, Doviak et al. (1976) explains that variability in the true drop-size distribution within the storm produce standard deviations in fall speeds near 1 m s^{-1} , and even greater errors (up to several m s^{-1}) are possible where fall speeds appropriate for liquid water are used for regions in which hail exists. Although this method of using empirically derived equations to assign fall speeds to hydrometeors introduces error into our analysis, it is a necessary factor in satisfying the dual-Doppler equations. Even so, Potvin and Wicker (2012) showed that when scanning at lower elevations such as those examined here, the dual-Doppler results were not severely affected by errors in the empirical fall speed relationship.

Finally, errors in dual-Doppler synthesis can occur due to temporal variation in the radar volumes. Chong et al. (1983) describes two effects of temporal variation on dual-Doppler analyses: the first is accounting for advection of the storm over the time period of one volume scan, and the second is the “intrinsic” temporal variation that occurs, or more simply, accounting for storm evolution over a volume. While storm evolution is difficult to account for and storms are often assumed to be in steady state over the update time interval, corrections for translation are possible.

Translational motion must be accounted for because the storm continues to move while a volume scan is being completed. If translation were not accounted for, the storm would appear to “tilt” with height in the direction of the mean storm

motion. If this is not accounted for, substantial errors could occur, particularly in our estimates of vertical velocity. In this case, accounting for translation is especially important given the large differences in sampling interval between the KTLX WSR-88D and RaXPol radars. In this study, the analysis accounts for advection by using a time-to-space conversion given the approximate storm motion, and advecting the upper levels backwards in time to account for storm motion. Chong et al. (1983) comments that the errors due to advection alone can be minimized as long as the advection velocity can be calculated to an approximate accuracy of $\pm 2 \text{ m s}^{-1}$.

The above examples describe only some of the errors characteristic of dual-Doppler analyses. While many are inherent in this type of analysis, care is taken to ensure the best possible results with the available data. Specifics regarding the methodology provide some more information on errors associated with this study and will be the focus of Chapter 3.

1.1.3 Dual-Doppler Observations of Tornadic Supercells

For nearly 50 years, observations of convective storms have been made utilizing more than one Doppler radar. The first dual-Doppler study (or “two-Doppler measurement,” as it was first referred to) of a convective storm took place in central Oklahoma (Brown and Peace Jr, 1968). In this study, the Cornell Aeronautical Laboratory (CAL) Doppler radar located at Mustang Airport in El Reno, OK, and the NSSL Doppler radar located near Chickasha, OK, were used to observe a convective storm. Data were compared to the NSSL upper-air network (observing sites approximately every 50 km) and to the NSSL surface network and revealed good agreement with observations. Initial limitations of this technique were well documented, including the trade-off between data coverage area and data resolution, as well as the lack of near-surface data as the radars become more widely

spaced. Nevertheless, this technique had proven the ability to resolve the wind field with much higher resolution (i.e., meso- γ scale) than ever before.

In 1975, research using the dual-Doppler technique continued with the construction of two S-band Doppler radars placed 41 km apart: one in Norman, OK, and the other near Cimmaron Field, OK. After years of studies using X-band radars, S-band radars were used because they suffer from less attenuation in heavy rain and have the ability to resolve unambiguously higher wind speeds associated with convective storms (Brown et al., 1975).

At approximately 1530 UTC on 20 April 1974, a tornadic storm with a hook echo passed through the dual-Doppler lobes of the Norman and Cimmaron radars, which were synchronously collecting data. The dual-Doppler analyses revealed two dominant, cyclonic circulations associated with the storm, both coincident with regions of strong updraft (Ray et al., 1975). Additionally, one of these updrafts was located in the weak echo region and was centered on the tornado path. A downdraft was also noted to the southwest of the main circulation. These features, discovered during the infancy of severe storms research, characterize tornadic supercells as we know them today. Ray (1976) examined vorticity and divergence fields from 8 June 1974 where he found similar features to the 20 April 1974 case, such as a downdraft southwest of the main updraft, albeit weaker than observed in the earlier case. Ray concluded that changes in vertical vorticity during the analysis were most likely caused by vorticity advection as the supercell approached a steadier state, but he hypothesized that tilting and stretching of vorticity were likely important sources of rotation for the storm.

Brandes (1977a) performed a similar analysis using the same radar network. On 6 June 1974, a cluster of severe storms moved through the dual-Doppler domain. Several of the storms in this cluster were characterized by movement to the right of the mean storm motion, something that had yet to be seen using a

dual-Doppler synthesis. Six cyclonic circulations were observed on this day by the horizontal wind synthesis, and Brandes (1977a) postulated that the cyclonic circulations were well correlated with wind-damage reports, perhaps indicating the two were interconnected. Brandes also discussed the possibility that hook echoes, of which there were several observed this day, were due to the interaction between the inflow and outflow of the storm. With observations of severe convection becoming more frequent, dual-Doppler observations allowed researchers to observe many new features of these storms for the first time. In a separate paper, Brandes (1977b) examined the Oklahoma City and Harrah storms from 8 June 1974 and observed tornadogenesis in a region along the gust front where strong convergence coincided within a region of strong shear in both tornado cases.

In a more in-depth analysis of these two tornadoes, Brandes (1978) noted that the low-level wind field showed an influx of moisture into the storm, suggesting it was feeding on the warm, moist air. A prominent hook echo was again observed as the low-level wind field continued to intensify, and the cyclonic vortex maximum approached $5 \times 10^{-2} \text{ s}^{-1}$ (Brandes, 1978). He also noted a northward tilt of the vortex. Comparing the dual-Doppler data from the two storms on this day, Brandes concluded that cyclogenesis had the tendency to take place along shear and convergence zones. He also made mention that as tornadogenesis was about to occur, the tangential component of the wind field and the average inflow angle at low levels was increased. Brandes concluded that tornadogenesis is associated with a “breakdown” of the parent mesocyclone into more of a two-cell axial vortex structure as opposed to the single-cell structure which initially is observed.

Eagleman and Lin (1977) even further investigated the 8 June 1974 Oklahoma City storm and in particular discussed the observation of counter-rotating cyclonic and anticyclonic vortex couplets aloft in the supercell. The study suggested that the mesocyclone would act as a solid body, and as environmental air flowed around

the cyclonic vortex an anticyclonic vortex couplet was created. Additionally, Eagleman and Lin suggested that the tornado was produced as the dynamic updraft acted to intensify associated cyclonic rotation in the parent mesocyclone. Heymsfield (1978) further examined the Harrah storm, and more specifically looked at the effects on vertical vorticity throughout the lifetime of the supercell. He concluded that the cyclonic/anticyclonic couplets observed at mid-levels in the storm could be initially attributed to tilting of vorticity by the updraft. This also explained why the couplets initially straddled the updraft. A strengthening downdraft and its effects on the absolute vorticity resulted in enhanced anticyclonic flow and a shift in the vortex “doublet” northward. The result was a cyclonic couplet collocated with the updraft. The collection of data on 8 June 1974 was undoubtedly one of the most important cases of tornado observations in history. Not only did it begin to pave the way for scientists to better understand tornadic supercells, but it provided a better understanding of radar and its potential uses for dual- and multiple-Doppler analyses.

On 20-21 May 1977, an outbreak of 16 tornadoes occurred near the Norman-Cimmaron Doppler radar network. Ray et al. (1981) examined data from storms near Del City and Fort Cobb, specifically examining characteristic similarities in the structure and morphology of the storms. In both cases, a horseshoe-shaped updraft was observed, and a shallow region of weak subsidence existed at low levels just prior to the formation of a tornado. More importantly, this study was the first to discuss that, despite the usefulness of the dual-Doppler technique, the temporal resolution of the analysis was too coarse to provide any reliable information on the dynamics of the storm. In any case, enhanced temporal resolution could provide a better understanding of storm dynamics, especially near tornadogenesis (Ray et al., 1981). Johnson et al. (1987) examined the same cases, identifying that the origins of midlevel rotation were likely different from those responsible for tornado

production. Although low-level data were not available to examine tornadogenesis more closely, it was concluded that descending vorticity from midlevels was unlikely to be the cause of tornadogenesis in the case.

A study from Brandes (1981) examined specifically the Del City, OK storm again in “unprecedented detail.” Brandes hypothesized from the analyses that vorticity was transported downward from aloft in the storm and then tilted into the tornado, but the primary source of intensification was due to stretching in the updraft. He concluded that updrafts, not the “rear downdrafts” as suggested by Lemon and Doswell III (1979), was the primary driver of the tornado. Several years later, Brandes (1984b) again examined vertical vorticity in both the Del City and Harrah storms, as discussed previously. Brandes (1984b) concluded that, in this case, the downdraft/updraft interaction was important as it played a direct role in the intensification of low-level vorticity, and ultimately tornadogenesis. Brandes (1984a) and Hane and Ray (1985) both used radar data for thermodynamic retrieval to attempt to understand the effects of buoyancy and perturbation pressure on the tornado and mesocyclone. Brandes (1984a) hypothesized that rear downdrafts in tornadic supercells were due to intensification of low-level vorticity and an associated reversal of the vertically directed perturbation pressure gradient force. Eventually, air parcels from aloft fill the mesocyclone leading to its demise Brandes (1984a).

After nearly 20 years of ground-breaking data collection, the same network of NSSL S-band radars continued collect new data which was the subject of research into the 1990s. Bluestein and Woodall (1990) analyzed a splitting supercell, and the first ever dual-Doppler data of a low-precipitation (LP) supercell. In this study, a cyclonic/anticyclonic couplet was again observed at midlevels, and observations linking vertical vorticity to enhancements in vertical motion observationally confirmed findings of early numerical studies (e.g. Rotunno and Klemp, 1985). Prior

to storm splitting, the existence of multiple updrafts was observed at low levels as the storm first split aloft. A short-lived tornado also accompanied the supercell, and it was noted that a weakening in the updraft occurred near the time of tornadogenesis, possibly associated with a downdraft and increased precipitation (Bluestein and Woodall, 1990). Several years later, Dowell and Bluestein (1997) examined the Arcadia, OK, storm observing a number of similar features described in Brandes' research; however, further understanding of supercell characteristics helped them to form separate conclusions. Dowell and Bluestein noticed increased convergence below the low-level mesocyclone, which acted to rapidly increase vorticity in the storm. Like those hypotheses presented by Lemon and Doswell III (1979), the rear-flank downdraft aided in the near-surface convergence and a simultaneous doubling of low-level vertical vorticity which could be a major key to the formation of a tornado (Dowell and Bluestein, 1997).

In 1994-1995, the first Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX) took place across the Great Plains (Rasmussen et al., 1994). This collaborative effort attempted to bring together much of the work that had been done over the past 25 years in tornado research. Years of work studying tornadoes brought forth many new ideas and hypotheses about tornadoes and how they form, but one major link remained missing. Although mid- and upper-level rotation had been observed on numerous occasions by NSSL (Cimmaron) and WSR-88D radars, no worthy explanation existed for how a tornado is then formed. Through use of ground-based radars, and the additions of the NOAA P-3 aircraft equipped with a Doppler radar, the NCAR ELDORA Doppler radar, a mobile, W-band radar, and a portable-Doppler radar (Bluestein et al., 1993), the radar network for VORTEX was especially advanced. The use of aircraft and portable-Doppler radars made the radar network more versatile, allowing for movement of the instruments as needed. An additional benefit of airborne radar was that

dual-Doppler analyses could be performed using a technique known as the fore-aft scanning technique (FAST; Jorgensen et al., 1995) which only required one radar. As the aircraft flew near the storm at high speeds, scanning in both the fore and aft directions could then be combined to create a dual-Doppler wind synthesis. Dowell et al. (1997) analyzed data collected using this method when the P-3 flew near two tornadic supercells on 29 May 1991 during the Cooperative Oklahoma Profiler Studies (COPS-91; Hane et al., 1993). An LP storm near Ulysses, KS, showed a similar evolution pattern to the LP storm in Oklahoma discussed by Bluestein and Woodall (1990), where the storm was observed to transition from an LP supercell to classic supercell just prior to tornadogenesis. Both supercells died shortly after the short-lived tornadoes dissipated. The second supercell, near Woodward, OK, contained far more precipitation than the Ulysses storm, and cyclic mesocyclogenesis was observed during this case as new mesocyclones formed to the right of the previous (Dowell et al., 1997). Additional cases of dual-Doppler analyses using the FAST technique from the VORTEX project occurred on 29 May 1994 near Newcastle, TX. In a study by Wakimoto and Atkins (1996), a tornado is observed to develop when a strong low-level shear zone was vigorously stretched by an updraft. Interestingly, the F3 tornado was *not* collocated with the mid-level mesocyclone, atypical of a strong tornado. It was concluded that strong tornadoes could actually originate from the boundary layer instead of descending from aloft. Subsequent examinations of the same case revealed that while the tornado and mid-level circulation may not have been collocated during tornadogenesis, the low-level and mid-level circulations were likely undergoing a merger at this time (Ziegler et al., 2001).

Data from the second aircraft, NCAR’s ELDORA, were collected on 16 May and 8 June 1995 near Garden City, KS, and McLean, TX, respectively. Wakimoto et al. (1998) identified baroclinically generated horizontal vorticity, which had been

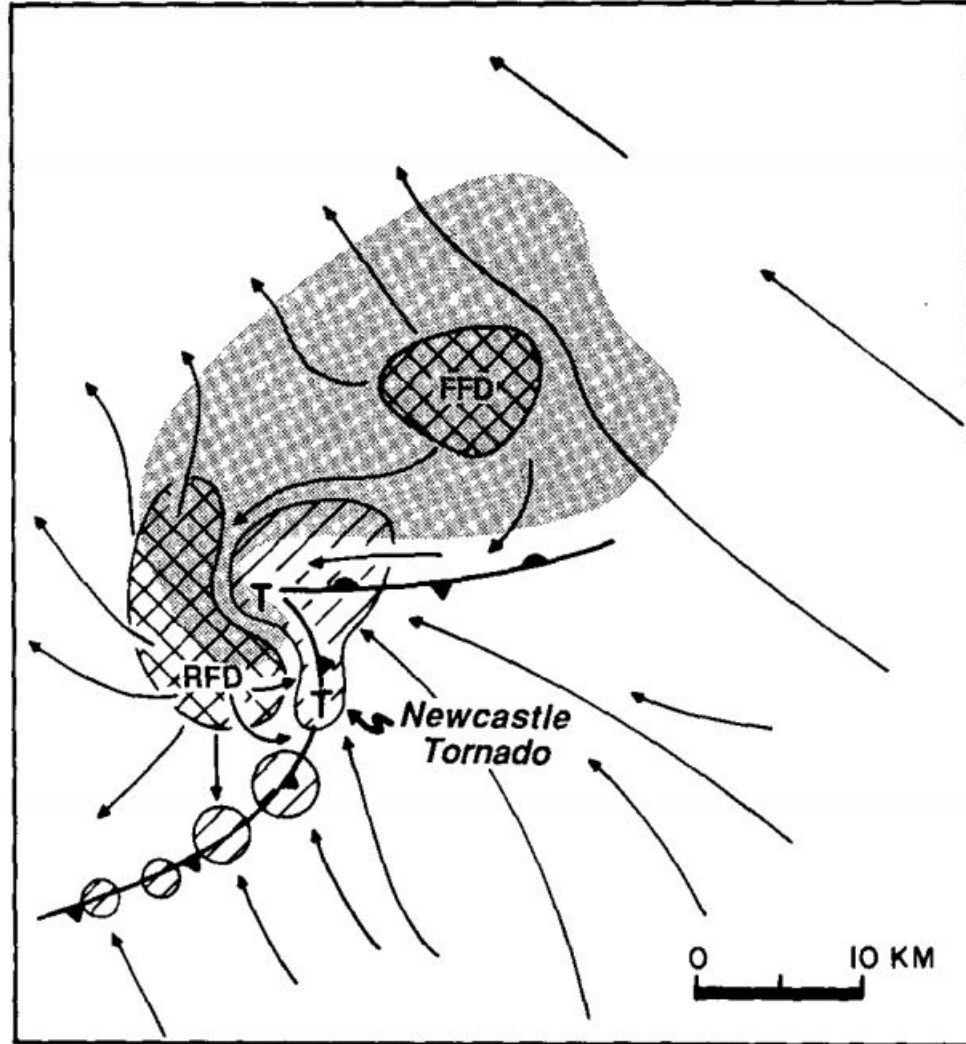


Figure 1.2: Schematic plan view of a supercell thunderstorm at the surface adapted from Lemon and Doswell III (1979) and Davies-Jones (1986). The gray shade encompasses the radar echo. The gust front structure is depicted using a solid line and frontal symbols. Surface positions of the updraft are hatched. Forward-flank downdraft (FFD) and the rear-flank downdraft (RFD) are crosshatched. Streamlines relative to the ground are also indicated. Likely tornado locations are shown by encircled T's. The location of the Newcastle tornado is indicated. Figure from Wakimoto and Atkins (1996).

tilted and stretched, to be an important source of vorticity for the mesocyclone. In a second paper, Wakimoto and Liu (1998) observed an occlusion downdraft just before “mesocyclonic vortex breakdown.” As they described, vortex breakdown in the mesocyclone could lead to tornadogenesis. It was suggested that the occlusion downdraft could play a major role in tornadogenesis, and identification of a strong downdraft in Doppler radar analyses could suggest that a tornado is imminent. Dowell and Bluestein (2002a) examined one of the first cases of cyclic tornadogenesis using dual-Doppler analysis.¹ Although the resolution of the data was still too coarse to resolve the tornado vortices, this data set had scans with high enough temporal resolution to provide better understanding of the cyclic nature of this supercell (Dowell and Bluestein, 2002a,b). In each case of cyclic mesocyclogenesis, tilting along the forward flank of the updraft was observed to create new maxima in vertical vorticity. As these new maxima were tilted and stretched, they intensified to tornado strength. Each consecutive vorticity maximum (and associated tornado, if one existed) would work its way towards the rear of the updraft (in a storm-relative sense), ultimately dissipating as it moved farther from the updraft. Dowell and Bluestein hypothesized that cyclic tornado formation may be a result of the tornado translating more slowly than the parent mesocyclone, ultimately splitting from the source of vortex tilting and stretching. In some instances, new vortices were spawned and were observed to be absorbed into larger vortices that may have persisted longer than average for this particular storm. This result was significant because it provided an explanation of a phenomenon that had previously been observed, but had not been observed at a high enough spatial and temporal resolution to gain a true understanding. These studies, and many others, utilizing airborne Doppler radar began to enlighten researchers on just how tough cracking

¹Although dual-Doppler analyses of cyclic tornadogenesis had already technically been performed, this was the first study performed where data quality was such that useful analyses could be conducted.

the mystery of tornadogenesis could be. Technological advances made observing tornadic supercells much easier, and new findings related the importance of downdrafts, baroclinic vorticity generation, etc., to tornadogenesis. Nevertheless, the mystery remained.

Although clear improvements were made in dual-Doppler radar analyses between fixed-site, ground-based data collection methods and airborne data collection methods, mobile, ground-based radars became increasingly popular in the mid 1990s. Although the first DOW was operational during VORTEX (Wurman et al., 1997), it was not until the second and third iterations of the DOWs were operational that dual-Doppler observations were made possible using truck-mounted, mobile-Doppler radars. The first truly ground-based, mobile radar was the 3-cm frequency-modulated and continuous-wave (FM-CW) radar (Bluestein et al., 1993). Although this radar had a nearly a 5° beamwidth, it was a major innovation in the world of tornado research. With this portable radar, scientists could now preposition the instruments based on the synoptic environment. The first usable, high-quality data collected from the FM-CW radar occurred near Northfield, TX, on 25 May 1994 (Bluestein et al., 1997). Data were examined in tandem with airborne radar data, photographs, and video to produce one of the most well-documented tornado cases to date. FM-CW data revealed a tornado damage path that was actually wider than the condensation funnel, with maximum wind speeds of nearly 75 m s^{-1} . Despite collecting successful data with the FM-CW, a 3-mm wavelength radar from the University of Massachusetts at Amherst was soon adopted in part because of its enhanced resolution (0.7° compared to the FM-CW's 5° beamwidth; Bluestein et al., 1995). With increased resolution of analyses and improved scanning rates, the benefit of ground-based, mobile radars was clear. Mobile-radars would change tornado research forever, and continue to improve our understanding of such phenomenon with each new study.

On 26 May 1997, near Glenpool, OK, 60 seconds of dual-Doppler data were collected with two ground-based, mobile, X-band Doppler radars (Wurman et al., 2007b). Despite the short length of time in which the radars were synchronously collecting data, the results showed the major improvements of this technique, which had the capability of resolving substantially finer details with remarkable scan-to-scan consistency. Even though X-band tends to suffer from attenuation in high precipitation events, this study showed that given the average range of mobile radars from their target storms (often $< 10\text{km}$), the decrease in resolution at longer wavelengths is not worth the trade-off for these types of studies. DOW radars also briefly collected dual-Doppler data of a weak tornado on 21 May 1998 (Wurman et al., 2007a). Although the data collection period was short, they were able to capture one such time in which both radars were scanning. The analysis further confirmed the importance of tilting of horizontal vorticity at the cusp between the RFD and updraft, although the magnitudes of these effects could not be diagnosed from observations at just one time. Wurman et al. (2007a) ultimately concluded that in spite of the quality of data that had been collected, the need for rapid scan to observe changes in the storm more quickly was crucial.

A number of cases in the early 2000s using only mobile, X-band radars brought about a number of studies in which data resolution was finally high enough (isotropic, 3-D grid size of 100 m) to resolve finer-scale features than had ever been observed before. Beck et al. (2006) analyzed “the first ground-based high-resolution dual-Doppler dataset of a cyclic supercell,” which occurred near Kress, TX, on 29 May 2001, albeit non tornadic. Subsequent cases from Marquis et al. (2008) of a tornadic supercell near Crowell, TX, on 30 April 2000, and Wurman et al. (2010) near Orleans, NE, on 22 May 2004, observed and documented secondary surges of RFD, a feature that had been rarely observed prior to this time. Marquis et al. (2008) remarked that this was likely due to a paucity of high-enough-resolution

data to resolve these surges as opposed to the rarity of these features in supercells. Frame et al. (2009) analyzed the 23 May 2002, Lipscomb County supercell and found that parcels originating in the forward flank *did not* end up in the updraft like suggested in previous models. It was suggested that the lack of a baroclinically generated vorticity could have played a role in the weakness of the low-level mesocyclone in this case. Tanamachi et al. (2012) performed a dual-Doppler analysis of the 4 May 2007, Greensburg, KS, tornado using a WSR-88D and the UMass X-pol mobile radar. Polarimetric observations were also examined with respect to the dual-Doppler synthesis. In this case, “mode-1” tornadoes and “mode-2” tornadoes were observed. In a mode-1 case, tornadoes that moved left of the parent updraft were weak and short-lived, while the mode-2 tornado (the violent Greensburg tornado) moved with the parent updraft. These findings were consistent with previous studies from Dowell and Bluestein (2002a). Although many of these studies were “high resolution,” they all still lacked the ability to resolve a tornado-scale circulation adequately (Kosiba and Wurman, 2010). In order to answer some of the most elusive questions about tornadoes, the push for even higher spatiotemporal resolution continued.

In 2009 and 2010, the second Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX2; Wurman et al., 2012) took place. This project built on the first VORTEX experiment, bringing together recent findings and technological advances from the 15 years between VORTEX and VORTEX2. This project sought to provide even further understanding to the origins of low-level rotation in tornadoes, continuing to build on those from VORTEX. Numerous data were collected and analyzed on multiple storms spanning the two year period of the field project, many of which included dual- and multiple-Doppler analyses (e.g., Marquis et al., 2012; Atkins et al., 2014; Skinner et al., 2014; Wienhoff et al.,

2014; Davenport and Parker, 2015). On 5 June 2009, data were collected on a tornadic supercell in Goshen County, WY. This tornadic supercell was the among the most widely documented cases in history (e.g. Markowski et al., 2012a,b; Kosiba et al., 2013; Wakimoto et al., 2011, 2012; Atkins et al., 2012), with seven mobile-Doppler radars (DOW6, DOW7, SR1, SR2, MWR-05XP, NOXP, XPOL), one WSR-88D (KCYS), and mobile sounding units all simultaneously collecting data on this storm. Data from this case document the tornado (and parent supercell) from tornadogenesis through dissipation. Analyses and results from the Goshen County studies will discussed with respect to this study in chapter 4.

The history of tornado observations, specifically through the use of dual-Doppler analyses, have improved drastically from their beginning more than 40 years ago. The next 15 years will continue to see increases in mobile radar technology, with phased-array radars becoming increasingly common. As we continue to build our dual-Doppler data set, and aim to continue improving data quality, we hope that our observational understanding of tornadoes will continue to make drastic improvements.

1.2 Overview of Polarimetric Observations

1.2.1 Review of Differential Reflectivity

Differential Reflectivity, or Z_{DR} , is a dual-Polarimetric variable that is calculated by

$$Z_{DR} = 10 \log \frac{Z_H}{Z_V} \quad [dB] \quad (1.27)$$

where Z_H and Z_V are the horizontal and vertical radar reflectivity factor, respectively. Radar reflectivity factor is defined as

$$Z_{H,V} = \frac{\lambda^4}{\pi^5 |K_w|^2} \int_0^\infty \sigma_{bH,V}(D) N(D) dD \quad (1.28)$$

where $|K_w|^2 = \frac{m^2-1}{m^2+2}$ is the dielectric factor with m representing the complex index of refraction, λ is the radar wavelength, $\sigma_{bH,V}$ is the backscatter cross-section for horizontally/vertically polarized waves, and $N(D)dD$ is the number of drops with a diameter D for a particular range of drop sizes dD .

For the purpose of this paper, Z_{DR} will be used to examine the potential an updraft has to loft large, oblate raindrops (relatively high Z_{DR}) into regions of smaller, more spherically shaped drops or ice particles (relatively low Z_{DR}). These " Z_{DR} Columns" and their importance will be discussed in the next section.

1.2.2 Differential Reflectivity Observations in Convective Storms

Since polarimetric capabilities were introduced to meteorological radars several decades ago, dual-pol products have been instrumental in our understanding of the microphysical structure of weather phenomena such as supercell thunderstorms. One such feature that has been frequently observed since the early 1980s is the presence of a vertical extension of relatively high Z_{DR} above the freezing level, known as a Z_{DR} column. Defined by Kumjian et al. (2014), a Z_{DR} column is a narrow vertical extension of positive Z_{DR} above the environmental freezing level.

Although this feature has been documented for many years, only recently has its relation to storm intensity been investigated. The presence of a Z_{DR} column was first documented by Hall et al. (1980); however, this discovery was only mentioned and little explanation (or attempted explanation) was given as to the cause of this feature. Hall et al. (1984) again found evidence of a Z_{DR} column extending 1.5 km above the melting layer. The presence of these relatively high values of both Z and Z_{DR} with no change in the height of the melting layer throughout the rest

of the analysis suggested that the updraft was responsible for lofting supercooled droplets above the melting layer.

In 1986, during the MIST field experiment (Dodge et al., 1986), Tuttle et al. (1989) collected dual-pol and triple-Doppler data using three National Center for Atmospheric Research (NCAR) radars. Data from this experiment again showed a vertical extension of the Z_{DR} column 2 km above the melting layer near the triple-Doppler-derived updraft maximum. With each subsequent updraft pulse, coincident Z_{DR} columns were observed at each time with the most intense updraft pulses resulting in the deepest Z_{DR} column tops, which intuitively suggests that stronger vertical velocity have the capability of lofting larger drops higher in the storm resulting in the observed Z_{DR} columns. Once the storm begins to form a downdraft, the Z_{DR} column disappears as the storm updraft begins to decay

Bringi et al. (1991) used aircraft radar data and particle probe data to analyze Z_{DR} and the storm updrafts with respect to Z_{DR} columns. Particle probes observed similar features to those describe in previous studies (i.e. Caylor and Illingworth, 1987; Illingworth, 1988) in which the beginning stages of cloud development resulted in large raindrops in low concentration, leading to larger concentrations of smaller drops as the cloud extends vertically, accompanied by the growth of the Z_{DR} column. Bringi et al. (1991) also observed similar vertical extension depths above the melting layer, near 2 km. The largest Z_{DR} was also observed below the melting level, displaced from the largest Z suggesting that collection of drops as they fall through the updraft can maintain the Z aloft yet still result in the highest Z_{DR} at a much lower altitude.

In the spring of 2002, The Joint Polarization Experiment implemented dual-polarization capabilities to the KOUN radar in Norman, OK, designed to test the usefulness of dual-pol products for National Weather Service forecasters (Schuur et al., 2004). Scharfenberg et al. (2004) used this data to examine the usefulness of

Z_{DR} columns in thunderstorm forecast development by examining the Z_{DR} columns throughout the lifetime of several supercell thunderstorms. They concluded that not only was it useful for understanding the formation, location, and strength of the thunderstorm, but it offered useful applications to other entities such as aviation.

Kumjian et al. (2014) also suggested that the use of Z_{DR} columns in research and forecasting applications was “strongly advocated”. They found a positive correlation between the Z_{DR} column height and the strength of the updraft, which could be used as a proxy for determining a storm’s severity. They also found a strong, positive-lagged correlation between Z_{DR} column height and significant hail reports at the surface, providing much greater forecasting lead time of the occurrence of significant hail than previous indicators.

Snyder et al. (2015) concluded that use of the Z_{DR} column depth and trend thereof may provide useful insight to forecasters for issuing warnings on marginally severe storms. It was suggested that Z_{DR} columns do provide strong correlation to convective updrafts, both in observations and simulations; however, radar volume coverage patterns can provide complications to the observations, especially for shallower columns far from the radar. Use of such observations can provide the ability for forecasters to anticipate severe weather events and issue earlier warnings. In addition, further research into characteristics of Z_{DR} columns could lead to better understanding of storm evolution and hazardous weather threats (Snyder et al., 2015). Examination of Z_{DR} columns and updraft intensities from this case will be further examined in Chapter 5 and compared to previous studies, such as the aforementioned.

Chapter 2

Overview of Instrumentation

In this chapter, we will discuss the specifications of the two instruments used in this study: a mobile, rapid-scan, X-band, polarimetric, Doppler radar, and the Twin Lakes WSR-88D radar.

2.1 A Mobile, Rapid-Scan, X-band, Polarimetric, Doppler Radar (RaXPol)

The mobile, rapid-scan, X-band (3.1 cm wavelength), polarimetric, Doppler radar (RaXPol) is a high-powered radar unit which is mounted on a Ford F-550 truck chassis (Fig. 2.1; Pazmany et al., 2013). Maintained by the Advanced Radar Research Center at the University of Oklahoma, RaXPol can mechanically scan at a rate of $180^\circ s^{-1}$, making it one of the fastest mechanically-scanning radars in the world. Some selected radar specifications are discussed in Table 2.1.

Rapid-scan data collection provides many benefits over slower-scanning radars, most notably the ability to collect high-temporal resolution data of storm-scale data such as tornadoes and supercells. In addition, the “traditional”, mechanically-scanning setup of RaXPol has the added benefit of dual-polarization and higher spatial resolution capabilities unlike some other rapid-scanning radars, such as

phased-array antennas, that often suffer from wider beamwidths and are typically only singly polarized. However, faster scanning does provide a number of challenges to the data collection process. One such challenge is that the accelerated scanning rate means the antenna will spin one beamwidth in 5.6 ms, giving it significantly less time for pulse averaging than some of its slower scanning counterparts. Discussed by Pazmany et al. (2013), RaXPol typically averages 12 pulse pairs in 4.8 ms, and employs frequency hopping to do so. Frequency hopping is a technique that involves shifting the frequency of the radar pulses by a minimum of one band width to both speed the radar parameters to their respective means faster, and mitigate second trip echoes as with such a short PRT. This technique allows RaXPol to collect a larger number of samples faster than it would be able to without the use of frequency hopping.

To reduce the effects of beam smearing, RaXPol has the ability to work in “strobe mode” in which a batch of pulses at different frequencies are transmitted simultaneously and received simultaneously, allowing for a number of independent samples to be collected at once in order to preserve the angular resolution. One drawback to transmitting such a large number of pulses simultaneously is the associated blind range caused by a long period of transmitting before switching to receive. This is potentially problematic due to the close proximity of the instrument to the tornado and other features of interest.

2.2 KTLX: The Twin Lakes WSR-88D Radar

Across the United States, the NEXRAD (Next-Generation Radar) network is comprised of 160 S-band radars, called WSR-88Ds (**W**eather **S**urveillanc**e** **R**adar, **1988**, **D**oppler). KTLX is located southeast of Oklahoma City, OK, and provides radar coverage for most of central Oklahoma. Like RaXPol, the WSR-88Ds have dual-polarization capabilities allowing them to produce dual-pol fields like



Figure 2.1: RaXPOL collecting data on a nontornadic supercell on 20 April 2014, at Lake Thunderbird near Norman, OK. Photograph by the author.

	RaXPOL	KTLX
Center Frequency	9.73 GHz $\pm$ 20 MHz	2.91 GHz
Transmit Power	20 kW	475 kW
PRT	0.250 msec	0.821 msec (V_r)
Pulse Width	0.1 μs - 40 μs	1.57 μs - 4.5 μs
Antenna Diameter	2.4 m	8.5 m
Half-Power (3 dB) Beamwidth	1.0°	0.91°
Antenna Gain	44.5 dB	45 dB
Maximum Antenna Rotation Rate	180° s^{-1}	30° s^{-1}

Table 2.1: Selected characteristics of RaXPOL and the KTLX WSR-88D.

Z_{DR} , cross-correlation coefficient (ρ_{hv}), and specific differential phase (K_{DP}). The WSR-88D, however, scans at a much slower rate than rapid-scanning radars like RaXPol; thus, the temporal resolution of the data suffers. On days when severe convection is present inside a particular WSR-88D domain, volume coverage patterns (VCPs) are adjusted accordingly to provide better coverage for forecasters. On 19 May 2013, KTLX was in VCP 212, which includes the following elevations scanning at a rate of $30^\circ s^{-1}$: 0.5° , 0.9° , 1.3° , 1.8° , 2.4° , 3.1° , 4.0° , 5.1° , 6.4° , 8.0° , 10.0° , 12.5° , 15.6° , 19.5° . In VCP 212, KTLX will complete one full volume scan in just under 4.5 minutes, focusing more of the volume on the low-level scans in an effort to detect hazards near the ground. This study will focus on utilizing this particular data, in concert with data from RaXPol, and accounting for differences in both temporal and spatial resolution to produce a dataset that is useful in examining some of the storm-scale features of these tornadic supercells.

Chapter 3

Methodology

3.1 Data Quality Control

In general, radar data quality is subject to a number of variables, including the number of samples and air motions associated with the volume being observed by the radar. In the case of RaXPol, faster scanning means that the data are subject to a relatively small number of samples (compared to WSR-88D data) and beam smearing. In addition, the data examined in this paper include scans of two violent tornadoes. The turbulent motions associated with these tornadoes, and areas of high lateral shear, often cause the data to have high spectrum width near tornadoes, and especially near the ground where large amounts of debris may be present. For this reason, quality control of the data is extremely important to the analyses.

Data were edited by hand using the National Center for Atmospheric Research’s SOLO3 software. The data were first thresholded on Normalized Coherent Power below 0.24, a signal-quality index which utilizes both signal-to-noise ratio (SNR) and spectrum width. Ground clutter was removed from the lower-level sweeps, and the velocities were dealiased. Velocity aliasing often occurred within the tornado and at anvil level where the wind speeds often exceeded the 31 m s^{-1} Nyquist velocity. Velocity aliasing was particularly prevalent during RaXPol deployment

3 where the Nyquist velocity was only 15 m s^{-1} . Finally, data were despeckled to remove remaining noise from low SNR data regions.

3.2 Observation Processing and Wind Synthesis

Data from both RaXPol and KTLX were mapped to a Cartesian grid following editing using Observation Processing and Wind Synthesis. To do so, a two-pass Barnes analysis was used (Barnes, 1964). Objective analysis parameters were then chosen based on the coarser of the two radar’s resolutions valid at the location of the mesocyclone. More specifically, before each two-pass Barnes analysis was completed, the maximum distance of the mesocyclone from the radar locations was carefully measured and this distance was used to calculate the grid parameters.

On 19 May 2013, data were collected during six deployments (Table 3.1). For deployment 2 (Carney-Edmond tornado), the dual-Doppler lobe size (Fig. 3.1) was large enough that objective analyses were performed twice with both grid spacings of 250 m and 500 m . This was done in order to maximize the resolution of the analyses where possible, but also to be consistent with $\delta = \Delta/2$, where Δ represents the raw data spacing and δ represents the grid spacing in Cartesian coordinates (Koch et al., 1983). Smoothing parameters of $\kappa = 0.716 \text{ km}^2$ and $\kappa = 1.077 \text{ km}^2$ for 250 m and 500 m analyses, respectively, consistent with $\kappa = (1.33\delta)^2$ (Pauley and Wu, 1990). Additionally, a second-pass convergence parameter (γ) of 0.3 was used (Majcen et al., 2008). To account for advection of the upper-level sweeps during volume scans, the speed of the mesocyclone was determined by tracking the speed of the updraft (Z_{DR} column). For both resolutions, the mesocyclone motion was taken to be $(u, v) = (12.6, 4.6) \text{ m s}^{-1}$. Accounting for storm motion by tracking the mesocyclone allowed the upper-levels to be corrected for advection that occurred while the volume scan was taking place. Although the advection at

Deployment	Time	Location
1.1	2016-2020 UTC	North side of Guthrie, OK
1.2	2040-2048 UTC	North side of Guthrie, OK
2*	2145-2221 UTC	3 mi. ENE of Wellston, OK
3*	2301-2321 UTC	4 mi. NNW of Bethel Acres, OK
4*	2327-2331 UTC	2 mi. NW of Bethel Acres, OK
5*	2341-0005 UTC	3 mi. SW of Shawnee, OK
6	0119-0125 UTC	Okemah, OK

Table 3.1: Complete list of RaXPOL deployments from 19 May 2013.

Deployments considered in this thesis are denoted by (*).

each level is undoubtedly variable, this method provides a reasonable correction especially near the updraft and tornado/mesocyclone.

For deployments 3 and 5 (Norman-Shawnee tornado; Fig. 3.2), a grid resolution of 250 m was used in both cases, along with a smoothing parameter $\kappa = .400\text{ km}^2$ and a storm motion of $(u, v) = (8.55, 6.28)\text{ m s}^{-1}$. The above parameters are consistent with recommendations from Pauley and Wu (1990), and those used in the dual-Doppler analyses performed by Kosiba et al. (2013).

3.3 A Spatially Variable Advection Correction (SVAC) Technique

One of the major challenges associated with this work was attempting to account for the drastic difference in temporal resolutions between the two radars. To the author’s knowledge, this is the first time rapid-scan radar data and WSR-88D radar data have been combined to perform a dual-Doppler analysis, which further enhanced the challenge of finding a suitable solution. In order to account for

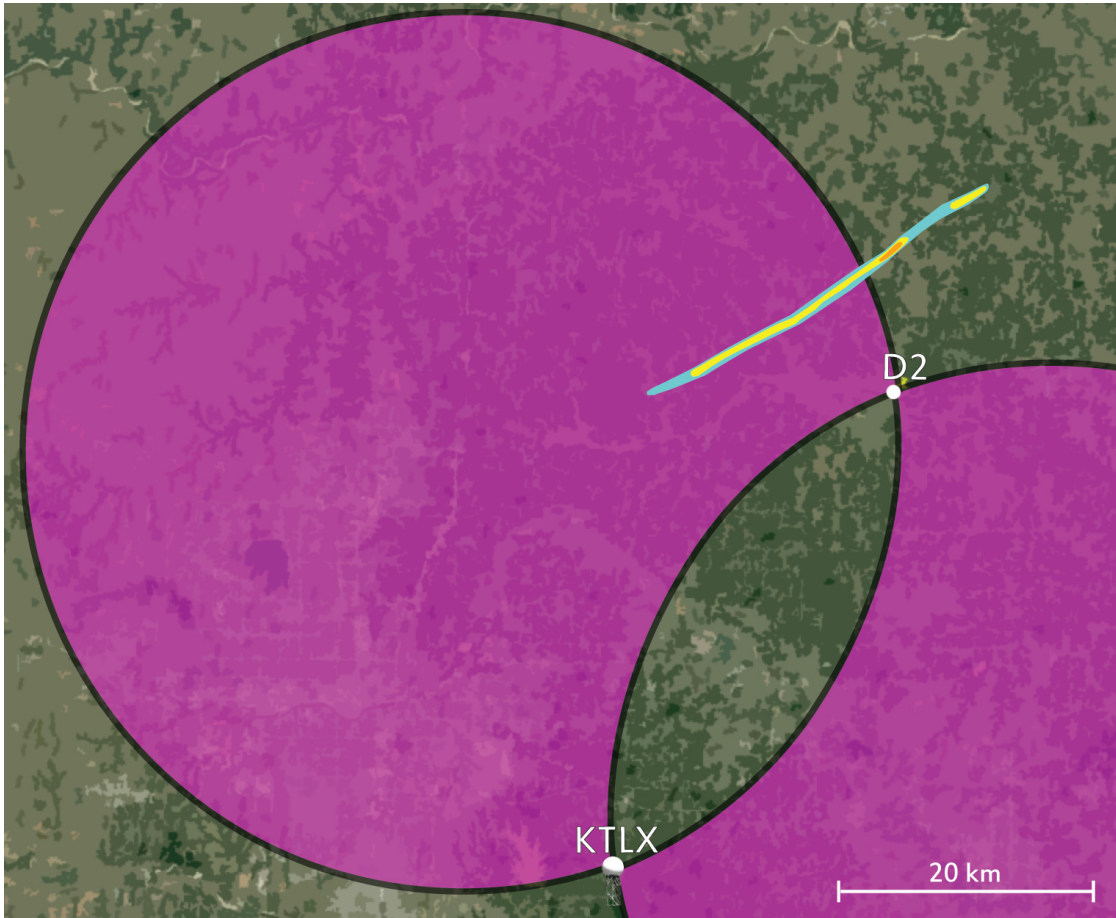


Figure 3.1: KTLX and RaXPol deployment 2 (D2) are shown with corresponding 30° dual-Doppler lobes. The NWS damage swath is also color contoured within the northern lobe. Blue, green, yellow, orange represent EF-0, EF-1, EF-2, and EF-3 rated damage, respectively. Although deployment 2 documented the tornado for 40 minutes, only 20 minutes were suitable for dual-Doppler analysis.

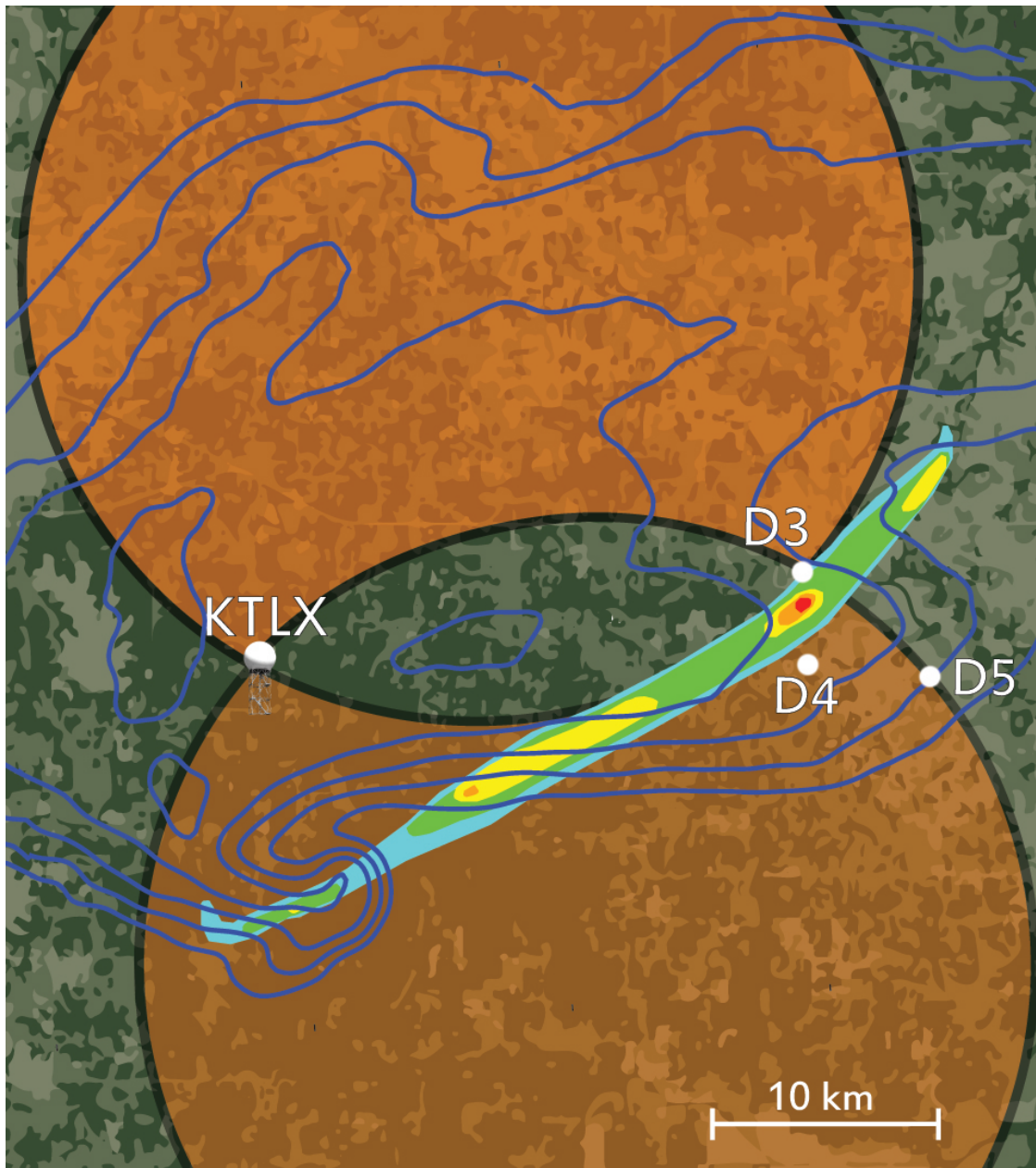


Figure 3.2: KTLX and RaXPol deployments 3, 4, and 5 (D3, D4, D5) are shown with corresponding 30° dual-Doppler lobes from deployment 3. The NWS damage swath is also color contoured within the northern lobe. Blue, green, yellow, orange, and red represent EF-0, EF-1, EF-2, EF-3, and EF-4 rated damage, respectively. Dark blue contours show the supercell at 23:02:22 UTC for scale perspective.

this difference, a spatially variable advection correction technique was used to linearly interpolate (in a Lagrangian sense) between two separate KTLX radar volumes using spatially variable U and V advection fields (Shapiro et al., 2010). This technique, which uses a cost function minimization with a constraint on the frozen-turbulence hypothesis, was first proposed by Gal-Chen (1982) for constant U and V fields. Gal-Chen (1982) showed that in order to obtain the constant advection fields, U , V , and $R(x, y)$ must be solved for iteratively to minimize the cost function. When applied to reflectivity fields, the frozen-turbulence hypothesis is represented through

$$\frac{\partial R}{\partial t} + U \frac{\partial R}{\partial x} + V \frac{\partial R}{\partial y} = 0, \quad (3.1)$$

where R represents the reflectivity. The frozen-turbulence hypothesis suggests that reflectivity is conserved throughout time. Because this is not exactly the case for real radar data, Shapiro et al. (2010) make use of a cost-function minimization with (3.1) as a weak constraint, resulting in the following cost function equation

$$J = \iiint \left[\alpha \left(\frac{\partial R}{\partial t} + U \frac{\partial R}{\partial x} + V \frac{\partial R}{\partial y} \right)^2 + \beta |\nabla_h U|^2 + \beta |\nabla_h V|^2 \right] dx dy dz, \quad (3.2)$$

where $\alpha = \alpha(x, y, t)$ is a “footprint function”, or a data coverage function which reflects where there is data within the analysis domain, and where there is not. The parameter β represents a constant positive smoothing parameter which is taken to be 100 dBZ^2 throughout this analysis unless otherwise noted.

Once the advection fields have been properly solved for, the radar data can be advected forwards and backwards in time to create linearly interpolated reflectivity fields. When given two consecutive radar volumes, the number of intermediate time steps is user specified and chosen to best match the new, interpolated volume scans to those volumes collected by RaXPOL. The U and V advection-corrected fields are also applied to fields such as radial velocity and Z_{DR} , creating new,

complete volumes of the radar data interpolated from real KTLX volumes. These new volumes were then paired with correspond data collected by RaXPol, and a dual-Doppler synthesis was performed and vertical velocities calculated using upward integration of the anelastic mass continuity equation.

The benefit to performing the analysis using the spatially variable advection correction was that we are able to maintain the value of rapid-scan radar data and supplement single-Doppler fields with a dual-Doppler synthesis. The main objective of this paper is to examine the validity of this technique, and weigh its potential benefit as a supplement to radar data where a second radar might be available but had previously been discounted due to differences in temporal resolution.

3.4 Analysis of Environmental Vertical Wind Shear

A portion of the work in Chapter 4 will focus on the temporal and spatial variation of the low-level vertical wind shear profiles in the near-storm inflow region of the supercell during the third RaXPol deployment on the Shawnee supercell, and in particular as it undergoes direct interaction with the tornadic circulation. For the purpose of this thesis, the near-storm inflow (hereafter referred to as inflow) is defined as the region south of the forward-flank gust front (FFGF) and east of the rear-flank gust front (RFGF). In this case, storm-relative hodographs obtained from the dual-Doppler analyses also discussed in Chapter 4 are examined both visually in terms of length and shape (i.e., how does the length and curvature change), and quantitatively in terms of 0-1 km and 0-3 km storm-relative helicity (SRH; Davies-Jones et al., 1990). From the 3D wind field obtained in our analyses, the following equation was used to obtain SRH:

$$SRH(c) = \int_0^h (\vec{v} - \vec{c}) \cdot \omega_h dz, \quad (3.3)$$

where $\vec{v}$ represents the vector wind field and $\vec{c}$ represents the storm motion. It is worth noting that because vertical velocity is calculated via upward integration, it is particularly sensitive and prone to errors associated with errors in the horizontal wind field. This can lead to substantial errors in our calculations of horizontal vorticity and subsequently SRH. Additionally, due to atmospheric refractivity and the Earth's curvature, it is important to note that hodograph evaluations at the surface are purely extrapolated as the beam failed to collect data below 150 m AGL at any time during deployment 3.

Chapter 4

Dual-Doppler analyses of the Carney-Edmond and Norman-Shawnee tornadoes

This section will briefly cover the synoptic setup from 19 May 2013, and will examine the results of the dual-Doppler analysis on both the Carney-Edmond and Norman-Shawnee Tornadoes.

4.1 19 May 2013: A Synoptic Overview

On 19 May 2013, a large-scale trough and significant mid-level speed maximum ejected eastward from the southern Rocky Mountains and overspread the Great Plains (Fig. 4.1a). Strengthening upper- and mid-level flow provided ample deep-layer shear in excess of 25 m s^{-1} supportive of supercells ($> 20 \text{ m s}^{-1}$), and 850 mb flow in excess of 15 m s^{-1} from the south-southeast provided sufficient low-level shear (Fig. 4.1b) to support tornadoes. The 1800 UTC sounding from Norman (Fig. 4.1e) sampled a classic “loaded-gun” environment with steep mid-level lapse rates in excess of $8.0 \text{ }^{\circ}\text{C km}^{-1}$, and a moist layer extending above 850 mb with a surface dewpoint of 72°F (22°C).

High CIN in excess of -100 J kg^{-1} , which was sufficient to suppress early-morning convection throughout the warm sector, was slowly decreasing during the

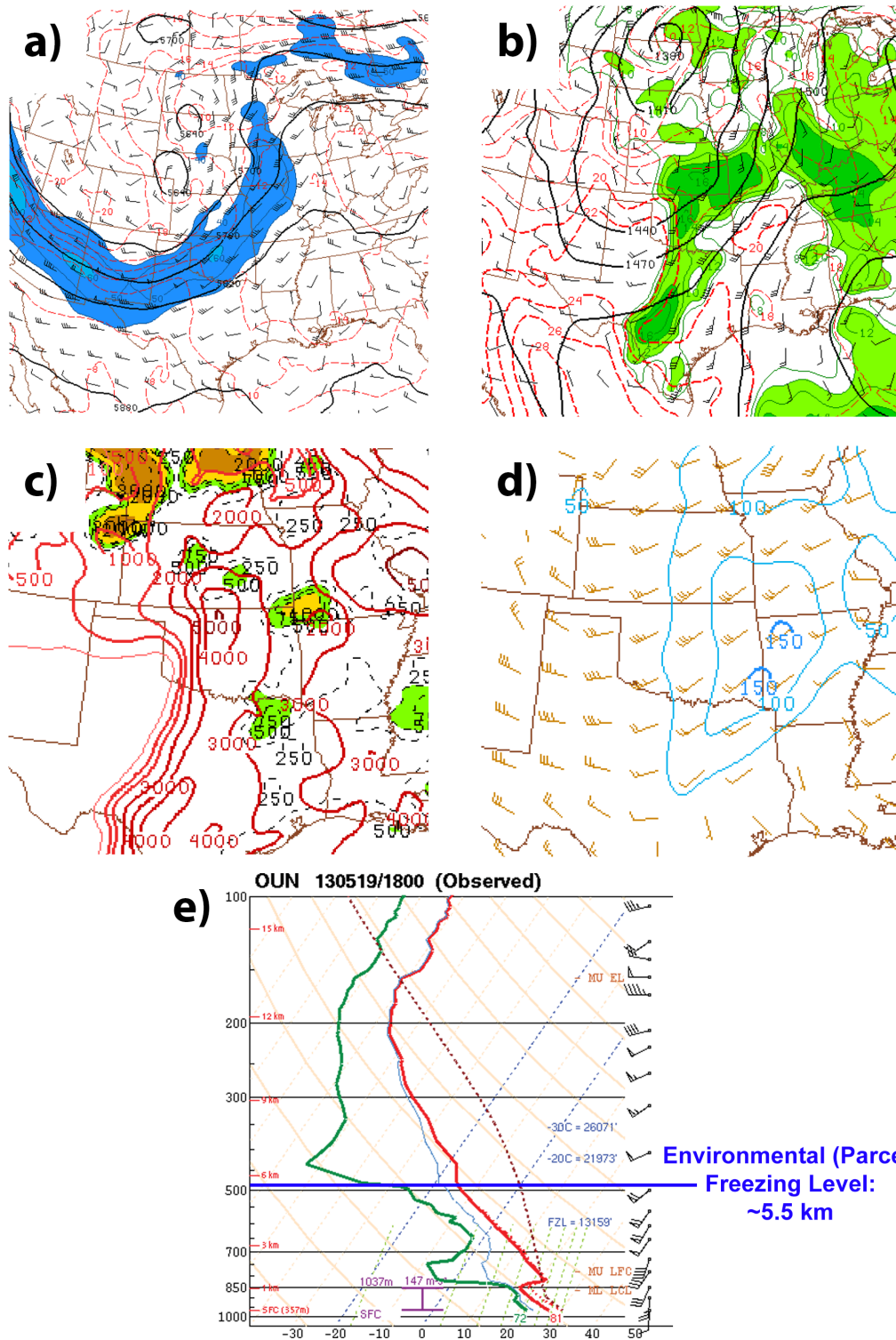


Figure 4.1: Overview of the synoptic setup on 19 May 2013 at valid at 18Z. Pictured above are (a) 500 mb winds, (b) 850mb winds, (c) MUCAPE, (d) 0-1 km SRH, and (e) 18Z sounding from KOUN (Norman, OK).

morning hours. A number of shortwave impulses impinging on central Oklahoma from the west aided in the removal of CIN through adiabatic cooling due to rising motion. Surface CAPE nearing 5000 J kg^{-1} and 0-1 km SRH in excess of $100 \text{ m}^2 \text{ s}^{-2}$ characterized the environment (Figs. 4.1 c and d; 0-1 km SRH of $147 \text{ m}^2 \text{ s}^{-2}$ as seen in the 1800 UTC OUN sounding (Fig. 4.1e). A deepening surface low over northwestern Oklahoma was associated with a rapidly-mixing dryline, which propagated eastward across Oklahoma toward the I-35 corridor. Increasing westerly flow aloft promoted enhanced effects of lee cyclogenesis, leading to a sharper surface trough and increased convergence along the dryline. As the convective temperature was reached due to strong boundary layer heating, the necessary conditions were in place for long-lived supercells with the potential for strong-to-violent tornadoes.

4.2 The Carney-Edmond Tornado

As the dryline propagated across central Oklahoma via vertical mixing, supercells formed just to the west of the I-35 corridor and moved east-northeast towards the Oklahoma City metro area. Initially, the northernmost storm, located north of the OKC metro, moved towards Edmond, OK. A maturing supercell with a southward protrusion resembling a hook echo moved toward Edmond at 21:41:01 UTC (Fig. 4.2), right at the time of tornadogenesis according to the NWS damage survey (21:41 UTC). The velocity couplet contained a maximum wind speed of 23.5 m s^{-1} (Fig. 4.2b). At 21:41 UTC, some semblance of a tornado debris signature was observed, particularly in reduced ρ_{HV} (Fig. 4.2c), but was still rather subtle as the tornado was weak. By 21:58 UTC, the tornado had strengthened substantially and a clear tornado debris signature was observed (Fig. 4.3). Low ρ_{HV} and Z_{DR} coincident with a debris ball in reflectivity and strong velocity couplet all indicated that a strong tornado was occurring (Fig. 4.3). RaXPol data collection began at 21:46:31 UTC; thus, dual-Doppler data collection commenced at this time (data

collection actually began several minutes earlier, but a radar hardware malfunction resulted in unusable data until this time). Figure 3.1 offers a glimpse of the field operation setup and shows the 30° dual-Doppler lobe that is considered in these analyses with color contours representing the location of the damage survey's estimated tornado track. Davies-Jones (1979) examined the effects of radar coverage on dual-Doppler analyses, noting that a between-beam angle of greater than 30° was satisfactory for resolving velocity components in the analysis. While the entire storm was not inside the 30° dual lobe, the tornado, mesocyclone, and updraft were inside the domain for a major portion of the data collection (approximately 20 minutes).

At dual-Doppler analysis times (Figs. 4.4-4.8), the horizontal wind field (black vectors), and black (blue) contours represent positive (negative) vertical velocity and vertical vorticity are shown over the color-filled KTLX reflectivity. In comparison with Fig. 4.2, the first dual-Doppler analysis, which is more than six minutes after tornadogenesis, exhibits a more organized storm structure, a supercell shape, and a hook echo. There is a strong confluence zone along the forward flank of the storm, which is maximized at low levels, and the wind field veers with height likely due to the combined effects of the lack of friction above the boundary layer and warm advection (veering environmental winds with height). Where friction plays a role (near the surface), cyclostrophic balance, or the wind balance between the pressure gradient force (PGF) and centripetal force (the apparent centrifugal force), is destroyed. Near the surface, the PGF dominates, resulting in a more easterly component (toward the mesocyclone) of the wind and a strong confluence zone along the forward flank which is not observed aloft. Additionally, strong convergence of inflow into the storm is observed, particularly in the lowest level (Fig. 4.4). Near-surface convergence associated with the updraft likely played a role in

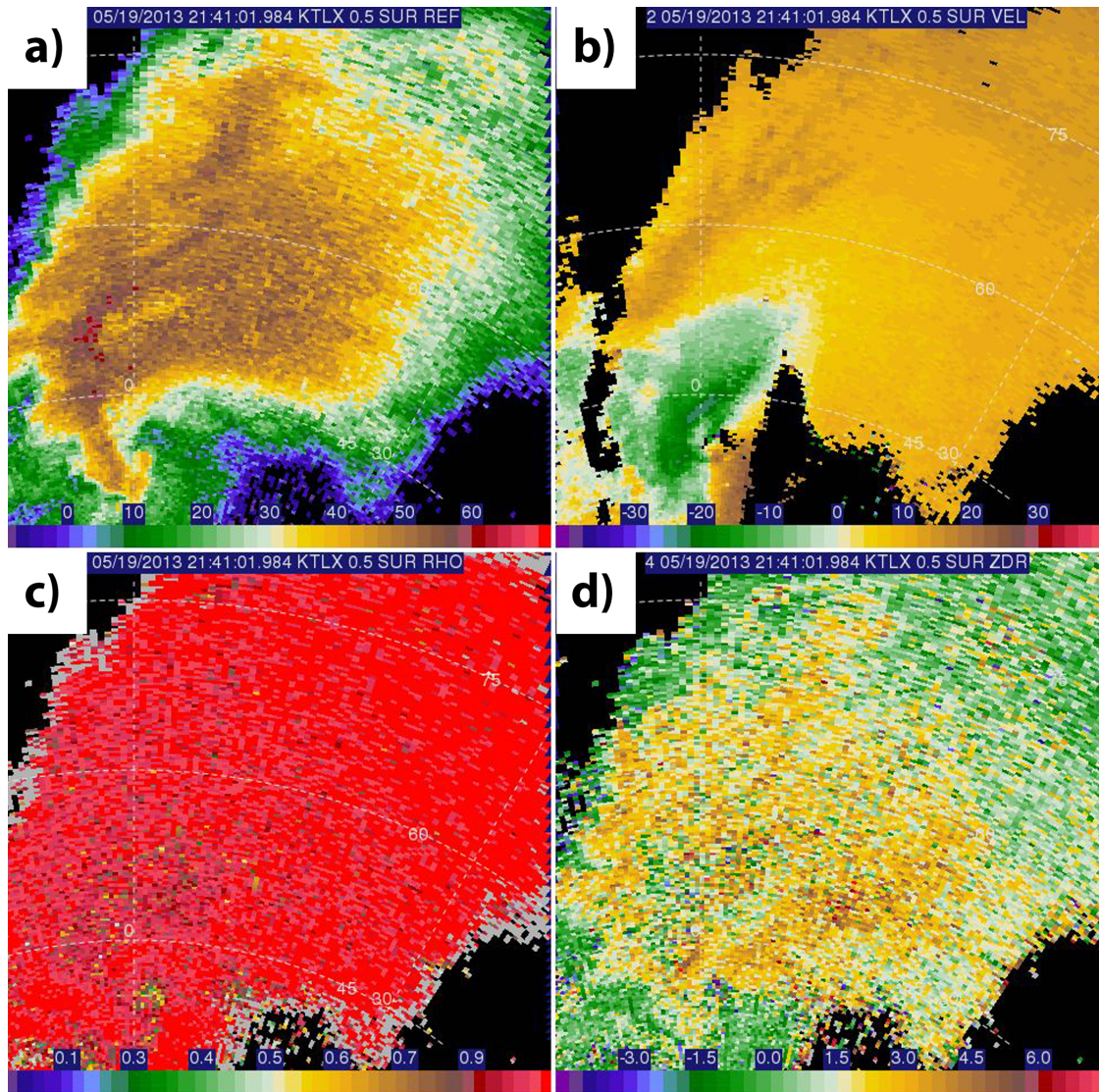


Figure 4.2: KTLX depiction (0.5° elevation) of the Carney-Edmond supercell at 21:41:01 UTC, right as tornadogenesis was occurring according to NWS damage survey reports. Radar variables shown are (a) radar reflectivity, (b) radial velocity, (c) cross-correlation coefficient, and (d) differential reflectivity.

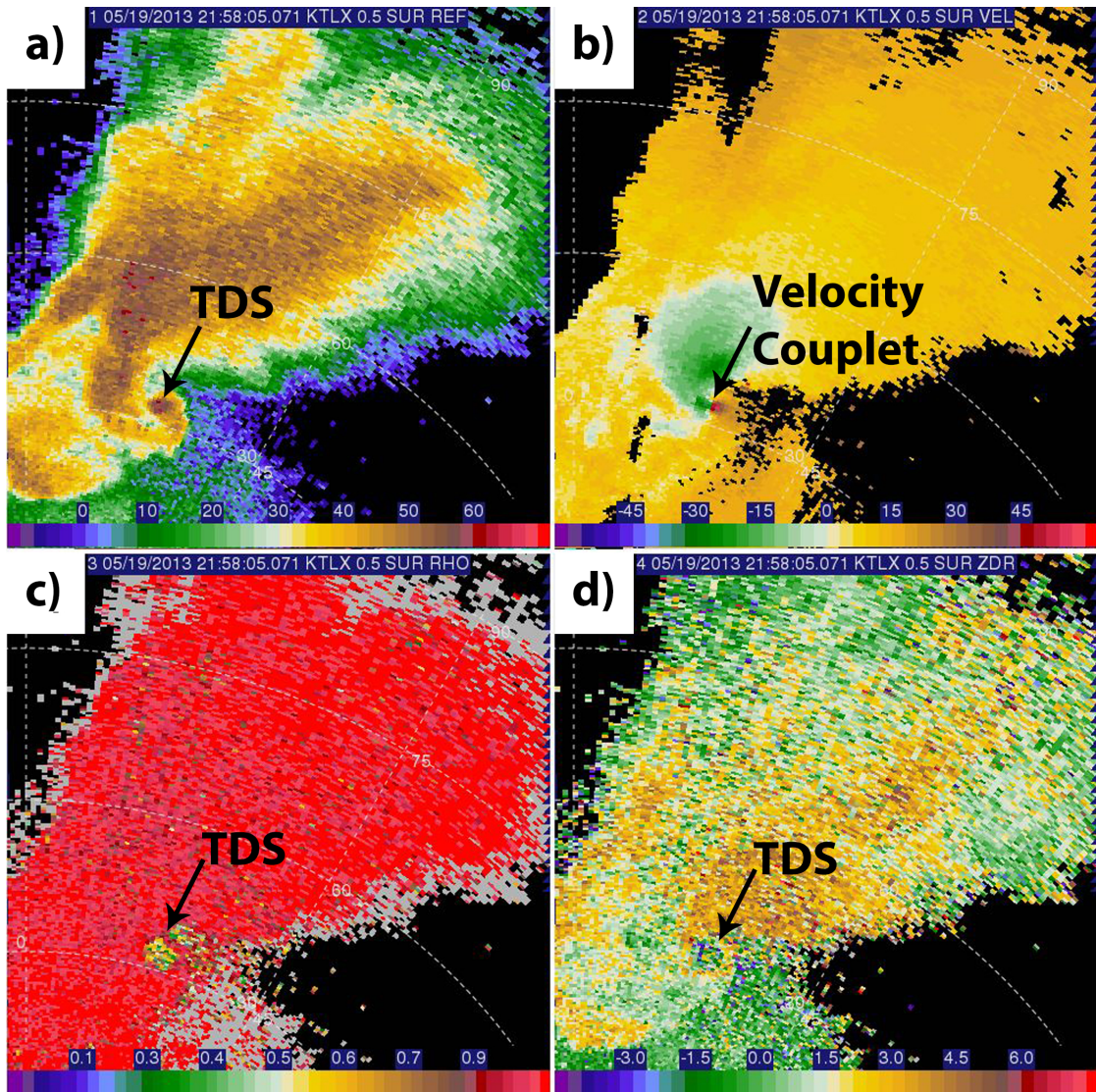


Figure 4.3: KTLX depiction (0.5° elevation) of the Carney-Edmond supercell at 21:58:05 UTC, just prior to the tornado's maximum intensity. Radar variables shown are (a) radar reflectivity, (b) radial velocity, (c) cross-correlation coefficient, and (d) differential reflectivity. The location of the tornado debris signature (TDS) is marked, for reference.

the intensification of vertical vorticity near the surface as well, strengthening the tornado (e.g., Ward, 1972; Lewellen, 1993; Dowell et al., 1997).

Strong wind speeds nearing 40 m s^{-1} were observed within the inflow region at 21:46:31 UTC, confirming that the near surface wind speeds were rapidly intensifying as the tornado strengthened and approached maturity. The maximum radial velocity near the surface (1.0° elevation) from RaXPol data was 40.2 m s^{-1} ; however, the largest magnitude of Doppler velocity (52.4 m s^{-1}) was recorded roughly 2.5 km above ground level (AGL), suggesting that early on the wind speeds were strongest aloft. The dual-Doppler analysis confirms that the wind field did intensify with height, but the observed magnitudes are drastically different. The maximum surface wind speed near the tornado was near 20 m s^{-1} , nearly half that of RaXPol measured maximum wind speed. With a preconceived idea that the wind speeds would be much stronger given the intensity of this tornado, this discrepancy was concerning. However, because the data are objectively analyzed to a **maximum** grid spacing of 250 m, it is expected that the analyzed wind speeds will be less, even significantly so, compared to raw RaXPol radial velocity data. Resolving any feature requires a minimum of three data points; thus, we are unable to resolve any features smaller than 500 m in scale. Therefore, although we are interested in the behavior of the tornado, we are unable to resolve the actual tornadic circulation itself but are more likely capturing a circulation somewhere between the scales of the tornado and mesocyclone. Smoothing of the data is one likely cause of the significant decrease in wind speeds observed in the objective analyses from those in the raw RaXPol data in the tornadic circulation, given that the more smoothing that is applied the less small-scale maxima are likely to be represented in the data. Another possible reason that the horizontal wind vectors fail to show stronger wind speeds is that the vectors are sampled every 1 km, making it unlikely that an individual vector will fall *exactly* on the radius of maximum winds within

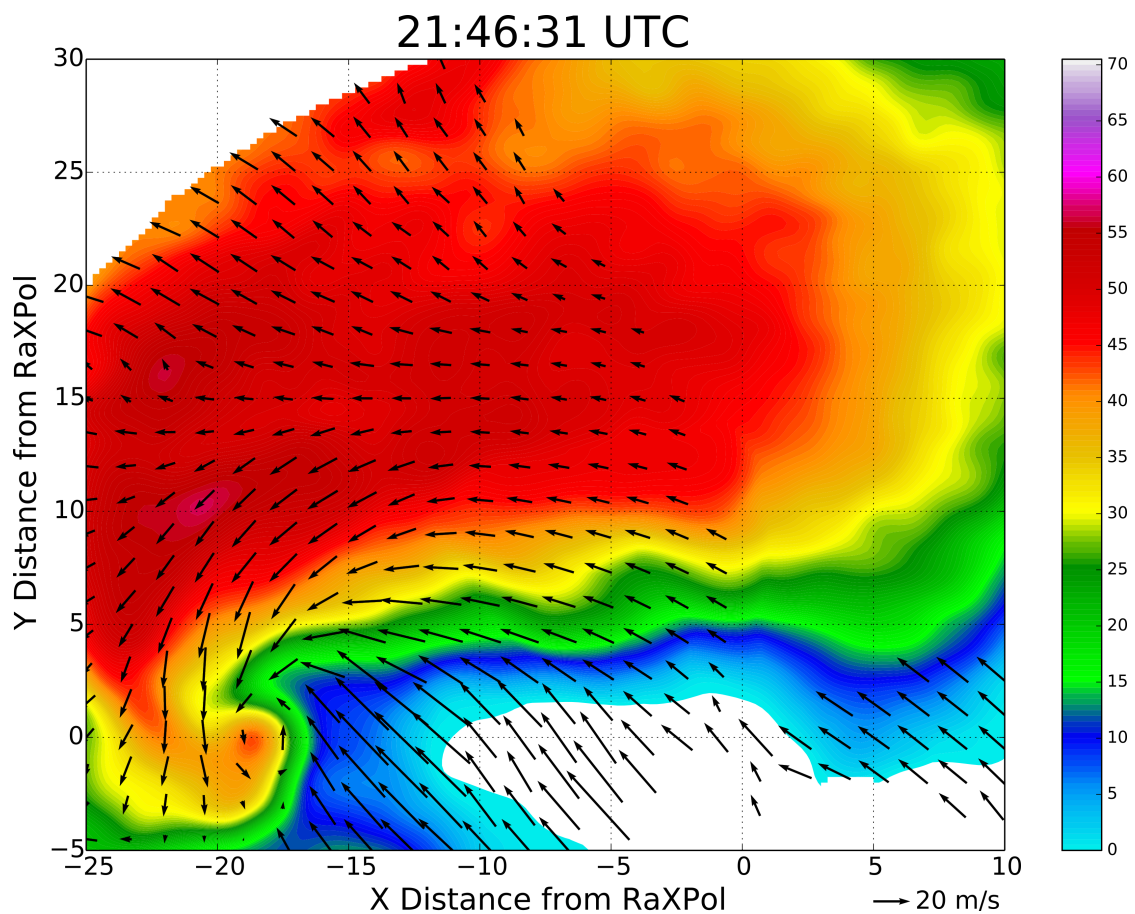


Figure 4.4: The first dual-Doppler analysis time from the Carney-Edmond tornado valid at 21:46:31 UTC. The tornado touched down approximately 6 minutes before the analysis began. This figure illustrates the supercell, with horizontal wind vectors overlaid. The tornado last approximately 40 minutes, but the storm moved out of the dual-Doppler lobe well before the tornado dissipated.

the circulation. Taking these into consideration, the analysis reasonably represents changes in both wind speed magnitude and direction with height, which can be equally useful in understanding how the storm is evolving.

With the primary focus of this study being placed on the updraft and dominant circulation, the domain has been cut and shifted in Figs. 4.5-4.8 from Fig. 4.4. Data are plotted at 750 m and 3.5 km AGL so that the mid-level circulation can be examined with respect to the low levels, and vice versa. Because the WSR-88D was sufficiently far from the supercell, at no time during the Edmond-Carney analysis are the data valid below 660 m (i.e., they are extrapolated below this level). For this reason, data below 750 m are not presented and will only be loosely considered in later examinations of vertical profiles of vertical vorticity.

Strong convergence and an associated region of $w > 10 \text{ m s}^{-1}$ was located on the eastern periphery of the hook echo at 21:46:31 UTC. Albeit still relatively weak at low levels, the updraft resembles a horseshoe-like shape which has been observed in numerous other studies (e.g., Ray et al., 1981). At 3.5 km, the updraft retains its same relative horseshoe shape, but the magnitude of the velocities have increased dramatically as expected. The main body of the updraft exhibits a maximum w in excess of 40 m s^{-1} . The location of the most intense updraft is close to, but not centered on, the bounded weak echo region (BWER¹; Chisholm, 1973) which is rather curious given that we would expect them to coincide with the strongest vertical velocities, which will be discussed further shortly. Another feature to the southwest of the main updraft is a small, yet pronounced, downdraft. The vorticity ζ at 750 m is $2.0 \times 10^{-2} \text{ s}^{-1}$, representing a moderately strong vortex at the location of the tornado. The circulation intensifies rapidly up to 2.5 km where it is greater than $2.5 \times 10^{-2} \text{ s}^{-1}$. The vertical vorticity ζ then slowly decreases with height

¹Although Chisholm (1973) was the first to use the term weak echo region (WER), or bounded weak echo region (BWER), alternate terms such as “chimney” or “vault” were used prior to this time to describe the same phenomenon in studies such as Browning and Donaldson (1963).

21:46:31 UTC

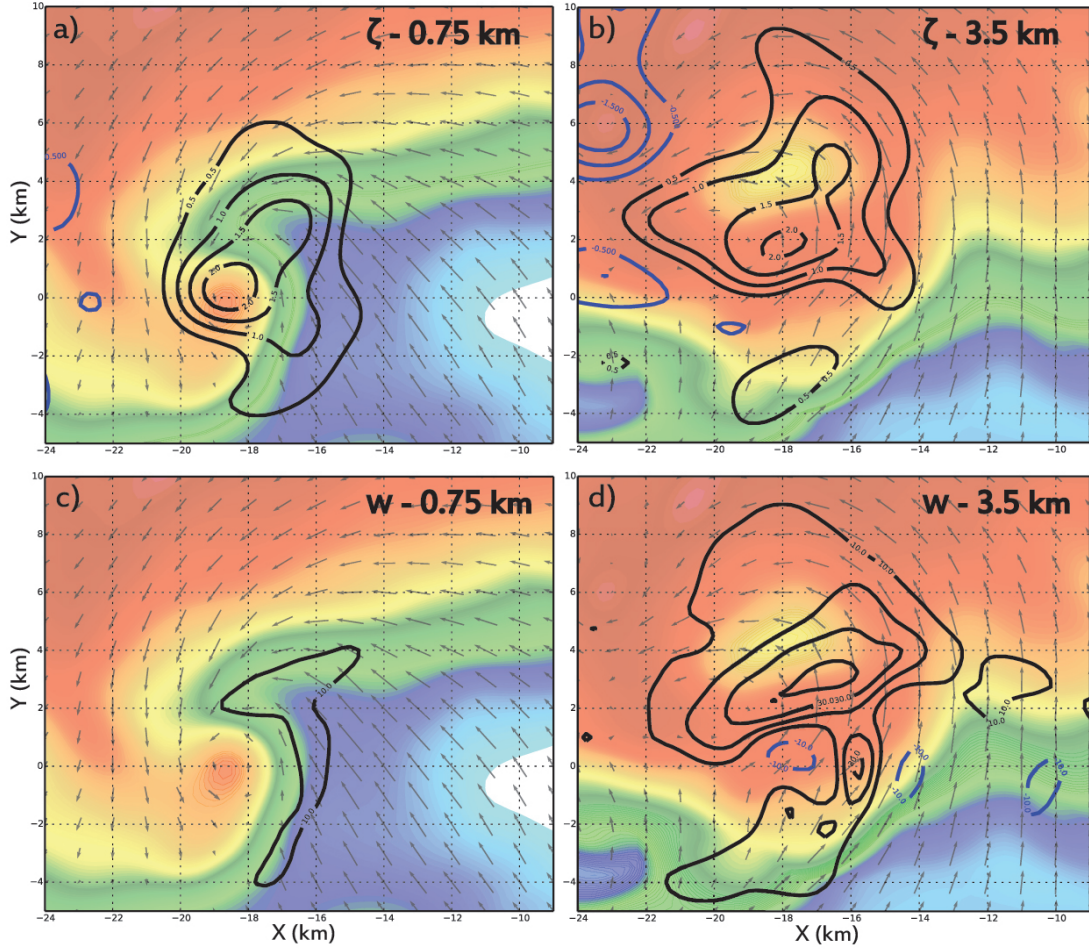


Figure 4.5: Dual-Doppler analysis valid at 21:46:31 UTC. Panels a and b show vertical vorticity near the hook echo and updraft region of the supercell at 0.75 km and 3.5 km, respectively, contoured every $0.5 \times 10^{-2} s^{-1}$. Panels c and d show the storm updraft at 0.75 km and 3.5 km, and are contoured every $10 m s^{-1}$. Reflectivity is shown (right), and a horizontal wind vector is shown for reference (bottom right).

21:49:59 UTC

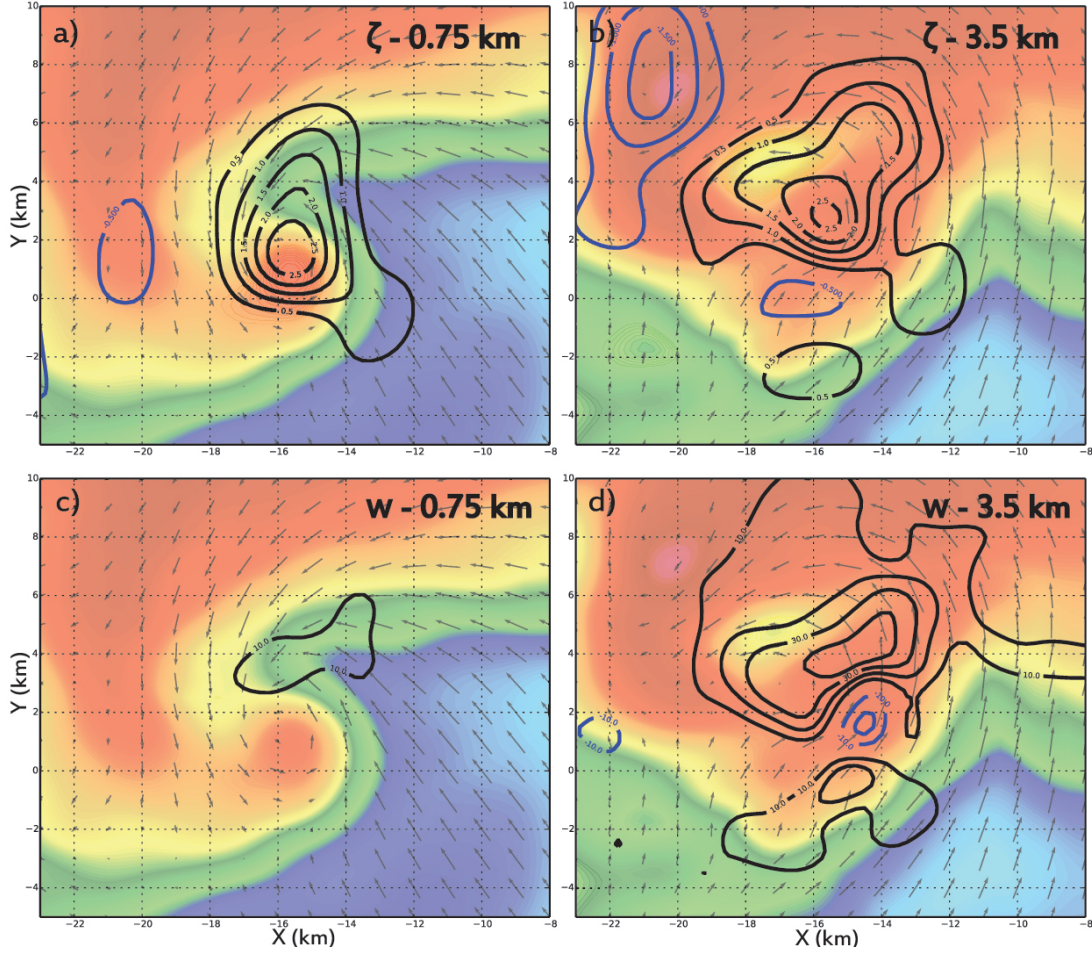


Figure 4.6: As in Fig. 4.5, but at 21:49:59 UTC.

21:53:27 UTC

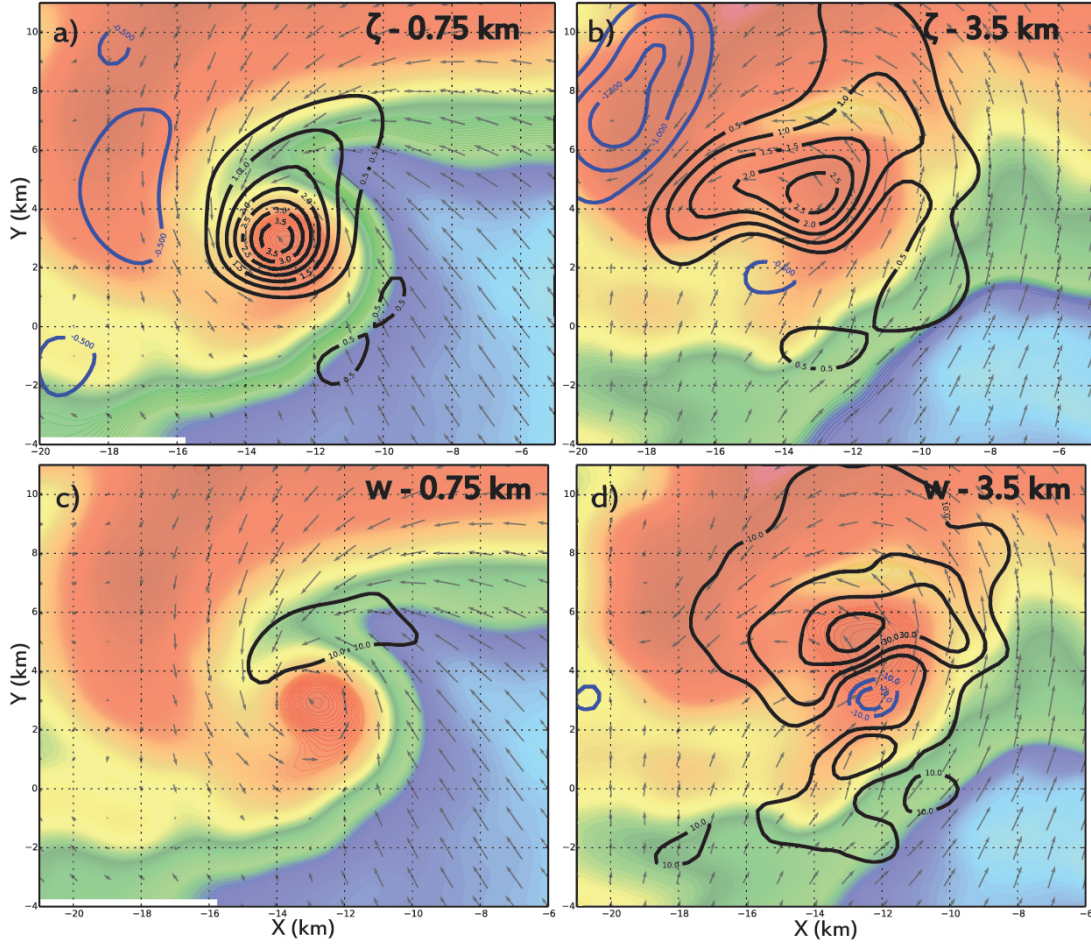


Figure 4.7: As in Fig. 4.5, but at 21:53:27 UTC.

21:57:46 UTC

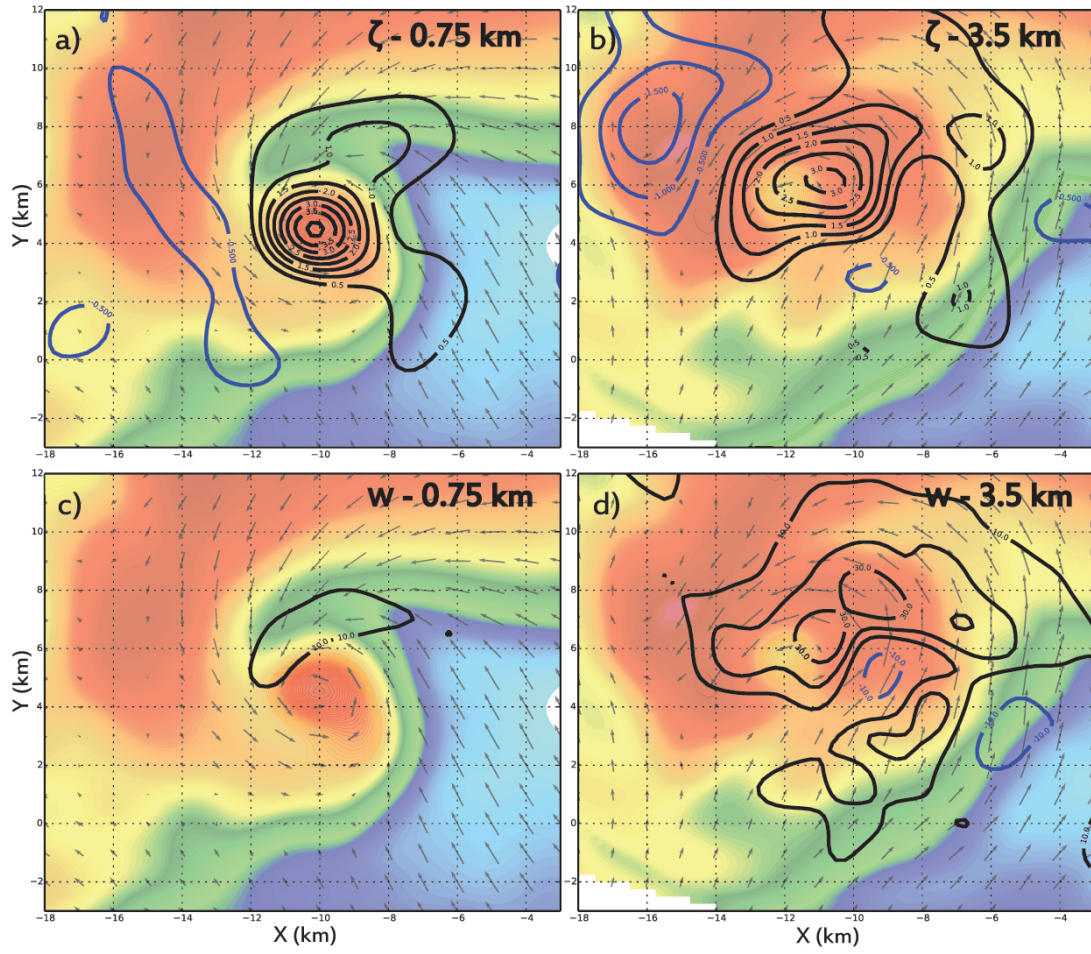


Figure 4.8: As in Fig. 4.5, but at 21:57:46 UTC.

farther into the mid levels. The tornado appears to tilt northeastward with height with the center moving almost 2 km north-northeast in the lowest 3.5 km. The maximum ζ also appear to coincide with the strongest updraft, both of which are displaced from the BWER.

At 21:49:59 UTC (Fig. 4.6), we see a number of similar features from the previous analysis. The updraft at 750 m is still located on the eastern periphery of the hook echo although has decreased slightly in size comparatively. The most intense updraft increases slightly in intensity although the main body of the updraft remains similar in both size and shape. As before, a small downdraft region south of the hook echo was still evident at this time. The vorticity ζ also intensified, both at 750 m and 3.5 km AGL, suggesting that the vortex continued to intensify at this time. Peak vertical vorticity was observed aloft near 2.5 km. Additionally, the anticyclonic couplet observed at 21:46:31 UTC broadened in size, but maintained its intensity, through 21:49:59 UTC.

At 21:53:27 UTC (Fig. 4.7) and 21:57:46 UTC (Fig. 4.8), the updraft continues to broaden with the same relative shape and intensity throughout, although it appears to weaken just slightly from 21:53:27 UTC to 21:57:46 UTC. At 21:57:46 UTC, the hook echo began to push northward towards the core, although this behavior was not observed aloft. The hook remains separated from the core with weaker ($< 30 \text{ dBZ}$) reflectivities to the north of the hook echo likely due to the presence of the updraft (and associated BWER). The vertical vorticity throughout the dual-Doppler analysis was steadily increasing in intensity. In each case, the steady increase in ζ near the surface corresponded to an increase in ζ aloft as well; the highest vorticity is found between 2 and 2.5 km AGL in each case. The vortex continues to lean with height but actually begins to lean more due north or north-northwest with height instead of the original north-northeast tilt that was seen very early in the deployment. At upper levels, and coincident with the downdraft

noted in all four w plots at 3.5 km AGL, a weak anticyclonic vortex is evident, on the south side of the primary circulation.

Observations of the BWER, which remained displaced from the maximum updraft through the dual-Doppler analysis, suggest some interesting and unexpected processes may be occurring. We generally expect that the updraft be collocated with the BWER since the BWER forms as a result of the updraft being too strong for raindrops to descend within it, or too strong to allow precipitation particles enough time to grow to radar detectable sizes (Browning and Donaldson, 1963). In this particular case, though, the two are separated with the BWER located slightly downstream of the maximum updraft with respect to the storm-relative wind field. This same result was seen in a number of studies, including Calhoun et al. (2013) and Markowski et al. (2012a). The Goshen County case (Markowski et al., 2012a) had strikingly similar comparisons between the updraft and the BWER, with the BWER again displaced downstream from the strongest vertical velocities. With time, the BWER begins to reside more closely to the strongest updraft. Additionally, the BWER is initially located between the cyclonic and anticyclonic vortex couplets. One reason for this observation likely deals with the maturity of the supercell. When the dual-Doppler observations began, the supercell and associated tornado had yet to reach a mature stage. When a supercell forms, initiation of an updraft tilts environmental horizontal vorticity vertically, creating a pair of counter-rotating vortices. With time, the cyclonic vortex in a right-moving mature supercell generally becomes collocated with the primary updraft. However, before this occurs, it seems possible that the BWER would be located in between the vorticity maximum and minimum until the storm reaches maturity. The reason that the vertical velocity is greatest slightly upstream could be due to the dynamic pressure effects associated with the strengthening cyclonic vortex, but a tilt of the updraft itself causes the observed displacement of these two features. The BWER

remains encompassed by the updraft throughout (i.e., $w > 10 \text{ m s}^{-1}$), but the maximum updraft could still lie upstream collocated with the ζ maximum for a period of time.

In each of the four cases there are notable and expected differences that occur, including a steady increase in ζ with time as the tornado slowly intensified, and an updraft that remained rather steady with time but slowly grew in spatial extent. Features such as the downdraft aloft south of the main circulation remain temporally consistent with time suggesting that they are real features and not simply artifacts of this analysis technique. While accuracy of the analyses are important, in cases such as this one it is nearly impossible to quantify the accuracy. Instead, scan-to-scan consistency with qualitatively reasonable results from previous observations and other studies are the best way to confirm the usefulness of this technique without additional data collection.

Timeseries graphs of ζ (Figs. 4.9 and 4.10) help quantify ζ at both 250 m grid spacing and 500 m grid spacing, respectively, and reflects the discussion from the previous figures. As previously mentioned, the analyses were rerun at 500 m resolution to extend the time period in which the analyses were valid. This helped us gain more understanding of the storm evolution after the 250 m analysis had concluded. The vertical vorticity in this case is slowly increasing with time as the tornado intensified. At 250 m grid spacing, vertical vorticity at all levels (considered here include 0.75 km, 1.5 km, 2 km, and 3 km) all steadily increase with time signifying an intensification of the tornado and associated low-level vertical vorticity. Vorticity at all levels is nearly equivalent when the analysis begins, near 0.025 s^{-1} . A steady intensification is evident at all levels early on; however, ζ at 750 m does not intensify as rapidly initially. The 500 m grid vertical vorticity timeseries (Fig. 4.10) spans 6 minutes longer than the 250 m grid analysis but shows a similar slow intensification. Although the intensification of the vorticity maximum is not

nearly as dramatic in the 500 m analysis, the slower increase is likely due to the decreased grid resolution and increased smoothing used in the objective analysis. In the 250 m analysis, the maximum in vertical vorticity comes at 21:57 UTC, when the peak ζ was in excess of 0.04 s^{-1} , after which it decreased slightly. The 500 m grid analysis start near 0.016 s^{-1} and slowly intensify. At approximately 21:57 UTC, the 500 m analysis shows a similar decrease in vertical vorticity like in the 250 m analysis, but is followed by another intensification immediately thereafter. This suggests continued intensification of the storm after the analysis had concluded.

The maximum ζ in the 500 m analysis occurred just prior to 22:00 UTC and is nearly constant at 0.025 s^{-1} through the end of the analysis. From observations of raw RaXPol data, it can be seen that the maximum in ζ is reasonable, corresponding to a maximum in radial velocity of approximately 74 m s^{-1} at 1.0° elevation at 21:59 UTC (Fig. 4.11b). Rapid intensification of velocities occur just prior to 21:58 UTC (Fig. 4.11a); maximum wind speeds are greater than 80 m s^{-1} just after 22:02 UTC (Fig. 4.11c). Maximum ζ at 250 m grid spacing in excess of $.04 \text{ s}^{-1}$ is consistent with previous studies, such as French (2006), in which 250 m grid spacing was also used. However, these vorticities represent strong low-level mesocyclones, but are an order of magnitude shy of those observed in high-resolution (single Doppler) observations of tornadoes (e.g., Wurman and Gill, 2000). In this case, we are certainly not resolving the tornado, but rather a larger vortex between the scales of the tornado and the mesocyclone.

Vertical profiles of maximum ζ (Fig. 4.12) are taken from the same times selected above in Figs. 4.5-4.8. In each case, vertical vorticity intensifies with time consistent with Figs. 4.9 and 4.10. At 21:46:31 UTC, the vertical vorticity is relatively uniform with height below 3 km, which is also consistent with information suggested in the timeseries. However, as ζ intensified, the profile exhibited a S-shape, with the vorticity maximized aloft somewhere between 1.5 and 3 km,

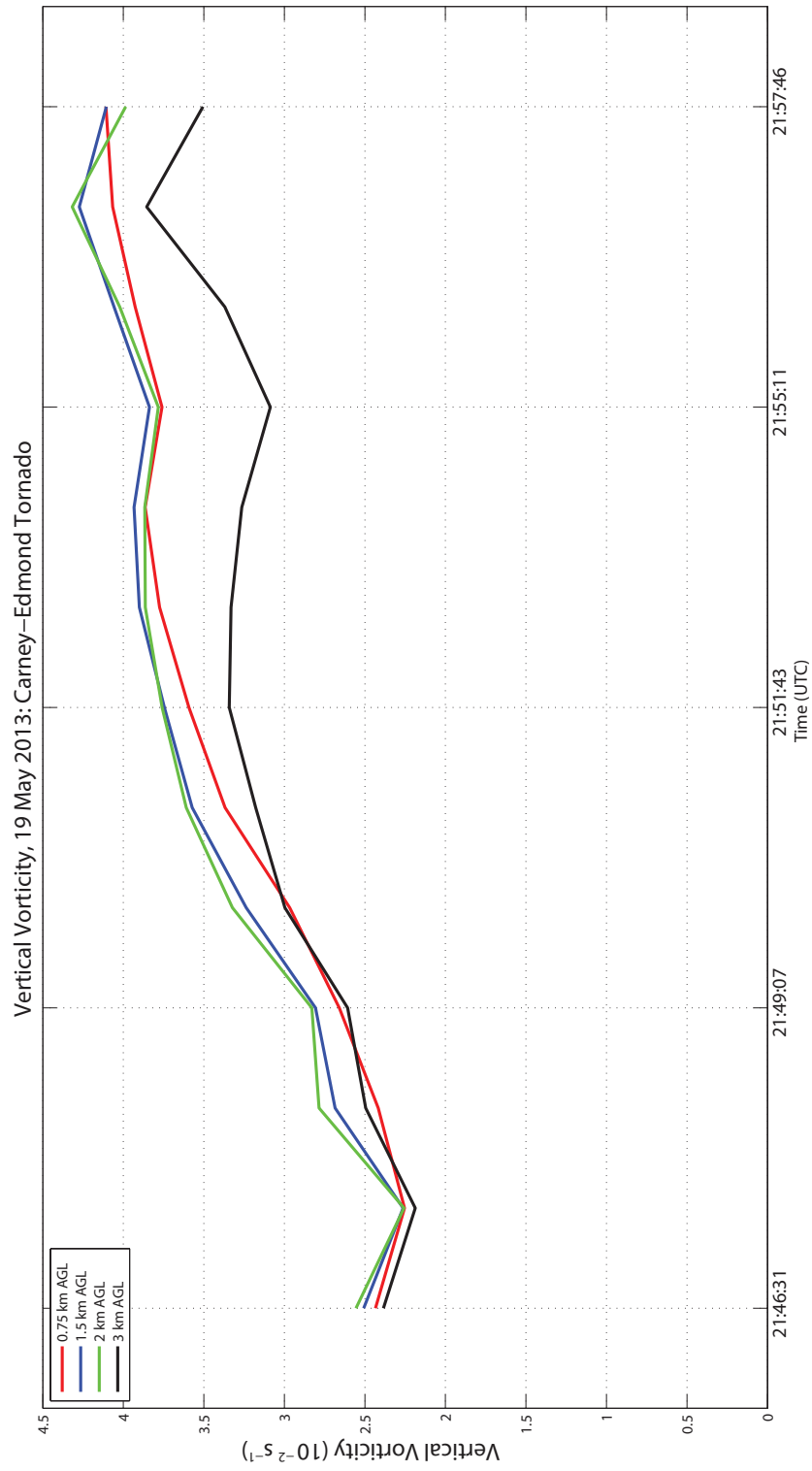


Figure 4.9: Vertical vorticity timeseries for the 250 m resolution analyses. Vertical vorticity ($10^{-2} s^{-1}$) is shown on the Y-axis.

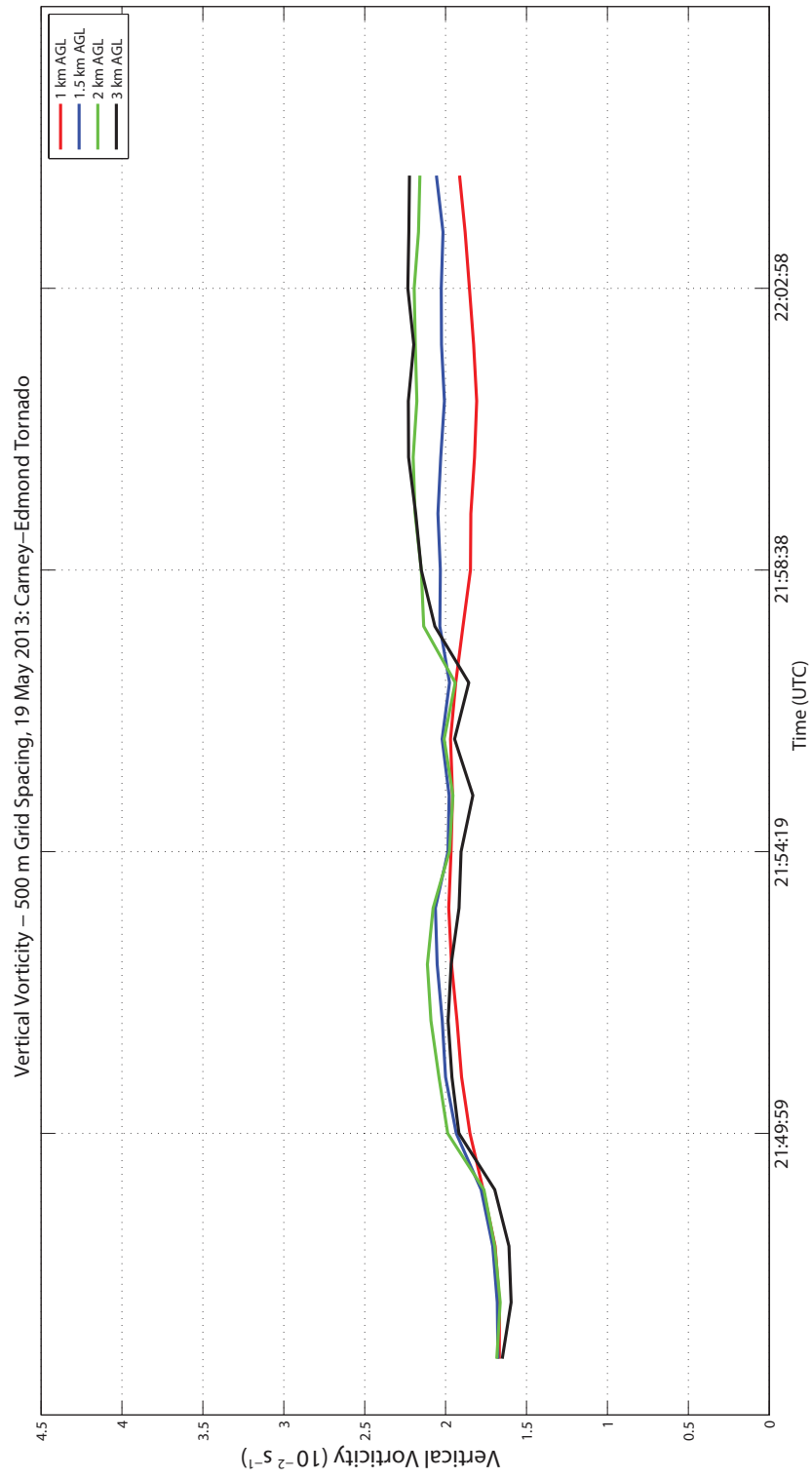


Figure 4.10: Vertical vorticity timeseries for the 500 m resolution analyses. Vertical vorticity ($10^{-2} s^{-1}$) is shown on the Y-axis.

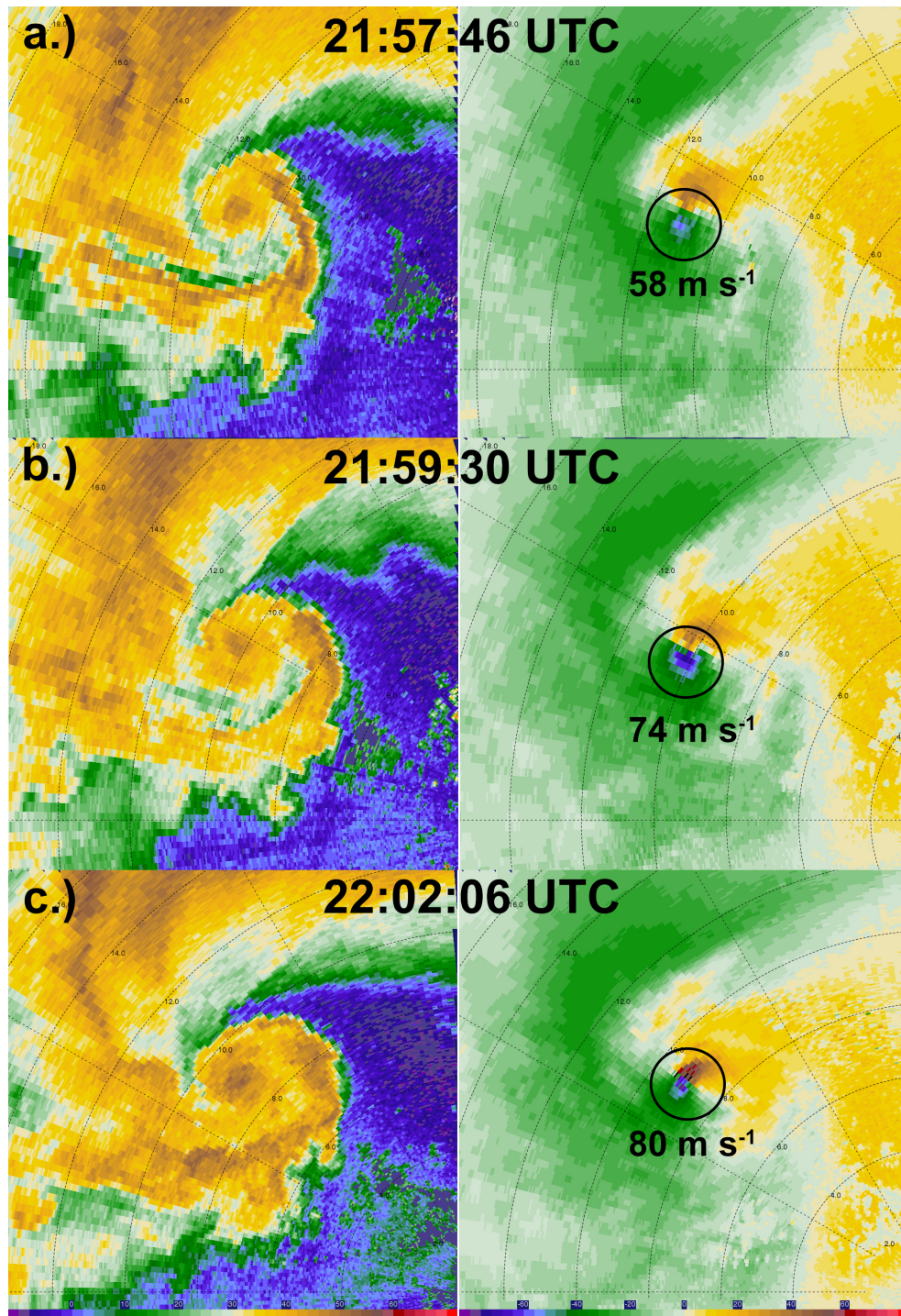


Figure 4.11: Raw RaXPol data illustrating the intensification of the tornado. Data are valid at a.) 21:57:46 UTC, b.) 21:59:30 UTC, and c.) 22:02:06 UTC. Radar reflectivity is in the left column, and Doppler velocity is in the right column.

and minimizing near 4 km. Additionally, as the tornado vortex intensified in both the raw radar data and the vorticity timeseries plots, the maximum ζ aloft appears to descend towards the surface while the minimum aloft stays relatively consistent. Although the maximum appears to be descending, it is more likely that the intensification is occurring near the surface first shifting the profile, and not that parcels with higher vorticity are descending. This result is consistent with previous studies, such as Wicker and Wilhelmson (1995), where ζ is maximized aloft for a time. However, these results differ from those of close-range, high-resolution, single-Doppler observations of tornadoes, which suggest the velocity (and therefore vorticity) should increase towards the surface (e.g., Alexander and Wurman, 2005; Bluestein et al., 2007; French et al., 2014; Houser et al., 2015). Bluestein et al. (2007) found that wind speeds in the surface friction layer increased by nearly 30% inside the core radius of the tornado, and by nearly 180% outside the core radius. Our results are again related to the 250 m grid spacing, which is too coarse to resolve the circulation nearest the ground. Additionally, we have no data below 660 m that could potentially supplement our analysis. Similar results are observed in the Norman-Shawnee dual-Doppler analysis; however, there are supplemental near-surface data which could potentially explain the difference in vorticity near the surface. These observations will be discussed in section 4.3.

A timeseries of maximum vertical velocity (Fig. 4.13) at 250 m grid resolution (red), and 500 m grid spacing (blue) depicts very unsteady vertical velocities at both resolutions. When the analysis begins, the vertical velocities in both case are relatively steady with 250 m resolution showing updraft speeds near 40 ms^{-1} . However, the data get particularly noisy after 21:59 UTC, when we see a spike in vertical velocity to almost 60 ms^{-1} . Although extremely rapid intensification of vertical velocities are possible due in updraft pulses, for example, it is highly unlikely that that is what is seen here. Moreover, what is likely observed is an

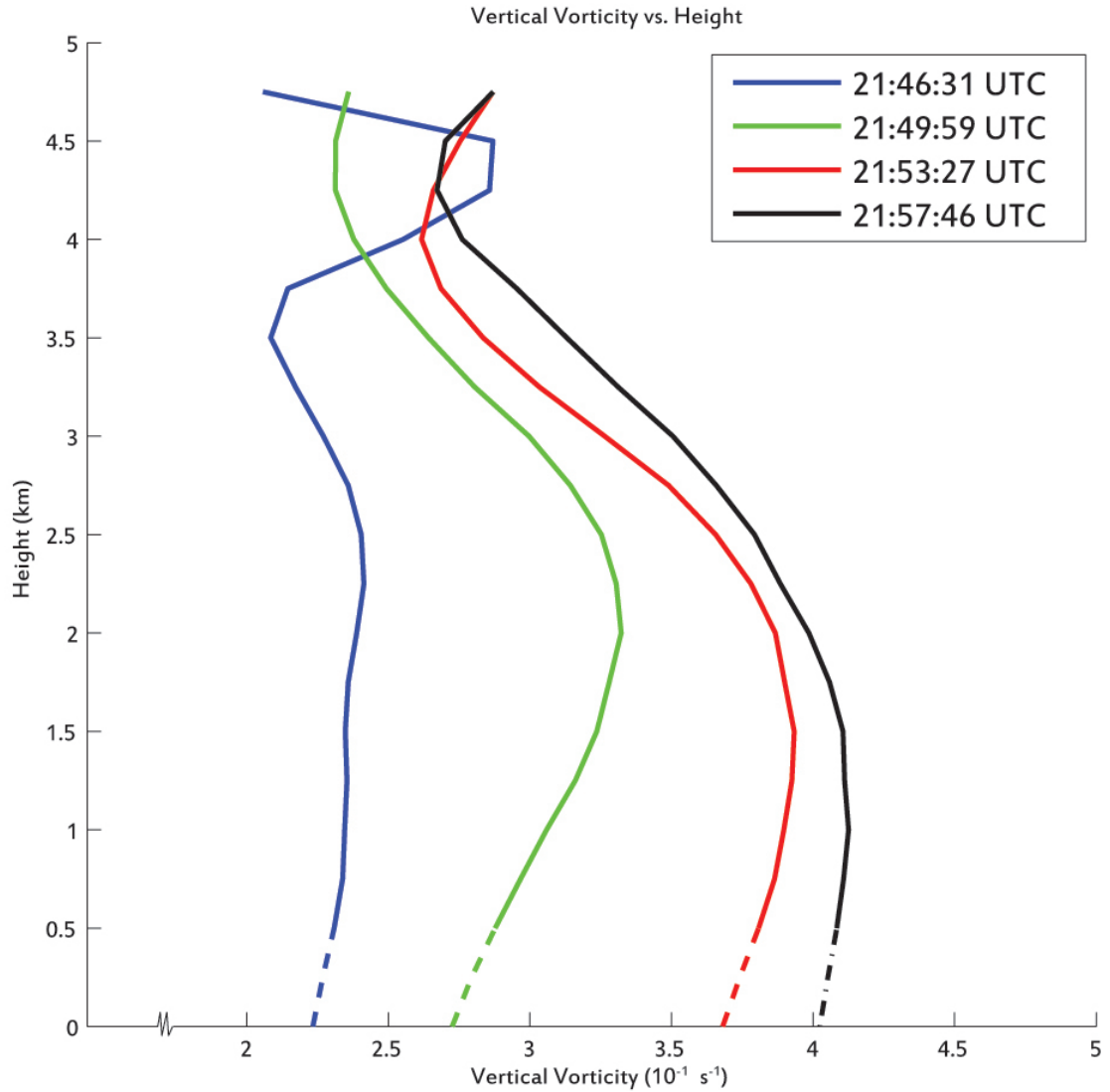


Figure 4.12: Maximum vertical vorticity (10^{-2} s^{-1}) plotted with height (km) for the times shown in figures 4.5-4.8. Dashed lines below 0.5 km represent data that is extrapolated and should be considered only as an estimate.

intensification in the maximum updraft speed as a result of lack of data as the storm began to move farther out of the dual lobe and the between beam angle decreased.

It is hypothesized that the reason for the sporadic nature of the maximum updraft intensity could be due to two things: the use of interpolation not allowing for changes in the evolution of the storm, or the movement of the storm towards the edge of the dual-Doppler lobe. Black hash marks represent actual volume collection start times, with the KTLX data in between interpolated using SVAC. Because RaXPol is collecting real-time data and KTLX data are interpolated, RaXPol will capture the evolution and non-linear changes in the storms during each update while the interpolated volumes will not. Given that the data are mostly steady and quantitatively reasonable throughout the first 8 minutes of the analysis, it seems unlikely that the result is due to interpolation. Additionally, the fluctuations are quite rapid and completely independent of the times in which “true” data are used from both KTLX and RaXPol. Toward the end of the analysis, the storm began to move closer and closer to the edge of the dual-Doppler lobe, resulting in data that likely falls outside of the usable data domain (where there is no radar data at all) and can cause inaccuracies. Additionally, the between beam angle begins to decrease with time. Even though the minimum beam angle throughout this analysis is never $< 30^\circ$, as the angle approaches 30° late in the analysis period, it is possible that this may be associated with the decline in data quality. Other studies have, too, noted the lack of reliable quantitative results of the w field in dual-Doppler analyses, especially when the between-beam angle is less than 30° (e.g., Tanamachi et al., 2012). Markowski et al. (2012a) comments that errors in data accumulate with height and should be only considered *qualitatively*. As is the case here, while the values are reasonable, magnitudes of w are prone to large errors and should not be trusted.

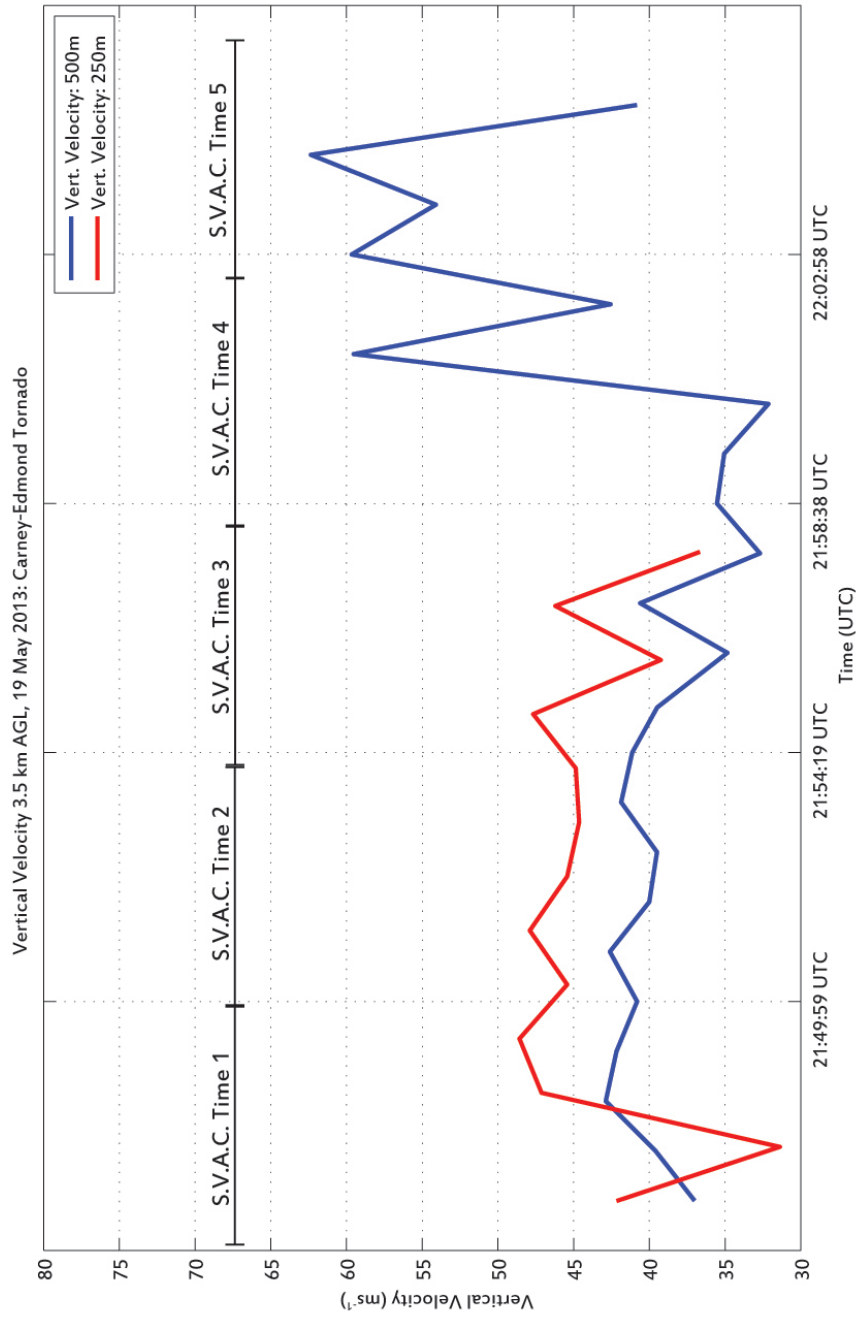


Figure 4.13: Updraft maximum vertical velocity timeseries for the 250m (red) and 500m (blue) resolution analyses. Vertical velocity is in units of m s^{-1} . SVAC time intervals are indicated with hash marks representing times at which true volume scans began, for reference.

4.3 The Norman-Shawnee Tornado

The second major tornado that occurred on 19 May 2013 began approximately 80 minutes after the first and occurred a little more than 50 km to the south, near Norman, OK. The third and southernmost supercell to fire off of the dryline that afternoon became tornadic at approximately 23:00 UTC, and the tornado persisted through 23:50 UTC. Approximately 15 minutes after tornadogenesis, the storm was already producing a strong tornado (Fig. 4.14). Low cross-correlation coefficient and Z_{DR} (Figs. 4.14c and d, respectively), coincident with a strong vortex couplet (Fig. 4.14b) all characterize a tornado debris signature. Maximum outbound wind speeds in excess of 50 m s^{-1} were observed by KTLX with a ΔV_{max} of 94 m s^{-1} at 1.8° elevation. In this case, a maximum wind speed of 58 m s^{-1} was found at 8.0° .

RaXPol began collecting data just after 23:02 UTC, about 2 minutes after tornadogenesis, and collected throughout the life of the tornado in three separate deployments. Deployment 4, which occurred less than 5 km from the violent tornado, was too close to perform dual-Doppler on, but single-Doppler data will be examined. Unlike the Carney-Edmond data set, this data set faced a number of challenges owing to the proximity of the storm to each radar. The location of the supercell at 23:02:22 UTC with respect to the dual-Doppler lobes and baseline is contoured in blue (Fig. 3.2). As the tornado propagated northeastward, the cross-beam angle briefly increased resulting in higher quality data for a period of time during the middle of deployment 3. However, as the region of interest began to approach the baseline, anomalous features began to appear which are not temporally continuous with time.

When the analysis began at 23:02:22 UTC, a tornado was ongoing within the dual-Doppler domain (not shown). A number of features similar to the previous case are observed, including a confluence zone along the forward flank of the storm,

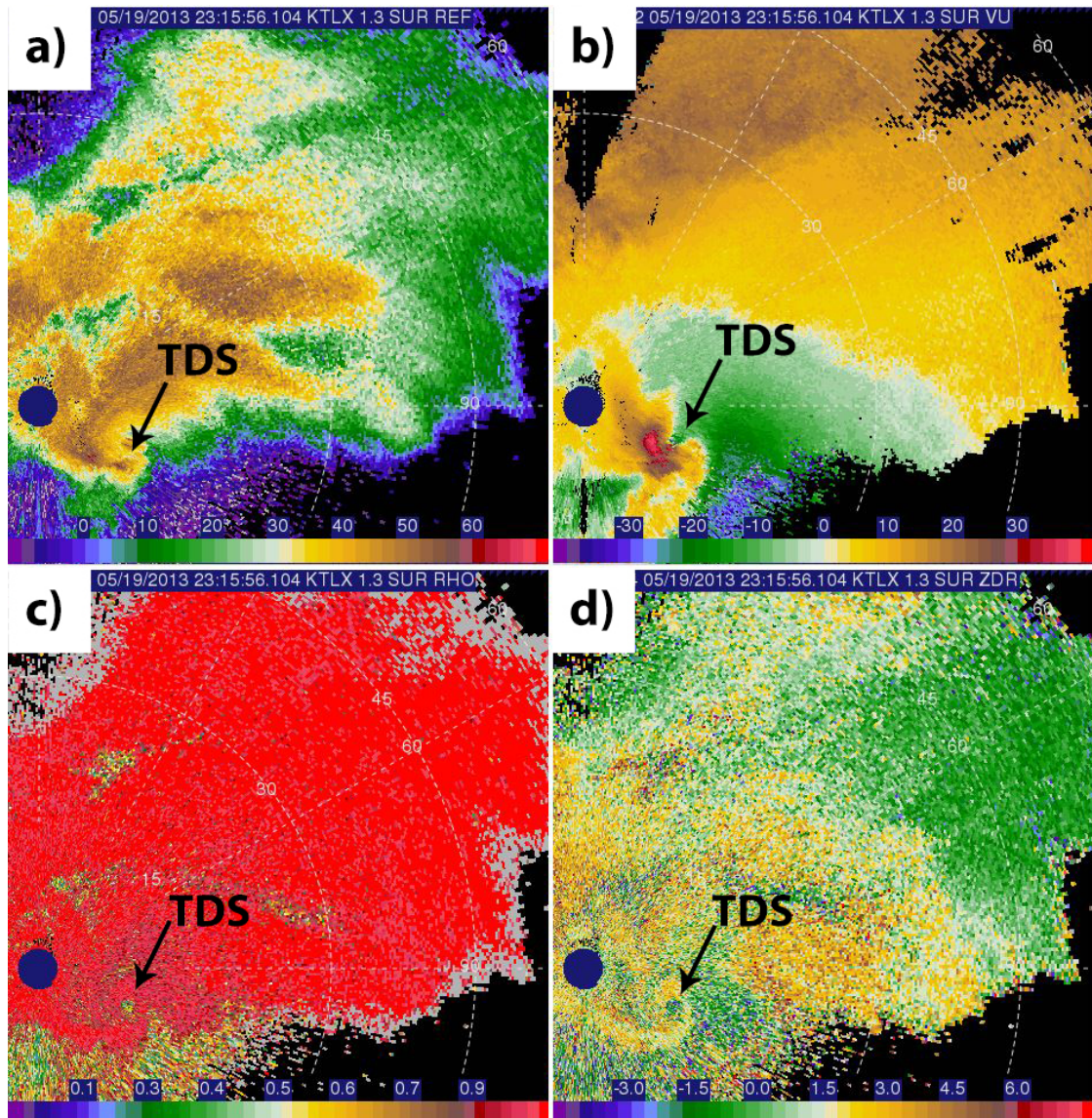


Figure 4.14: KTLX depiction (1.3° elevation) of the Norman-Shawnee supercell at 23:15:56 UTC, approximately 15 minutes after tornadogenesis. Radar variables shown are (a) radar reflectivity, (b) radial velocity, (c) cross-correlation coefficient, and (d) differential reflectivity. The location of the tornado debris signature (TDS) is marked, for reference.

and strong rotation on the north side of the hook echo. By 23:03:06 UTC (Fig. 4.15), ζ in excess of $3 \times 10^{-2} s^{-1}$ was observed at 750 m, located just north of the hook echo. Examination of the ζ maximum with height reveals a similar northward tilt as seen in the Carney-Edmond tornado. As the tornado continued to intensify both near the ground and aloft, the analyses reflect this intensification in maximum ζ . At 23:15:38 UTC (Fig. 4.15), vertical vorticity strengthened further to $> 4 \times 10^{-2} s^{-1}$ as the tornado and associated mesocyclone continued to grow in intensity. Reflectivity shows the hook echo begin to wrap northward into the forward flank. The vertical vorticity maximum is just slightly offset from this “debris ball,” however, the debris ball is located just south of the ζ maximum in the region of most intense winds.

As the tornado approached RaXPol, deployment 3 was cut short to relocate and redeploy the radar. Deployment 4 was a 4-minute deployment that collected very high resolution data less than 4 km from the strong tornado. Unfortunately, due to the orientation of the baseline, deployment 4 is not included in the dual-Doppler analysis. Deployment 5 began at approximately 23:43 UTC and spanned roughly 15 minutes after the Shawnee tornado had dissipated. At 23:43:02 UTC (Fig. 4.16), the tornado was still ongoing and the storm began to show evidence of cyclic tornadogenesis as the original ζ maximum began to propagate rearward with respect to the storm. The vorticity ζ in excess of $4 \times 10^{-2} s^{-1}$ persisted through this time as the tornado continued to do substantial damage despite a slight weakening and its rearward storm-relative motion. As the primary circulation that was associated with the tornado began to weaken and move rearward, a second area of weak-to-moderate ζ began to intensify coincident with the original hook echo. ζ approached $2.5 \times 10^{-2} s^{-1}$ in the second, eastern-most circulation as it briefly intensified. By 23:49:32 UTC, the first circulation had weakened substantially.

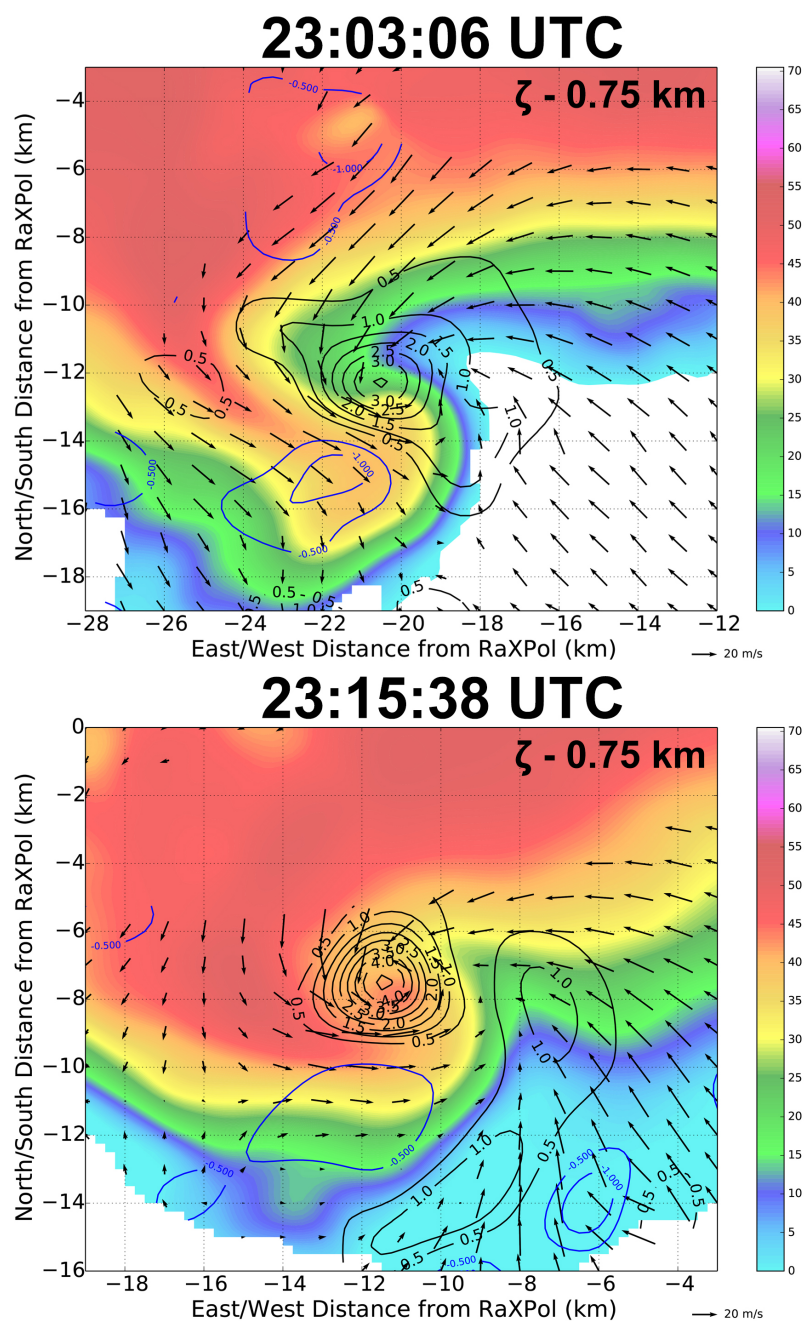


Figure 4.15: Dual-Doppler analyses of vertical vorticity at 750 m AGL valid at 23:03:06 UTC (Top) and 23:15:38 UTC (bottom). Positive vertical vorticity plotted every 0.5 s^{-1} in black, and negative vertical vorticity plotted every -0.5 s^{-1} in blue. KTLX reflectivity is color contoured.

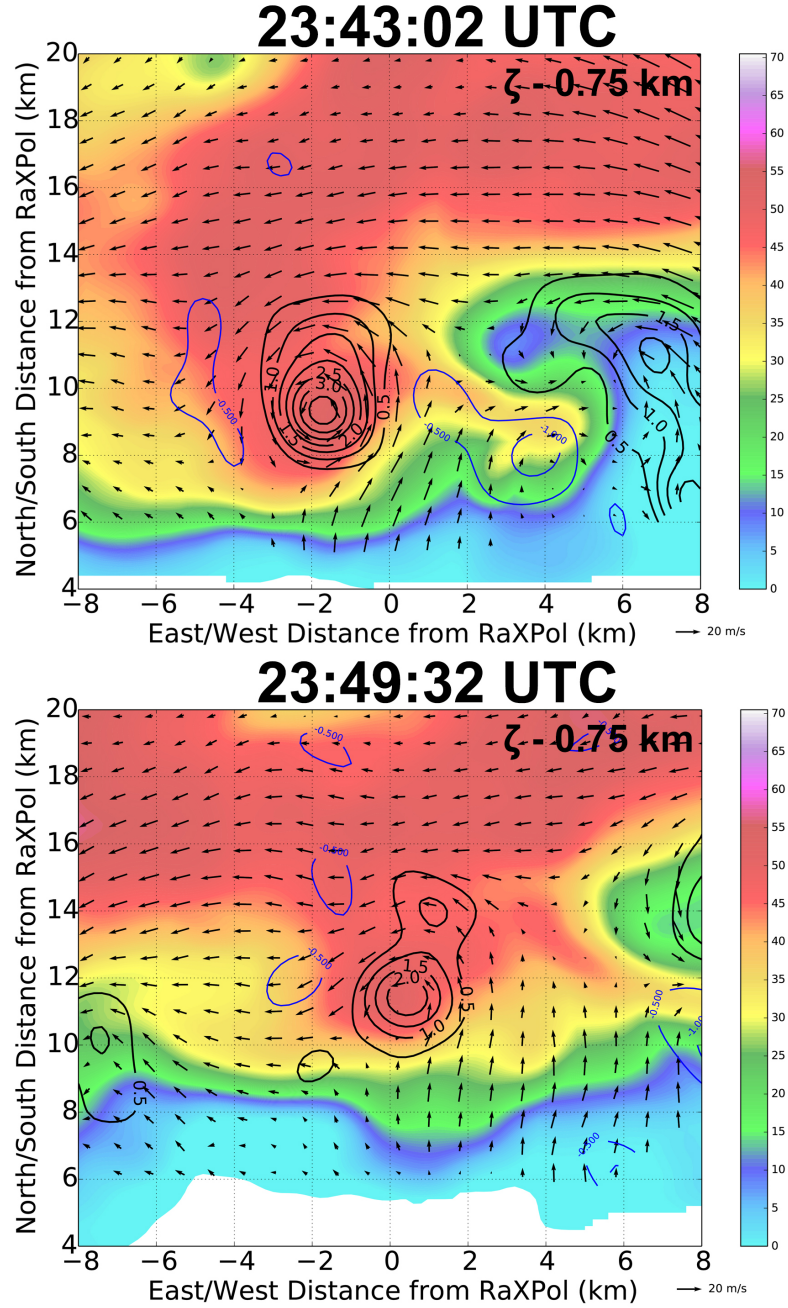


Figure 4.16: Dual-Doppler analyses of vertical vorticity at 750 m AGL valid at 23:43:02 UTC (Top) and 23:49:32 UTC (bottom). Positive vertical vorticity plotted every 0.5 s^{-1} in black, and negative vertical vorticity plotted every -0.5 s^{-1} in blue. KTLX reflectivity is color contoured.

The National Weather Service damage survey team estimated (<http://www.srh.noaa.gov/oun/?n=events-20130519>) that the Shawnee tornado dissipated at 23:50 UTC, seconds after the analysis in Fig. 4.16. As the original circulation continued to move rearward with respect to the storm and weakened, the second circulation seen first in Fig. 4.16 continued to persist; however, no substantial change (weakening nor strengthening) in vertical vorticity is observed. Three separate circulations are noted by 23:51:46 UTC, denoted with a “1”, “2”, and “3”, for reference (Fig. 4.17). Relative ζ maximum 1 is the remnant circulation that was associated with the Norman-Shawnee tornado. Circulation “2”, which was likely associated with cyclic mesocyclogenesis, was much stronger approximately 10 minutes earlier but had weakened substantially by 23:51:46 UTC. Circulation “3” formed to the west of the first two in the cusp of a southward surge in reflectivity, but was rather weak and short-lived. By 23:55:33 UTC, all three circulations were essentially non-existent as the entire storm weakened (Fig. 4.17).

As presented in the Carney-Edmond case, vertical ζ profiles obtained from dual-Doppler syntheses show the evolution of the profiles with height at selected times (Fig. 4.18). Note that since the beam height throughout D3 fails to reach below 150 m, the lowest grid point (i.e., the surface) has been extrapolated downward and therefore must be viewed with caution. Throughout D3, the vertical profiles of vorticity exhibited a characteristic “S” shape with a local minimum near the surface, a maximum aloft between 1.5 and 3 km AGL, and another local minimum between 3.5 and 4.5 km AGL. Maxima range from $2.5 \times 10^{-2} \text{ s}^{-1}$ to $5.9 \times 10^{-2} \text{ s}^{-1}$; the lowest ζ are at 0.25 km AGL, and tend to strengthen with height, as the peak vorticities range from $3.5 \times 10^{-2} \text{ s}^{-1}$ and $5.5 \times 10^{-2} \text{ s}^{-1}$ between roughly 1.25 to 3 km AGL. The vortex generally strengthens both near the surface and aloft with time and the relative maximum seen at each time appears to decrease in height. That is, as the vorticity at low levels strengthened, the height AGL of the local

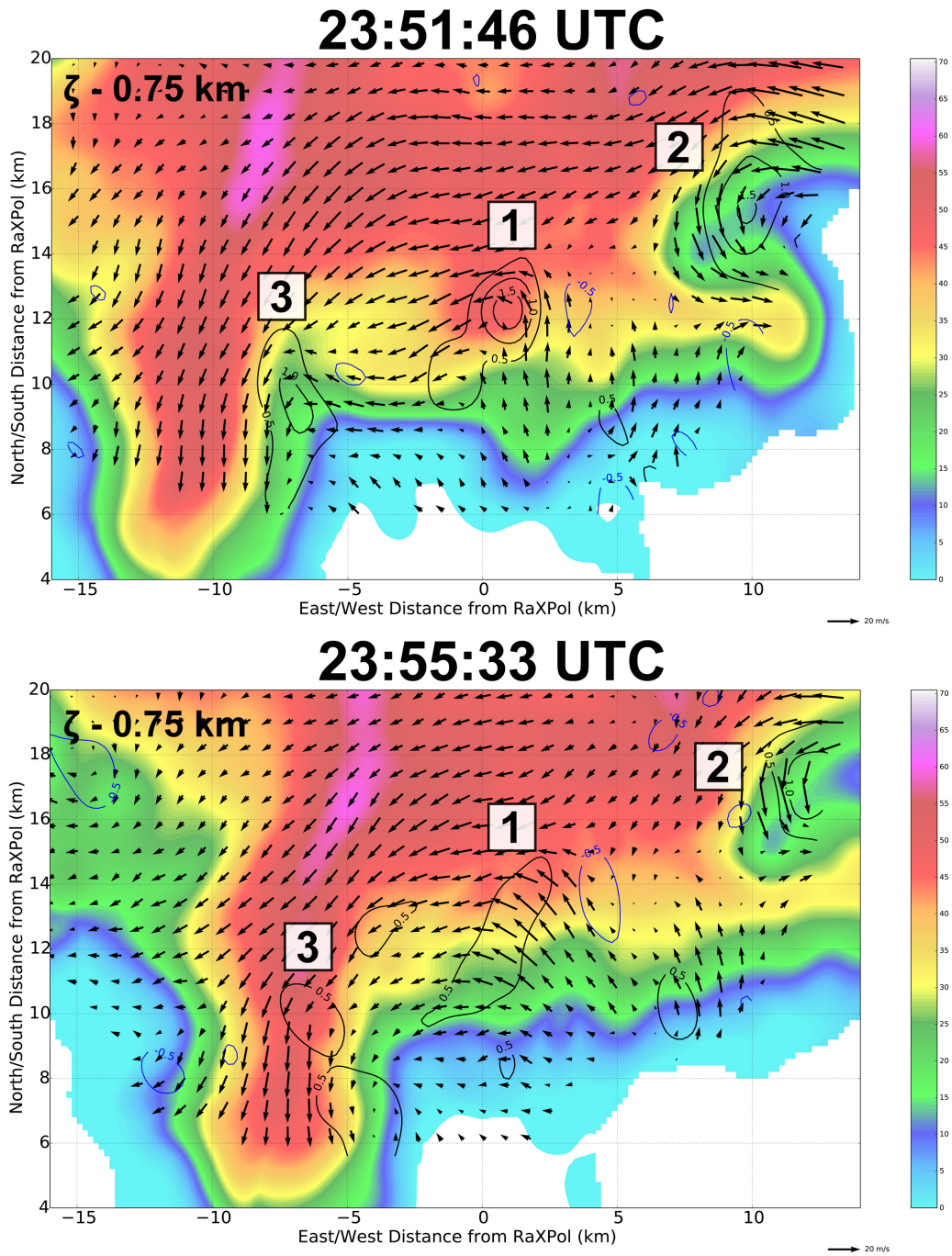


Figure 4.17: Dual-Doppler analyses of vertical vorticity at 750 m AGL valid at 23:51:46 UTC (Top) and 23:55:33 UTC (bottom). Positive vertical vorticity plotted every 0.5 s^{-1} in black, and negative vertical vorticity plotted every -0.5 s^{-1} in blue. KTLX reflectivity is color contoured.

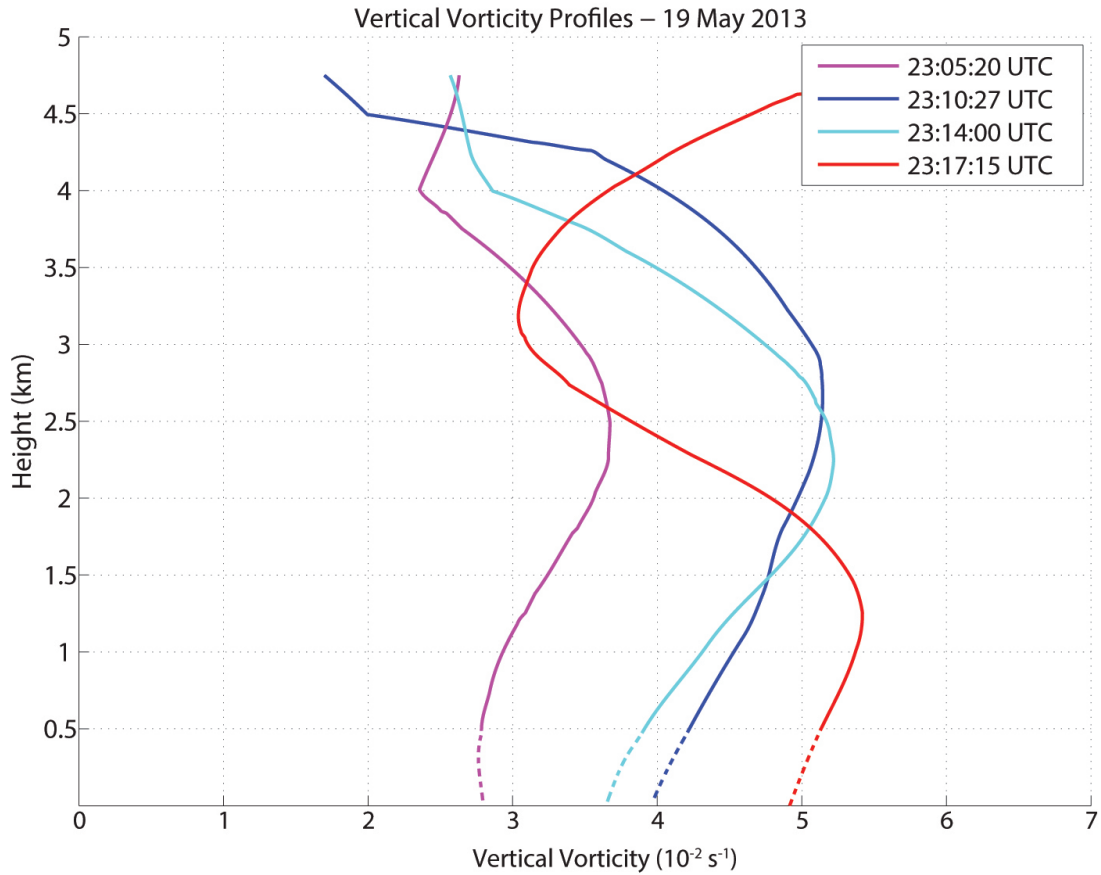


Figure 4.18: Vertical profiles of vertical vorticity obtained from the dual-Doppler analysis during D3. Profiles are taken through the vorticity maximum at each level for the corresponding analysis time.

maximum decreases. With respect to vertical vorticity above 3 km, no definitive trend is evident from this dataset.

At first, the vertical vorticity at 0.75 km AGL (Fig. 4.19) exhibits little change, but rapidly intensified after 23:14:00. By and large, the general trend at low-levels is a gradual strengthening within the first 13 minutes of data collection (two gaps of missing data exist due to a change in scanning strategy mid deployment). By 23:14:00, the vortex begins to rapidly intensify at low levels, but the opposite change is observed at mid-levels. As the vorticity increases to in excess of $4.0 \times$

$10^{-2} s^{-1}$ at both 0.75 and 1 km AGL, the 2 and 3 km vorticity decreased rapidly. The mid-level vorticity minimizes just prior to 23:20 at the same time in which the surface vorticity peaks.

Vertical ζ profiles exhibited a characteristic “S” shape, or in profiles of ΔV_{max} (the maximum change in wind speed across the vortex, which can be related to vertical vorticity through consideration of the core radius), which corresponds well with previous studies above the surface (e.g., Tanamachi et al., 2012; French et al., 2013). One major difference between this case and previous observations and numerical modeling studies is the absence of an increase in vorticity as we near the surface. In this case, the reason for the lack of an observed intensification of vertical vorticity is likely a result of missing data near the surface. First, because these observations occur at a distance (both radars never simultaneously observing below 150 m AGL), we fail to directly observe the inertial sublayer, where frictional convergence spins up preexisting vertical vorticity. The second reason is that our spatial resolution is such that we are unable to resolve features smaller than 500 m owing to the use of a 250 m grid spacing. The vorticity sampled is therefore only valid on the mesocyclone scale, not the tornado scale. Wicker and Wilhelmson (1995) found that vertical vorticity was maximized aloft in the mesocyclone due to tilting and subsequent stretching acting on horizontal vorticity that had been generated baroclinically. There is a “sweet spot” aloft where the combined effects of tilting and stretching of this vorticity are the greatest.

At the end of D3, RaXPol moved south to avoid being hit by the tornado. D4 provides volume scans at a range of less than 4 km to the center of the tornadic circulation. Although dual-Doppler is not available in this case because the radar was too close to the tornado, data were obtained with minimum beam heights between 40 and 66 m. According to Lewellen et al. (2000), the height of the inertial layer can be approximated as 0.5 times the core radius of the tornado

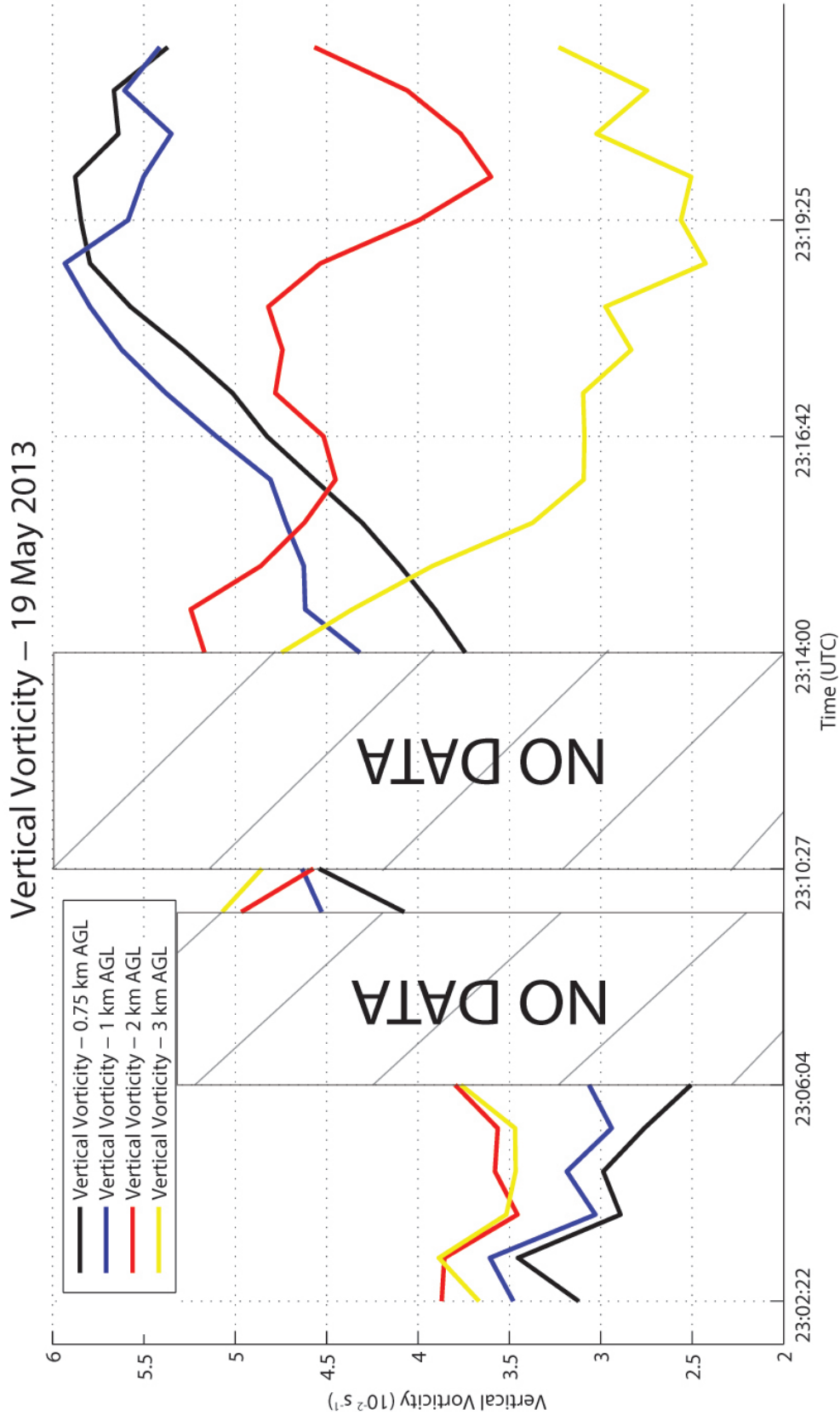


Figure 4.19: Time series of maximum vertical vorticity (s^{-1}) at 750 m (black), 1 km (blue), 2 km (red), and 3 km (yellow) obtained from dual-Doppler analysis. Times in which the data was unavailable is shown in the figure, for reference.

at the ground. In this case, the core radius is estimated to be 180 to 200 m, resulting in an inertial layer nearly 90 m deep. One important difference between D4 and D3 is that because dual-Doppler coverage is not available in D4, the vertical vorticity is determined assuming that the vortex is axisymmetric, and is calculated as $2\Delta V/\Delta X$. Additionally, because only RaXPol determines the resolution in this case, a horizontal resolution of 50 m was used, which is capable of resolving the tornadic circulation.

Vertical profiles of axisymmetric vertical vorticity (AVV) were calculated from D4 (Fig. 4.20) data, and are plotted similarly to ζ profiles from D3 (Fig. 4.18) to help provide insight on the near-surface effects of vertical vorticity which are missing from the dual-Doppler analysis. The characteristic “S” shape is not as evident in the AVV plots; however, this is not surprising given the increased resolution and a maximum data height of 1.2 km AGL. One major difference is the noticeable change in vorticity below 1 km. In Fig. 4.18, vorticity largely increases with height, but here we see evidence that vorticity actually decreases with height below 1 km. In this case, boundary layer effects on vertical vorticity are likely being resolved. Vorticity at the surface varies only slightly, ranging from 0.45 s^{-1} to over 0.5 s^{-1} over the short deployment period. The lack of any appreciable variation in vorticity at this time could be due to the short time period during which RaXPol was deployed, or that the tornado had reached peak intensity and was relatively steady (the tornado was producing EF-4 damage at this time; Fig. 3.2). At all times during D4, vorticity decreases nearly linearly with height, stays nearly constant with time, and reaches a minimum at the top of the domain.

4.3.1 Discussion of vorticity profiles

There are a number of theories that describe mechanisms that may cause vortex intensification near the surface. One theory, the dynamic pipe effect (Leslie,

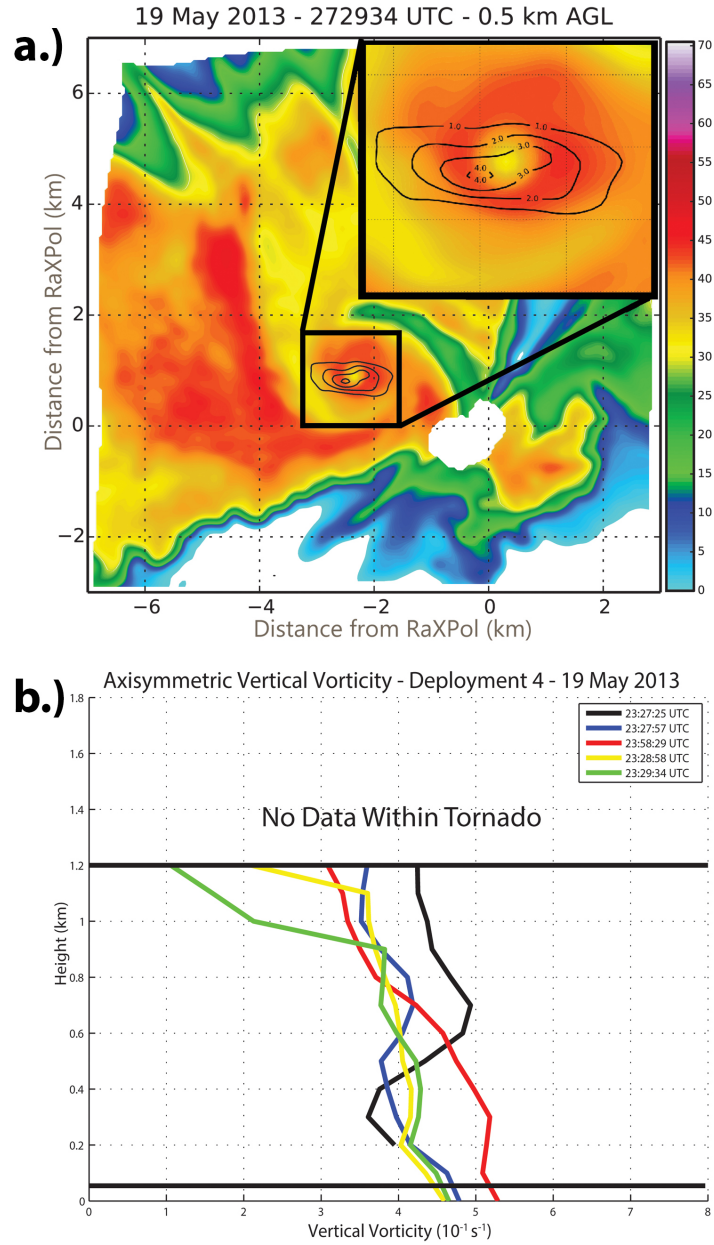


Figure 4.20: Objectively-analyzed radar reflectivity from RaXPOL deployment 4 at 27:29:34 UTC are shown in (a). Vertical vorticity (black contours; $10^{-1} s^{-1}$) is illustrated in a zoomed panel located at the top right of figure (a), and is valid at $z = 0.5$ km AGL. Figure (b) presents vertical profiles of axisymmetric vertical vorticity plotted in $10^{-1} s^{-1}$. The solid black line at 66 m AGL represents the minimum beam height of RaXPOL, below which values have been estimated by extrapolation.

1971), states that vortex intensification first occurs aloft and builds downward to the surface. A second theory suggests that with the aid of friction, near-surface convergence acting on weak, preexisting near-surface vorticity can intensify it to tornado strength (Ward, 1972). Because we are interested in boundary layer effects on the tornado, we will focus only on the second of the two theories. Within the tornado boundary layer, flow that is characterized by swirling motions perpendicular to a central axis is slowed by friction near the surface. Described in Bluestein (2013) and Rotunno (2013), the swirling motions of the vortex drive an inward directed pressure-gradient force. As the flow slows owing to friction, the centrifugal force decreases disrupting cyclostrophic balance. Because this has no effect on the inward directed pressure gradient force, the flow is accelerated inward toward the center of the vortex. Directly above the frictional layer, both a radial and an azimuthal component of velocity exist. As the flow moves radially inward, it encounters the “corner” region in which radial inflow is turned vertically, forming a strong vertical jet. Within the corner flow region, the friction layer decreases in depth owing to the vertical turning of the flow (i.e., horizontal flow is no longer subject to surface friction), and tornado wind speeds have been shown to be the most violent very near the surface Bluestein et al. (2007). As the flow converges and turns vertically, preexisting vertical vorticity intensifies owing to convergence and stretching. If vorticity is enhanced by a large enough amount, tornadogenesis may occur. Examinations of results from D4 indicate an increase in vertical vorticity within the inertial layer. As described above, this is likely due to convergence in the friction layer very near the surface (below the minimum beam height even in this case). With time, vorticity that is concentrated near the surface is enhanced owing to convergence. As the flow reaches the corner, vertical acceleration of the air further stretches the vorticity and intensifies the vortex. In a strengthening vortex, the radial flow increases, becomes more convergent, and continues to enhance

vorticity with time. Near the ground, where the frictional effects are the strongest, we see a maximum in vertical vorticity which slowly decreases with height. This vorticity decreases throughout roughly the lowest 1 km AGL; however, the general trend is for vorticity to increase as it approaches the surface. This is also consistent with results shown in Wicker and Wilhelmson (1995), in which a relative maximum exists near the surface and above 1 km. While both a maximum aloft **and** intensification near the surface was not observed in either D3 or D4, the combined results reflect these findings. While there are no data from D3 near the surface, pertinent data was collected aloft at this time. The maximum aloft is likely on the mesocyclone scale and is a result of horizontal vorticity generated baroclinically, tilted, and then advected upwards. The region above approximately 1 km provides a “sweet spot” where this vorticity is maximized. Once we are able to resolve the tornado-scale circulation and collect data within the inertial layer, we are able to see the effects of convergence on the vortex as it intensifies near the surface.

By 23:14:00 UTC, the vertical vorticity at all levels rapidly intensified; however, the low-level and mid-level rotation appear inversely associated with one another. That is, as the low-level vorticity intensified, the vorticity at 2 and 3 km generally weakened. One explanation for this is that as the core radius grew with height as the tornado intensified with time, the vertical vorticity dropped. Because vorticity is a function of ΔV_{max} across the tornado and the radius of maximum wind speed, one feasible explanation would be that the tornado broadened with height with the maximum wind speeds at that height held relatively constant. Vertical profiles of w (Fig. 4.21) may help to explain the reason for this inverse relationship between vertical vorticity near the surface and at mid-levels. If we consider the change in vertical vorticity with time (while neglecting frictional generation and baroclinic generation of ζ):

$$\frac{\partial \zeta}{\partial t} = -\vec{v} \cdot \nabla \zeta + \omega \cdot \nabla w, \quad (4.1)$$

where $\omega = (\xi, \eta, \zeta)$ is the 3-D relative vorticity vector. Breaking down the second term on the RHS in the vertical vorticity equation, we find that

$$\omega \cdot \nabla w = \left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z} \right) \frac{\partial w}{\partial x} + \left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \right) \frac{\partial w}{\partial y} + \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) \frac{\partial w}{\partial z}, \quad (4.2)$$

where the first term on the RHS of the equation represents tilting of x-oriented vorticity, the second term represents the tilting of y-oriented vorticity, and the third term represents the stretching of vertically-oriented vorticity. With respect to Fig. 4.21, stretching of vertical vorticity (or in this case shrinking) likely plays a role in the decrease of vertical vorticity at the mid-levels while vorticity in the low levels continually intensifies. The vertical velocity profile at 23:14:33 UTC exhibits a change in the sign of w at 3 km AGL. The result would be the continued intensification of vertical vorticity below 3 km, but a decrease in vertical vorticity at and above 3 km as the sign of $\frac{\partial w}{\partial z}$ flips. By 23:18:52 UTC, a kink in the vertical velocity profile descends to between 1.5 km and 2.5 km where $\frac{\partial w}{\partial z}$ approaches 0, and flips sign for a brief period of time before once again becoming positive. This likely explains the steady drop in vertical vorticity at 2 km, while the vertical vorticity at 3 km steadies and then increases after 23:19:25 UTC, and the vertical vorticity at 0.75 km and 1 km both generally strengthen with time before leveling off and then falling as the kink in the vertical velocity profile continued to descend.

Stretching of vertical vorticity by the updraft is one reasonable explanation for the inverse relationship in vertical vorticity at the low and mid levels. However, why are we observing this descending kink in vertical velocity with time as this tornado intensified? The explanation likely deals with the descent of the maxima observed in vertical vorticity profiles with height, which also descends with time, and has a significant contribution to the intensity of the updraft. As the vorticity

maximum descends, so does the kink in the vertical velocity, and the resulting scenario is one of the classic chicken and egg problem. The kink in the vertical velocity profile could have been associated with effects of the relative maximum in ζ , but the opposite is also true. Stretching of vorticity by accelerations in the updraft may be due to a change in sign of the updraft, but which occurs first? Although there is no way to definitively say from these data alone, vertical vorticity profiles at times presented in Fig. 4.18 exhibit very similar profiles to those seen in the Carney-Edmond tornado, a case in which we did not observe the same effects of vortex stretching on the vertical vorticity. In that case, steady intensification was observed to occur in relative unity at all levels throughout the analysis. As the tornado and associated mesocyclone intensified in both cases, the vorticity maximum descends with time, as well as shifts the entire profile to the right as ζ generally intensified throughout the entire column.

One explanation for this “descending vorticity maximum” is that the vorticity near the surface intensified with time, shifting the profile in the lowest levels to the right. The maximum in vertical vorticity would then appear to progress downward towards the surface. In essence, the descent of the relative maximum is not associated with the descent of parcels with higher relative vorticity, but rather the maximum approaches the surface as the source of vorticity intensification *is* the surface.

As the vertical vorticity maximum descended towards the surface, consideration of the dynamic perturbation pressure gradients can help us provide one possible explanation for these observations in the vertical vorticity timeseries. Consider the following relationship between the perturbation pressure and the three-dimensional wind and buoyancy field:

$$p' \propto \left[\left(\frac{\partial u'}{\partial x} \right)^2 + \left(\frac{\partial v'}{\partial y} \right)^2 + \left(\frac{\partial w'}{\partial z} \right)^2 \right] + 2 \left(\frac{\partial v'}{\partial x} \frac{\partial u'}{\partial y} + \frac{\partial w'}{\partial x} \frac{\partial u'}{\partial z} + \frac{\partial w'}{\partial y} \frac{\partial v'}{\partial z} \right) + 2\mathbf{S} \cdot \nabla_h w' - \frac{\partial B}{\partial z}, \quad (4.3)$$

where the first two terms on the RHS of the equation represent nonlinear dynamic pressure perturbations, the third term is the linear dynamic pressure perturbation, and the fourth term is the buoyancy pressure term. Through some careful manipulation, it can be shown that the above equation can be reduced to the following:

$$p' \propto \left[\left(\frac{\partial u'}{\partial x} \right)^2 + \left(\frac{\partial v'}{\partial y} \right)^2 + \left(\frac{\partial w'}{\partial z} \right)^2 \right] - \frac{1}{2} \zeta'^2 + 2\mathbf{S} \cdot \nabla_h w' - \frac{\partial B}{\partial z}, \quad (4.4)$$

where the second term, sometimes referred to as the spin term or nonlinear dynamic forcing term. For the sake of this analysis, let's consider effects of the spin term with height on the dynamic (no buoyant contribution in this case) perturbation pressure term. The following equation better illustrates the relationship between these two terms:

$$-\frac{\partial p'_d}{\partial z} \propto \frac{1}{2} \frac{\partial \zeta'^2}{\partial z}, \quad (4.5)$$

which relates the change of perturbation vertical vorticity with height (RHS) to the change of the dynamic perturbation pressure term with height. This relationship shows that as vertical vorticity increases with height, the perturbation pressure does the inverse. In this case, with the vertical vorticity maximum located aloft and descending, the sign of the RHS of 4.5 is positive below the maximum, becoming zero at the maximum and then becoming negative above. Therefore, below the vertical vorticity maximum, we will see the perturbation pressure decrease upwards below the maximum, or an upward directed perturbation pressure

gradient (related to the LHS of 4.5 as $-\alpha \frac{\partial p'_d}{\partial z}$) below the vorticity maximum, and a downward directed perturbation pressure gradient force above the descending vorticity maximum. This effect would likely have played a role in the kink observed in Fig. 4.21. Contributions to the updraft are broken down into dynamic and buoyant components by the following equation:

$$\frac{Dw}{Dt} = -\alpha \frac{\partial p'_d}{\partial z} - \alpha \frac{\partial p'_b}{\partial z} + B, \quad (4.6)$$

where the first term on the RHS of (4.6) represents the dynamic forcing, and the second and third terms on the RHS represent the buoyant forcing terms. As discussed above, the dynamic contribution to vertical velocity by the mesocyclone depends on the intensity of the vorticity maximum, or more specifically the change in vertical vorticity with height ($\frac{\partial \zeta'^2}{\partial z}$). Well above vertical vorticity maximum, the slope of ζ'^2 becomes small, resulting in a lesser contribution to vertical velocity. In these regions, buoyant contributions to $\frac{Dw}{Dt}$ dominate the dynamic contributions. Additionally, buoyancy can best be described by the following equation (in terms of potential temperature):

$$B \approx g \frac{\theta'}{\bar{\theta}}, \quad (4.7)$$

where θ' represents the perturbation potential temperature of the parcel from the environment, and $\bar{\theta}$ represents the mean environmental potential temperature. As an estimate, we can assess the buoyancy of the parcel based on the difference between the parcel line and the environmental temperature line. In this case, according to Fig. 4.1e, the parcel buoyancy begin to increase rapidly above the LFC (≈ 2 km) due to steep mid-level lapse rates, and is maximized near 7.5 km AGL. This further supports that buoyancy likely begins to dominate well above the maximum in vertical vorticity, providing further explanation as to why the velocity profile changes so rapidly only very near the maximum in vertical vorticity.

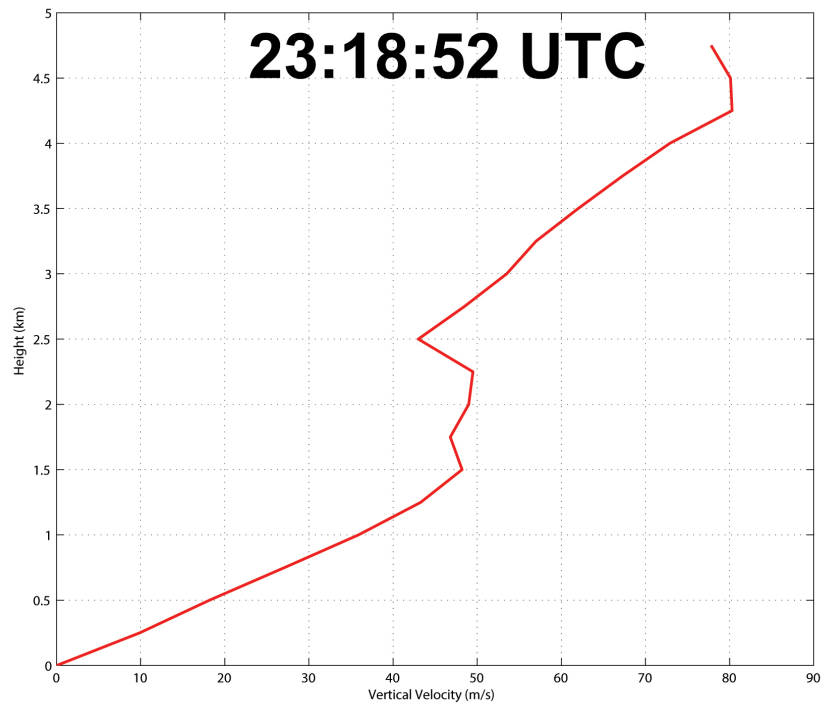
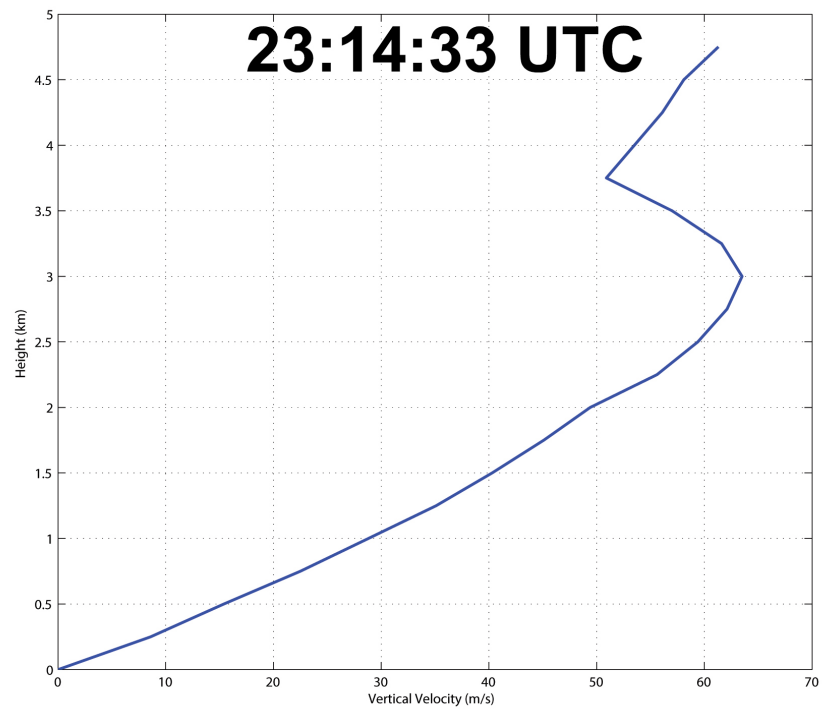


Figure 4.21: Selected profiles of vertical velocity ($m s^{-1}$) as a function of height (km) during D3.

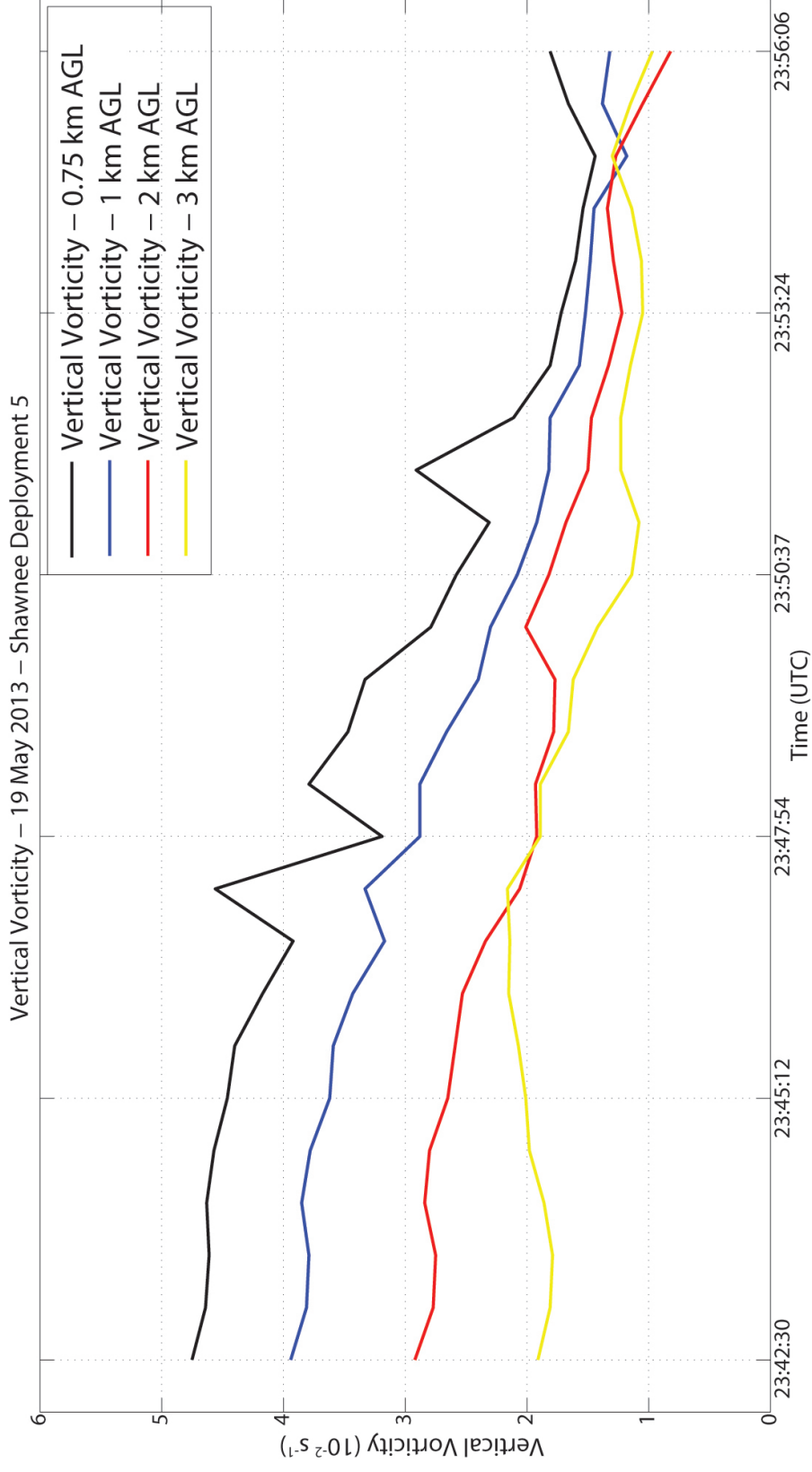


Figure 4.22: Time series of maximum vertical vorticity (s^{-1}) at 750 m (black), 1 km (blue), 2 km (red), and 3 km (yellow) obtained from dual-Doppler analysis as the tornado weakened. For reference, the tornado dissipated at approximately 23:50 UTC according to the NWS damage survey.

Below the vorticity maximum, vertical velocity is affected both by buoyancy and the upward directed perturbation pressure gradient force (PPGF) associated with the vorticity maximum. At and above the vorticity maximum, buoyancy continues to drive positive vertical velocities, but downward directed PPGF due to the vorticity maximum acts opposite the force of buoyancy. This effect decreases the overall vertical velocity, changing the sign of $\frac{dw}{dz}$ with height. Continuing vertically along the velocity profile, once the parcels are sufficiently far from the vorticity maximum, the downward directed PPGF is sufficiently small such that it no longer has a noticeable effect on the vertical velocity.

Due to the location of the storm within the dual-Doppler lobes during both D3 and D5, vertical velocity data in general were prone to substantial errors, especially early in D3. For this reason, we only consider the vertical velocities after 23:14:00 UTC to be qualitatively representative of the true storm environment. Complicating the issue further is the fact that the storm quickly approached the baseline late during D3; thus, the analyses lack numerous data points on and near the baseline. We can, however, get an idea of the updraft shape and intensity from the data that were usable in this region.

At 23:15:38 UTC, vertical velocities at 750 m resemble the familiar horseshoe-like shape to the updraft observed in the Carney-Edmond case and numerous previous studies (Fig. 4.23). Similar magnitudes near 10 m s^{-1} are observed in this case as well. At 3.5 km, vertical velocities in excess of 50 m s^{-1} are seen coincident with a weak echo region in reflectivity. Unlike in the Carney-Edmond tornado, the updraft is directly collocated with the bounded weak echo region. There are a number of possibilities for why this collocation exists in this storm, but not the Edmond-Carney storm. One potential explanation is that the updraft in the Carney-Edmond storm was tilted resulting in a slight displacement of the

bounded weak echo region from the maximum updraft, as discussed in the previous section. In this case, the supercell was much more intense during the analysis period, as indicated by both a stronger maximum vorticity and updraft intensity. Because the mesocyclone was much more intense, a greater contribution by the vertically oriented PPGF could have caused the updraft to be more erect than in the Carney-Edmond case, at least during the analysis period. In the Carney-Edmond storm, the displacement of the BWER from the maximum updraft decreased with time while the maximum updraft in this case stays closely located to the BWER throughout. The updraft, both at 750 m and 3.5 km, retains the same relative shape and intensity through 23:21:02 UTC (Figs. 4.24 and 4.25).

The relative consistency between w observations in the Carney-Edmond and Norman-Shawnee cases with previous studies give us confidence that this technique provides reasonably useful results. Further work is needed to understand the accuracy to which this technique captures the magnitudes of w , but the results from both cases are encouraging.

4.4 Analysis of Near-Storm Heterogeneity in the Norman-Shawnee Supercell

The primary purpose of assessing the inflow is to gain a better understanding of changes in the structure of the low-level vertical wind shear within a tornadic environment (similar to numerical simulations by Dahl et al. (2014)), more specifically the effects the tornado has on its own environment. In this section, the inflow is first examined in an Eulerian framework. Heterogeneity of the vertical wind structure near the tornado is then considered at multiple times, and these effects are quantified in terms of storm-relative helicity and compared in time as they interact with the tornado.

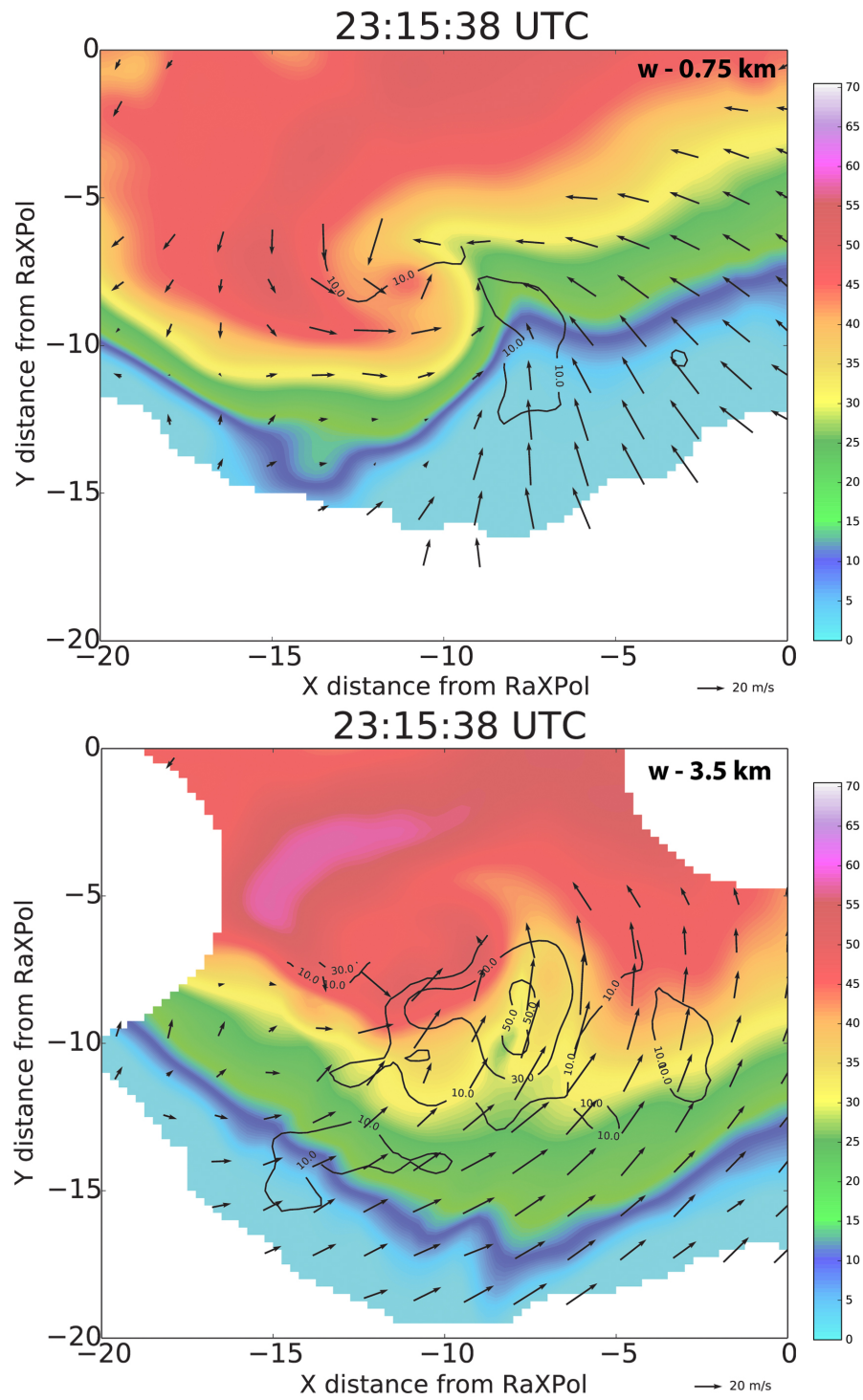


Figure 4.23: Dual-Doppler analyses of vertical velocity at 750 m AGL (top) and 3.5 km (bottom) valid at 23:15:38 UTC. Vertical velocity is plotted every 10 m s^{-1} in black. KTLX reflectivity is color contoured.

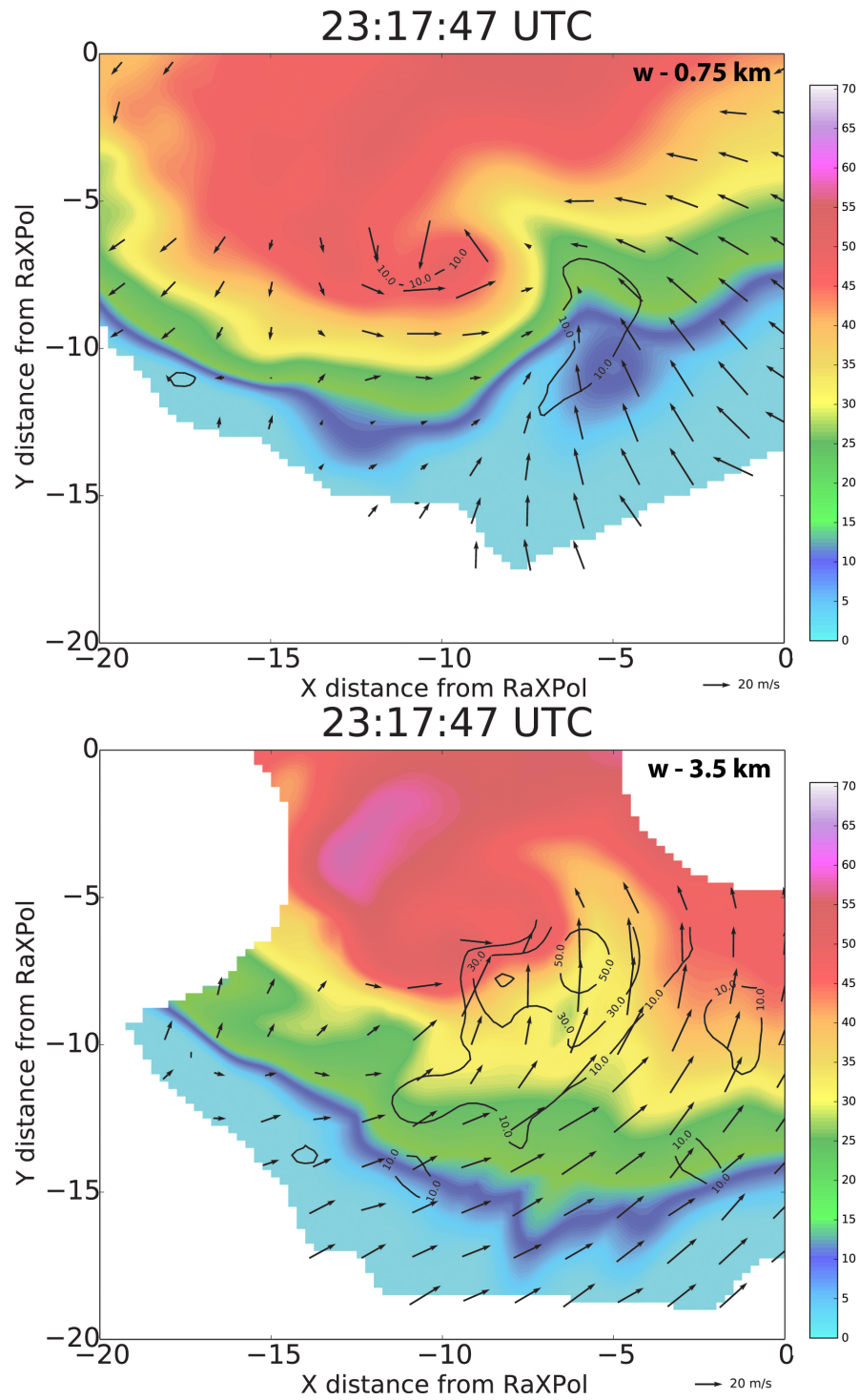


Figure 4.24: Dual-Doppler analyses of vertical velocity at 750 m AGL (top) and 3.5 km (bottom) valid at 23:17:47 UTC. Vertical velocity is plotted every 10 m s^{-1} in black. KTLX reflectivity is color contoured.

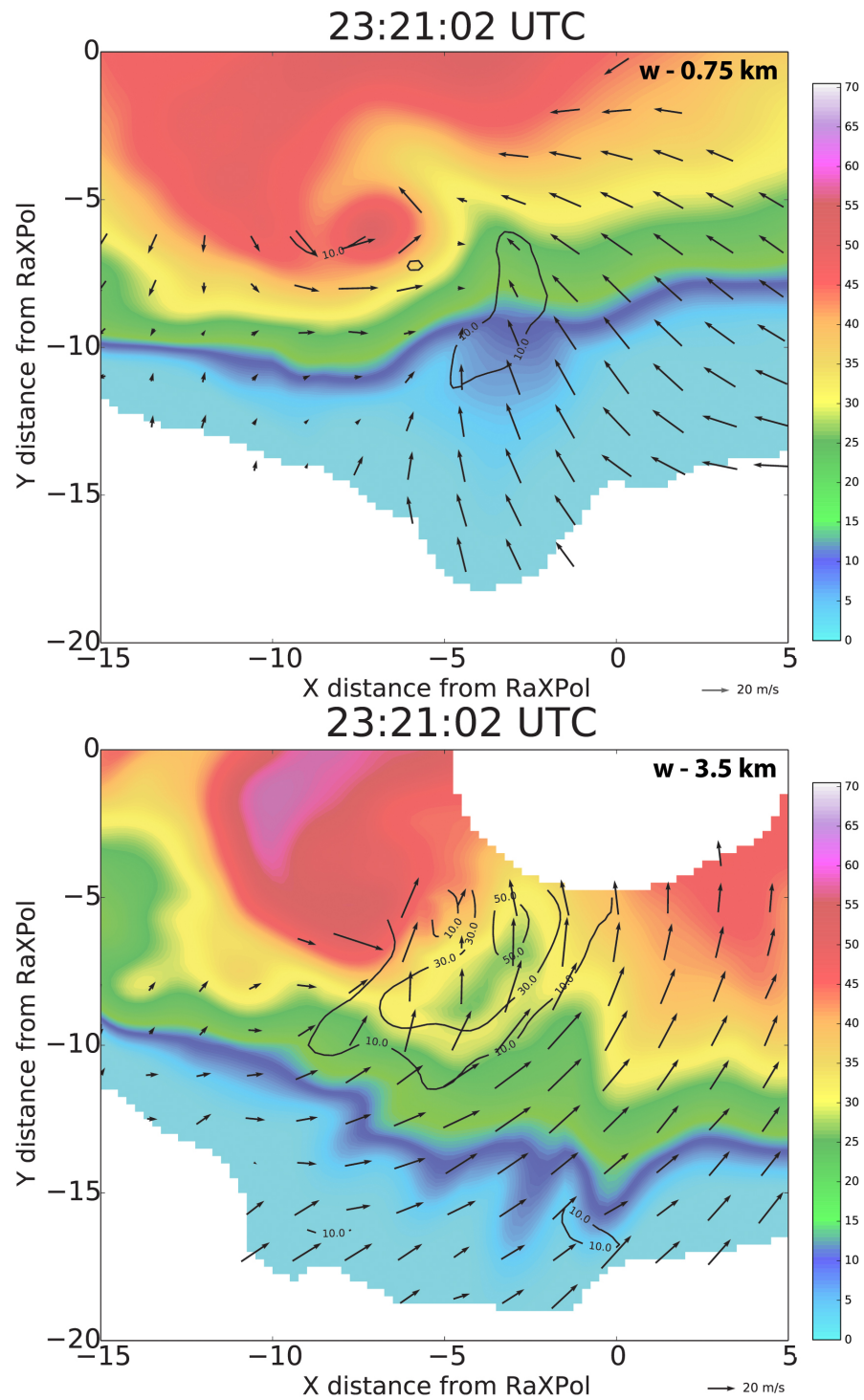


Figure 4.25: Dual-Doppler analyses of vertical velocity at 750 m AGL (top) and 3.5 km (bottom) valid at 23:21:02 UTC. Vertical velocity is plotted every 10 m s^{-1} in black. KTLX reflectivity is color contoured.

4.4.1 Variation of near-storm vertical wind shear within an Eulerian framework

Three separate analysis times of the Norman-Shawnee supercell in which a tornado was ongoing were chosen arbitrarily to span the analysis period (Fig. 4.26). A grid point (red dot) within the inflow region was chosen that was both in the inflow and in the approximate path of the tornado throughout D3 to examine effects of the tornado in an Eulerian reference frame (sensitivity to horizontal displacements at one specific time will be examined in the next section). A composite plot contains seven separate hodographs that reflect the changes in the wind field over the 20 minute period (Fig. 4.26d). Between 23:02:22 and 23:10:27, the 0-3 km vector wind difference is roughly 15 m s^{-1} at each time, and the overall shapes of each hodograph do not vary drastically; 0-1 km SRH varies between 8 and $12 \text{ m}^2 \text{ s}^{-2}$ and 0-3 km SRH varying between 22 and $27 \text{ m}^2 \text{ s}^{-2}$. This result suggests that well away from the tornado, the inflow region exhibits minimal change in time. That is, no significant variation is observed away from the tornado and mesocyclone as a function of time.

From 23:15:05 to 23:18:20, the effects of the tornado become apparent both in the magnitude of the vector wind difference and in hodograph curvature at low-levels. By this time, a long, sickle-shaped hodograph is evident at both 23:15:05 and 23:16:42, which is representative of the large increase in shear near the surface. The 0-3 km vector wind difference after 23:15:05 exceeds 40 m s^{-1} , with storm-relative surface wind speeds ranging from 30 to 34 m s^{-1} . This dramatic increase in low-level shear is also seen in SRH, with 0-1 km values nearing $70 \text{ m}^2 \text{ s}^{-2}$ and 0-3 km values in excess of $200 \text{ m}^2 \text{ s}^{-2}$.

As the supercell moves across the domain, it is evident that the steadiness of the inflow is rather sensitive to its location with respect to the tornado. Initially, both 0-1 km and 0-3 km SRH are relatively steady, increasing dramatically as the

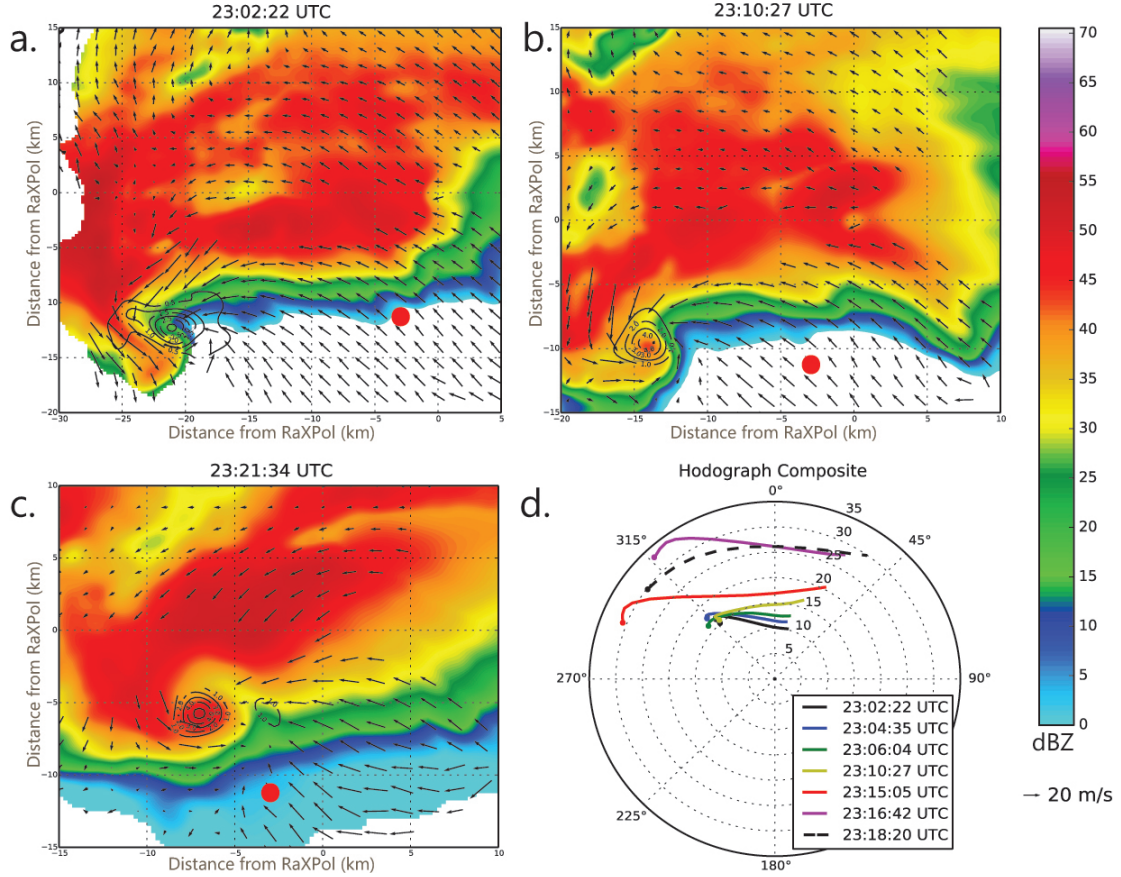


Figure 4.26: Dual-Doppler synthesized horizontal winds (black arrows; storm-relative), vertical vorticity (black contours; contoured every $1.0 \times 10^{-2} \text{ s}^{-1}$), and radar reflectivity factor (color contoured) at $z = 0.25 \text{ km}$ (a-c). The red dot in a through c represent the location of the stationary grid point from which 0-3 km hodographs in d are obtained, with respect to a moving storm. Hodograph wind speeds are in m s^{-1} , and each hodograph begins at the point indicated on the curve for each individual time.

tornado approaches (Fig. 4.27). As the tornado nears the arbitrary grid point, the values exhibit more variability, unlike the first eight volumes in which the SRH is nearly constant and little variation is observed in the hodographs. Once the tornado propagates sufficiently eastward, it nearly doubles the magnitude of both the shear vector and SRH within the observed inflow. This result indicates that the inflow of this tornadic supercell is largely unsteady, especially when brought under direct influence of the tornado vortex.

4.4.2 Examination of heterogeneity in tornado-modified inflow

As in the above section where the inflow was examined in an Eulerian sense, grid points were once again chosen arbitrarily based on the location of the tornado in order to best sample the inflow near the tornado. In this case, heterogeneity of the inflow in this region is examined at single times with 6 separate hodographs. Locations of all 6 points with respect to the supercell at 23:02:22 and 23:21:34 are color-coordinated (Fig. 4.28a, b) according to their respective hodograph (Fig. 4.28c, d). In each case, points were chosen that remained within the inflow air, but were close enough to the approaching tornado so that the effects of the tornadic circulation would still be realized. Points oriented along an east to west line are evenly spaced at 2.5 km intervals, with the northern (pink) and southern (yellow) points spaced 1.5 km from this line.

At 23:02:22, all hodographs display similar shapes with respect to one another and to results from the stationary point analysis above when the tornado was near the grid point. A distinct sickle shape is again evident from the 90° turn in the wind field at low-levels. Examining the points oriented from east to west, it is clear that as we progress closer to the tornado, both the magnitude of low-level shear and hodograph curvature increase, which is also manifested in changes of SRH (Fig. 4.29). Similarly, hodographs at 23:21:34 depict increasing shear

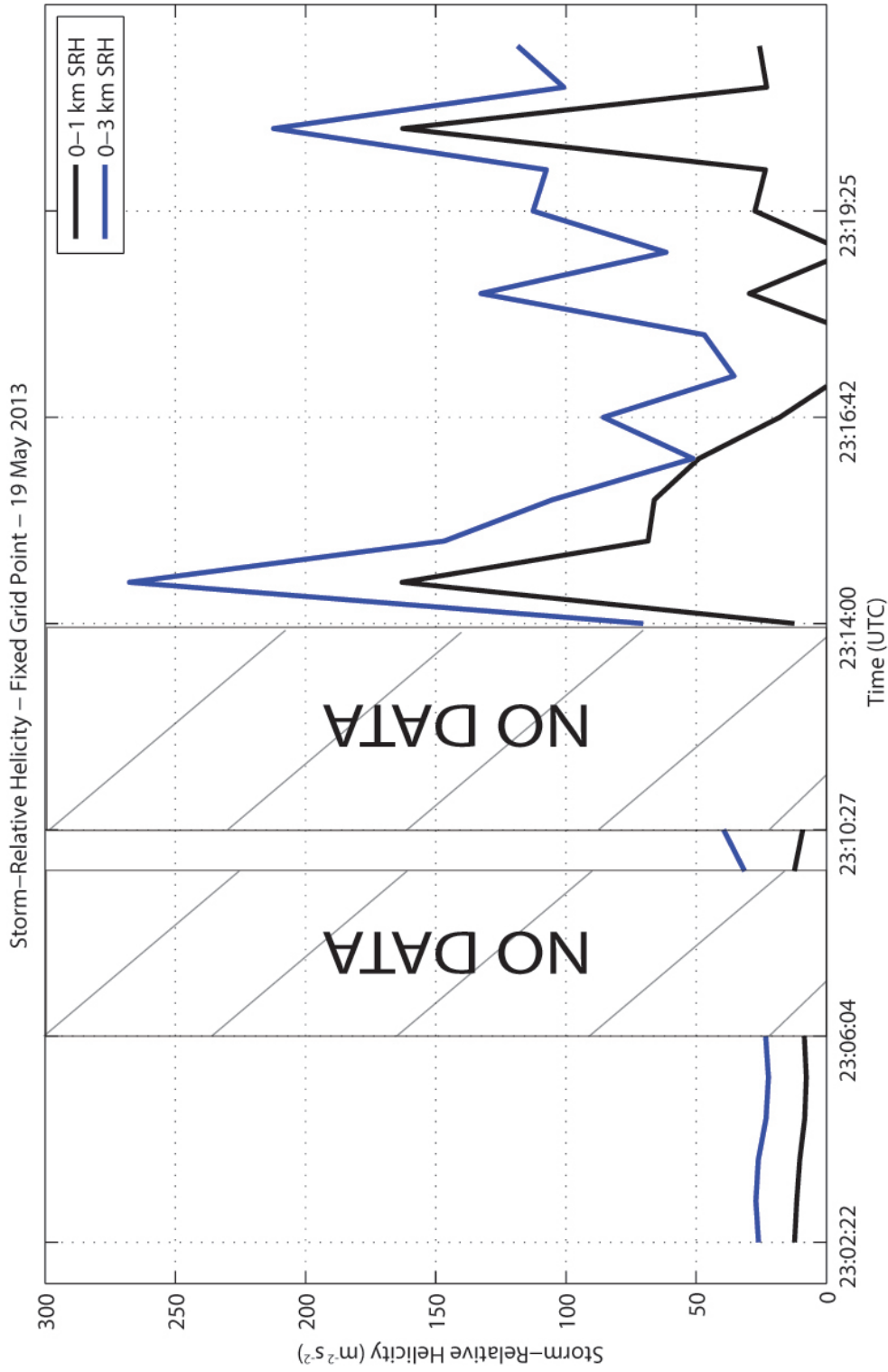


Figure 4.27: Time series of 0 – 1 km (black) and 0 – 3 km (blue) storm-relative helicity ($m^2 s^{-2}$) obtained from dual-Doppler wind syntheses valid at the stationary grid point indicated in 4.26. Data were unavailable where indicated.

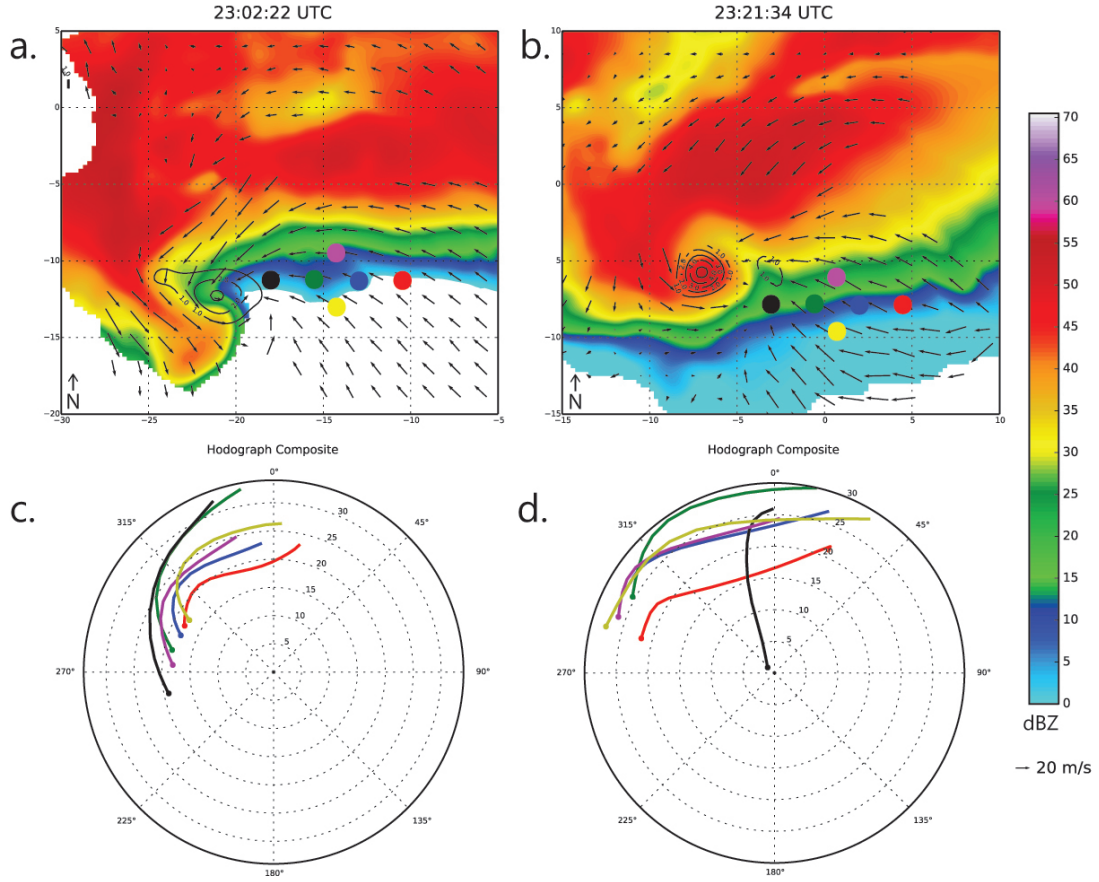


Figure 4.28: Dual-Doppler synthesized horizontal winds (black arrows; storm-relative), vertical vorticity (black contours; contoured every $1.0 \times 10^{-2} \text{ s}^{-1}$), and radar reflectivity factor (color filled) at $z = 0.25 \text{ km}$ are indicated (a, b). Colored dots correspond to hodographs in c and d, and are the locations from which each hodograph was obtained. Hodograph wind speeds are in m s^{-1} and each hodograph begins at the point indicated on the curve. Cardinal North is indicated in a and b.

nearer the tornado, except in the “black” example where the winds have slowed at low levels. This slowing is likely an artifact of the RFGF, and smearing of the boundary due to the grid resolution used in D3. In both examples, a noticeable distinction cannot be made between the north (pink) and south (yellow) points, nor in any additional times not presented here. In general, the north and south points tended to be representative of those which immediately surround them (blue/green), suggesting that proximity to the forward flank gust front is not nearly as influential as proximity to the tornado itself. Even though theory suggests that vorticity generated baroclinically in the forward flank is one major source of low-level rotation in a supercell (e.g., Rotunno and Klemp, 1985), it is not explicitly considered in this study due to the lack of thermodynamic data.

4.4.3 Discussion and consideration of SRH interaction with tornadic circulation

The previous sections presented two different examinations of low-level wind shear within the inflow region of a tornadic supercell. In the first case, it was shown that while inflow can be considered relatively steady on short time scales far from the tornado, the tornado has the potential to drastically modify its own environment over time. Heterogeneity was then examined with similar results verifying the sensitivity of SRH with respect to the tornado. However, these results indicated that other sources of horizontal vorticity, such as the FFGF, were not as influential on the vertical structure of the wind field. Figure 4.29 supports this statement; the SRH remains largely unchanged with time far from the circulation.

The low-level shear rapidly increases near the tornado owing to the modification of the environment by the tornado itself. At 23:04:35 (Fig. 4.29a), maximum SRH is near $150 \text{ m}^2 \text{ s}^{-2}$, increasing in strength to $> 200 \text{ m}^2 \text{ s}^{-2}$ by 23:06:04 (Fig. 4.29b). As the tornado strength increases, SRH continues to increase, but also seems to

be drawn inward toward the vortex. Maximum SRH was initially strongest east-southeast of the tornado and was pulled toward the strengthening tornado (Fig. 4.29c).

As the tornado reaches maturity, the supercell appears to occlude beginning around 21:15:05 (Fig. 4.29d); however, one major distinction must be clarified. Occlusion, initially defined by Lemon and Doswell III (1979), occurs when the tornado and mesocyclone are cut off from the inflow, and instead feed off of re-circulated air from the RFD. Despite insufficient thermodynamic data, high SRHs that were once being fed into the circulation from the environment have been completely cut off suggesting that the storm has begun to occlude (Figs. 4.29d-f). Even though we have no basis for determining whether or not the supercell has occluded based on the formal definition, we can make the case that the original source of SRH has been cut off from the circulation.

Throughout the lifetime of the tornado, as the low-level winds strengthen within the circulation, the associated environmental winds follow suit. This effect leads to the observed increases in low-level shear, which enhances SRH. Although it would seem as if one may be the cause of the other, the two likely occur simultaneously. While environmental SRH initially exists as a result of the large-scale environmental wind field, tornadogenesis must still act as a precursor to local modification of the low-level wind field. Prior to tornadogenesis, no local intensification of the low-level wind field exists. We can speculate that once tornadogenesis occurs, the tornado can enhance its own environment, but this is merely an effect, rather than a cause, of tornadogenesis.

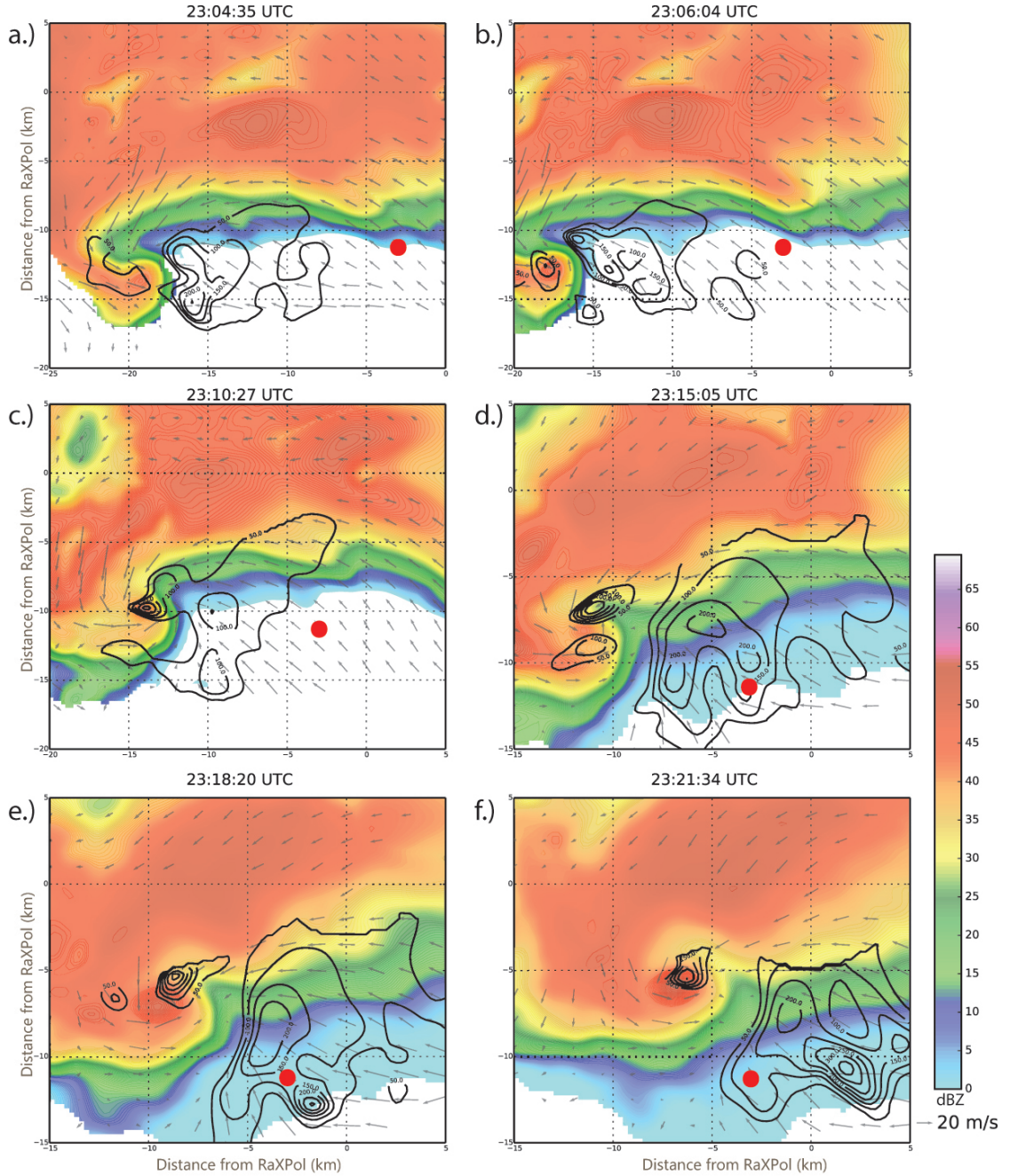


Figure 4.29: 0-3 km storm-relative helicity (black contours; $m^2 s^{-2}$) obtained from dual-Doppler wind syntheses are indicated every $50 m^2 s^{-2}$ (zero contour omitted). Storm-relative wind vectors are indicated by ghosted arrows, and radar reflectivity at $z = 0.25$ km is color filled. The red dot shows the location of the stationary grid point used in 4.26, and is indicated for reference.

4.5 Discussion of the relationship between the low-level vertical wind shear structure and tornadic vertical vorticity

To this point, this study has examined the vertical structure of vorticity in a tornado and its change in both height and time, as well as the structure of the low-level vertical wind shear in a tornadic supercell. Droegemeier et al. (1993) suggested that while the mean shear itself was not a good discriminator of rotational characteristics within a storm, the shear profile and storm motion were. They also suggested that SRH was a good indicator of both the longevity of a storm and the degree to which rotation was created. In this section, we examine the relationship that exists, if any, between SRH and the vorticity in the tornado vortex.

In Section 3 we discussed the stationarity and homogeneity of the inflow region of the storm. It was hypothesized that well upstream from the tornado and mesocyclone, the inflow was relatively steady. As the tornado approached a particular location, a rapid increase in low-level shear was observed owing to local modification of the inflow by the tornado (Fig. 4.26). The inflow was characterized by a large increase in low-level shear with minimal change upstream with respect to the tornado (Fig. 4.28). Shear tended to be maximized in the region just to the southeast of the RFGF/FFGF intersection, where the effects of the tornadic winds at low levels were the strongest.

Average maximum SRHs from timeseries data (Fig. 4.30) over the first 15 volumes obtained by RaXPol, sufficiently downstream of the tornado are approximately $120 \text{ m}^2 \text{ s}^{-2}$ between 0 and 1 km and $255 \text{ m}^2 \text{ s}^{-2}$ between 0 and 3 km. At 2000 UTC on 19 May 2013, a sounding was launched at KOUN, in which 0-1 km SRH of $131 \text{ m}^2 \text{ s}^{-2}$ was observed, and $262 \text{ m}^2 \text{ s}^{-2}$ for 0-3 km SRH. Relatively equivalent actual SRH and experimentally-obtained SRH (estimated from the dual-Doppler

analysis) give us a fair amount of confidence that the SRHs calculated well downstream of the tornado are representative of the ambient environment. Between 23:14:00 and the end of D3, SRH climbs rapidly to near $300 \text{ m}^2 \text{ s}^{-2}$ between 0 and 1 km and $375 \text{ m}^2 \text{ s}^{-2}$ between 0 and 3 km. Vertical vorticity exhibits a similar increase after 23:14:00, suggesting that the two are related. As a method of comparison, correlation coefficient values were calculated to determine if a correlation existed between the vorticity at 0.25, 1, 2, and 3 km AGL and the 0-1 km and 0-3 km SRH. The 0-1 km and 0-3 km SRH and vertical vorticity at 0.25 and 1 km were well correlated (> 0.70). However, vertical vorticity at 2 km was only weakly correlated with 0-1 km and 0-3 km SRH, and at 3 km the results were negatively correlated, which was expected given the decrease in ζ at mid-levels discussed earlier.

Although there is a strong correlation at low levels of vertical vorticity intensification with increases in SRH, the lack of any such relationship with mid-level vorticity is curious. As the tornadic circulation near the surface strengthens, it intensifies the low-level shear by increasing the surface wind speeds. If this is in fact the case, a positive feedback may exist between SRH and vertical vorticity. The vertical vorticity increases, which reduces the pressure at the center of the vortex as enhanced rotation creates a large negative perturbation pressure. This increases near surface wind speeds and SRH, and then the process starts over. Even though this idea seems plausible given the evidence provided herein, this process is much weaker (essentially non-existent) without tornadogenesis. For this to occur, an alternative forcing is needed (such as friction) to start the process.

Finally, at approximately 23:15, the mid-level vorticity is negatively correlated with the low-level vorticity (Fig. 4.19). One explanation was offered earlier, discussing the roles of the perturbation pressure gradient force on the updraft, and how stretching/shrinking of vorticity may have been occurring in this region. An

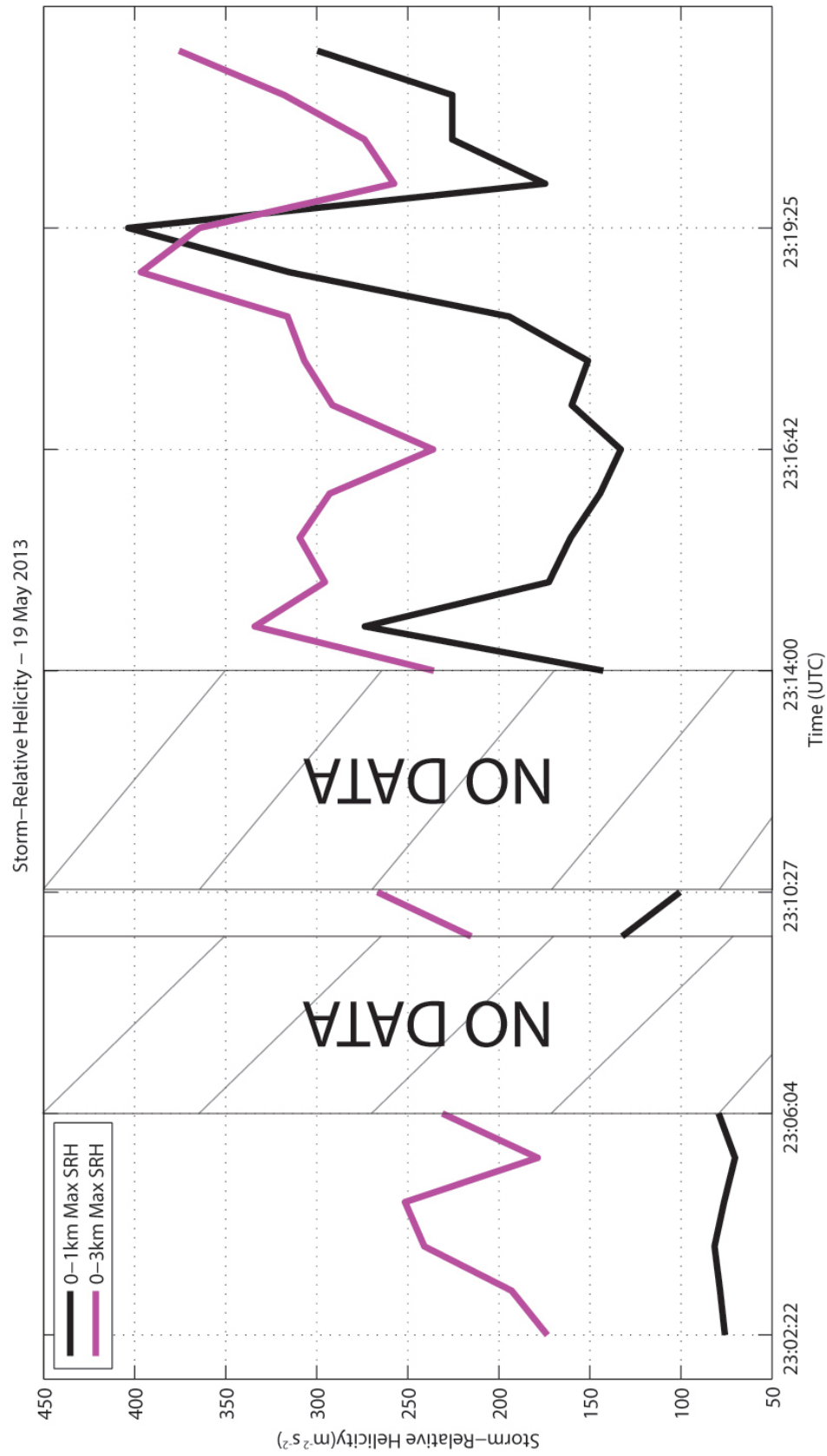


Figure 4.30: Time series of maximum 0-1 km (black) and 0-3 km storm-relative helicity ($m^2 s^{-2}$) obtained from dual-Doppler wind syntheses of the inflow region of the Norman-Shawnee supercell. Data were not available where indicated.

alternate explanation might be that a widening of the core radius of the vortex aloft due to vortex breakdown caused a substantial decrease in vertical vorticity, which is not observed near the surface. Bluestein (2013) suggests that vortex breakdown occurs when a vertical jet owing to near-surface friction drastically reduces in intensity, or even becomes a downdraft, in the end-wall vortex aloft. Flow that is initially supercritical in the upward jet near the surface becomes subcritical aloft as vorticity begins to decay, and an effect similar to a hydraulic jump is observed (Benjamin, 1962). This hydraulic jump causes flow that is laminar near the surface to widen, becoming substantially more turbulent in the process. The broadening of the vortex as a result of vortex breakdown would subsequently reduce the vertical vorticity in this region. This effect can be evaluated by examining ΔV_{max} to eliminate the potential bias associated with the size of the core radius, and will be examined in future studies.

Chapter 5

Analyses of Z_{DR} columns in the Edmond-Carney and Norman-Shawnee Supercells

This section will examine and discuss an analysis of Z_{DR} columns within the Edmond-Carney and Norman-Shawnee supercells. Z_{DR} will be examined with respect to the three-dimensional wind field derived through dual-Doppler analysis, specifically with respect to vertical velocity. This section will conclude by comparing differences between the two cases and considerations of vertical velocity strength with respect to maximum Z_{DR} column height.

5.1 The Edmond-Carney Supercell

Differential reflectivity (Z_{DR}) columns are a useful proxy for understanding changes in storm intensity, particularly in diagnosing and understanding changes in the updraft of the storm. One benefit to this study is that the dual-Doppler analyses were performed with two polarimetric radars, allowing us to diagnose not only the three-dimensional wind field, but also compare the wind field to dual-pol products.

As the supercell (and associated tornado) approached maturity, a clear Z_{DR} column¹ is observed associated with the main updraft in the supercell (Fig. 5.1a). One interesting and useful aspect of Z_{DR} columns is they also allow us to more accurately track the mesocyclone much more clearly than guessing the rough location based on theoretical models of supercells alone. This can provide insight on complex storm dynamics such as storm splitting, as well as mean storm motion of a supercell which is often difficult to track.

By 21:48:15 (Fig. 5.1b), Z_{DR} in excess of 2 dB at 3.5 km AGL closely coinciding with the regions of most intense updraft. Although we are interested in the Z_{DR} column, the data considered here (3.5 km AGL) are *below* the environmental freezing level (see Fig. 4.1e) because dual-Doppler was not possible above the freezing level during this analysis. The shape of the updraft also closely matches the shape of the underlying Z_{DR} column, further suggesting that the two are not only directly associated, but that one (the updraft) cause the other (the Z_{DR} column). However, one interesting aspect of this analysis is that, at first glance, the maximum Z_{DR} and most intense updrafts are not exactly collocated with one another. Superimposition of the horizontal wind vectors with the updraft and Z_{DR} shows that maxima in Z_{DR} generally follow that they are slightly downstream from each individual maximum in updraft intensity. Therefore, perhaps the large drops which obtain relatively high Z_{DR} s are then advected slightly downstream in the direction of the storm-relative wind speed at each level. At 21:48:15 UTC, visual representation of the updraft shape which is manifested in Z_{DR} , with exception

¹By definition, a Z_{DR} column is an extension of positive Z_{DR} **above** the environmental freezing line. In this case, the analyses are below the environmental freezing level (Fig. 4.1e), which by definition is not a true Z_{DR} column. However, the processes responsible for a Z_{DR} column (size-sorting) are identical to those that occur above the freezing level; thus, the region of relatively higher Z_{DR} will be referred to as a Z_{DR} column even though it is below the freezing level.

of the western periphery of the updraft. Dashed contours in Fig. 5.1c represent updraft intensities in excess of 5 m s^{-1} . Even though a portion of the Z_{DR} column falls outside of the $> 10 \text{ m s}^{-1}$, it can be seen that the Z_{DR} column remains supported by updrafts of near and greater than 5 m s^{-1} , and higher Z_{DR} continued to be advected out of the strongest regions of the updraft by the storm-relative wind.

Individual maxima in the Z_{DR} column height are observed at 21:48:15 UTC (Fig. 5.2). For reference, the Z_{DR} column top was calculated based on a $Z_{DR} < 1 \text{ dB}$ (i.e., the Z_{DR} column contains $Z_{DR} \geq 1 \text{ dB}$). Individual maxima in the Z_{DR} column top all correspond to separate maxima in updraft intensity. These corresponding maxima further support the finding that regions of greatest vertical velocity tend to closely correspond with the regions of enhanced Z_{DR} and maximum Z_{DR} column top.

By 21:51:43 UTC, observations of the Z_{DR} column are similar to those at 21:48:15 in that the shape of Z_{DR} column closely represents the updraft's shape, and the areas of highest Z_{DR} are closely associated with the strongest updrafts. The Z_{DR} maximum is still slightly displaced from the strongest updrafts at this time. This trend occurs not only throughout the analysis of this storm in the Z_{DR} data, but shares a similar trend in the location of the bounded weak echo region in reflectivity data (discussed in Chapter 4). It can be seen from the analysis of the maximum Z_{DR} column depth (Fig. 5.4) that the area of most intense w is growing spatially, which is reflected in the updraft data as well. In this case, as the storm grew and strengthened, continued intensification of the updraft can be observed in Z_{DR} data as well.

Toward the end of the Edmond-Carney analysis period, the results follow a similar pattern. The Z_{DR} column(s) corresponds to regions of locally stronger vertical velocity, but is slightly displaced downstream of strongest regions of vertical motion. There is qualitative agreement in the 21:57:46 UTC analysis between the

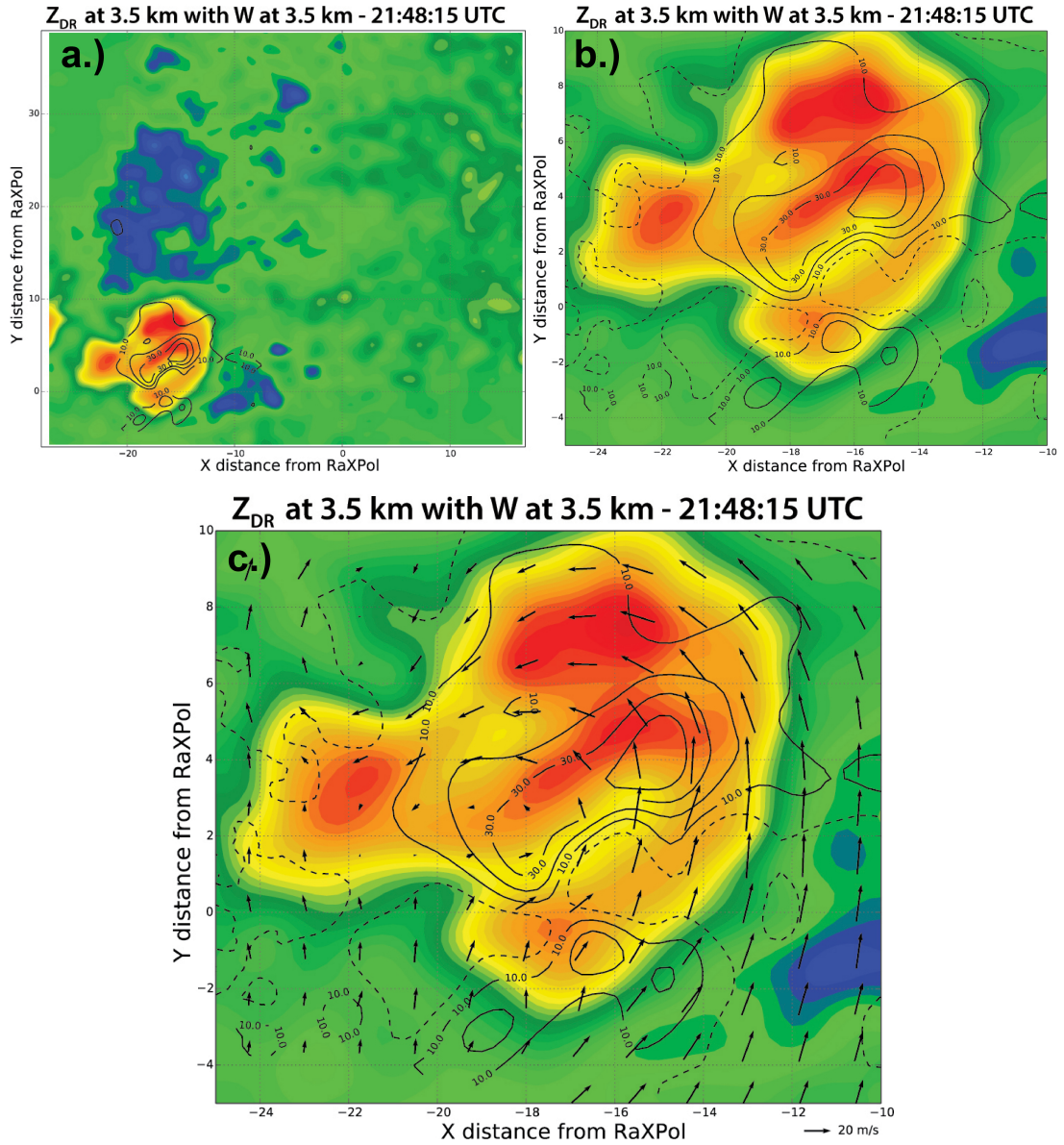


Figure 5.1: Analysis of KTLX Z_{DR} valid at 21:48:15 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in a but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid. 5 m s^{-1} vertical velocities show by dashed black contour, for reference.

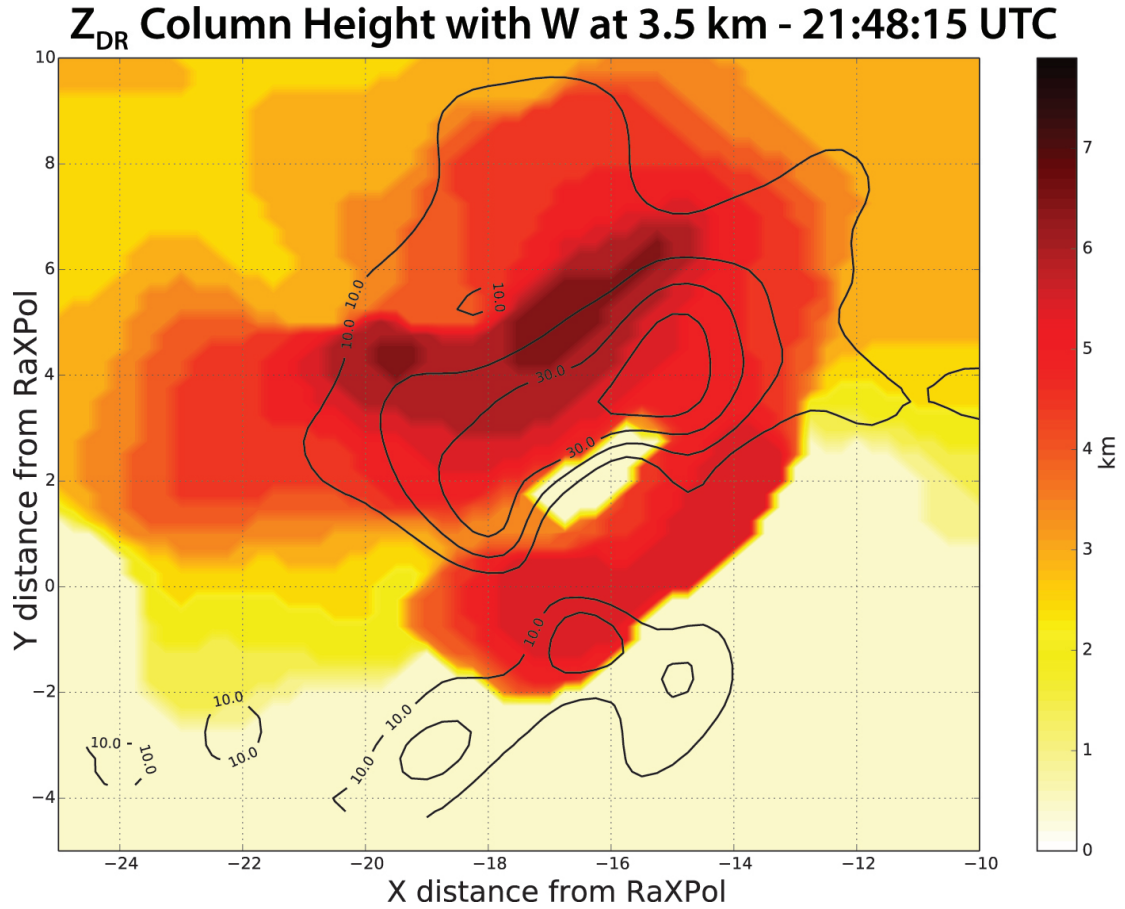


Figure 5.2: Analysis of KTLX Z_{DR} column depth at 21:48:15 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below 1 dB and is plotted in km AGL.

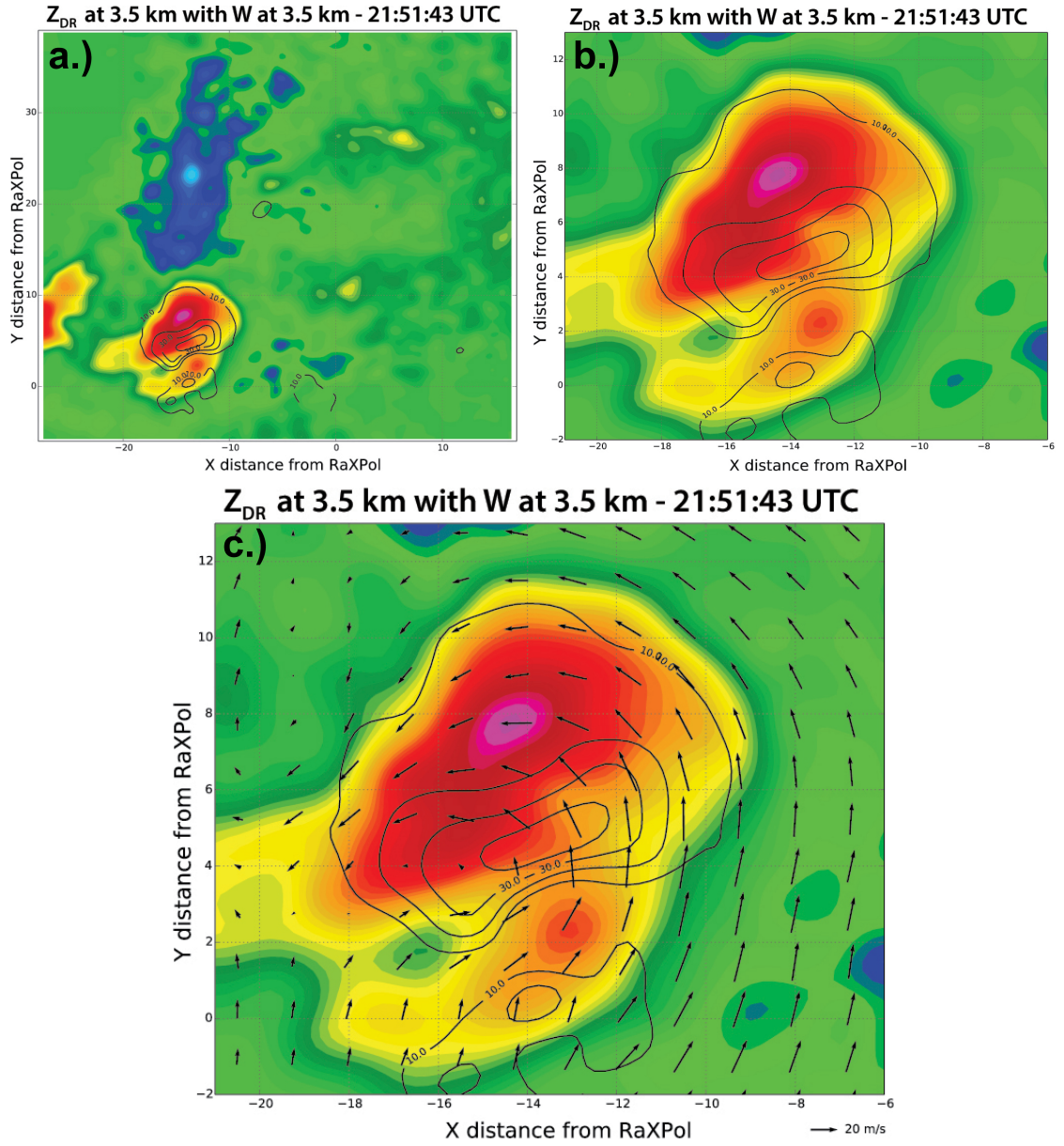


Figure 5.3: Analysis of KTLX Z_{DR} valid at 21:51:43 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in panel a but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid.

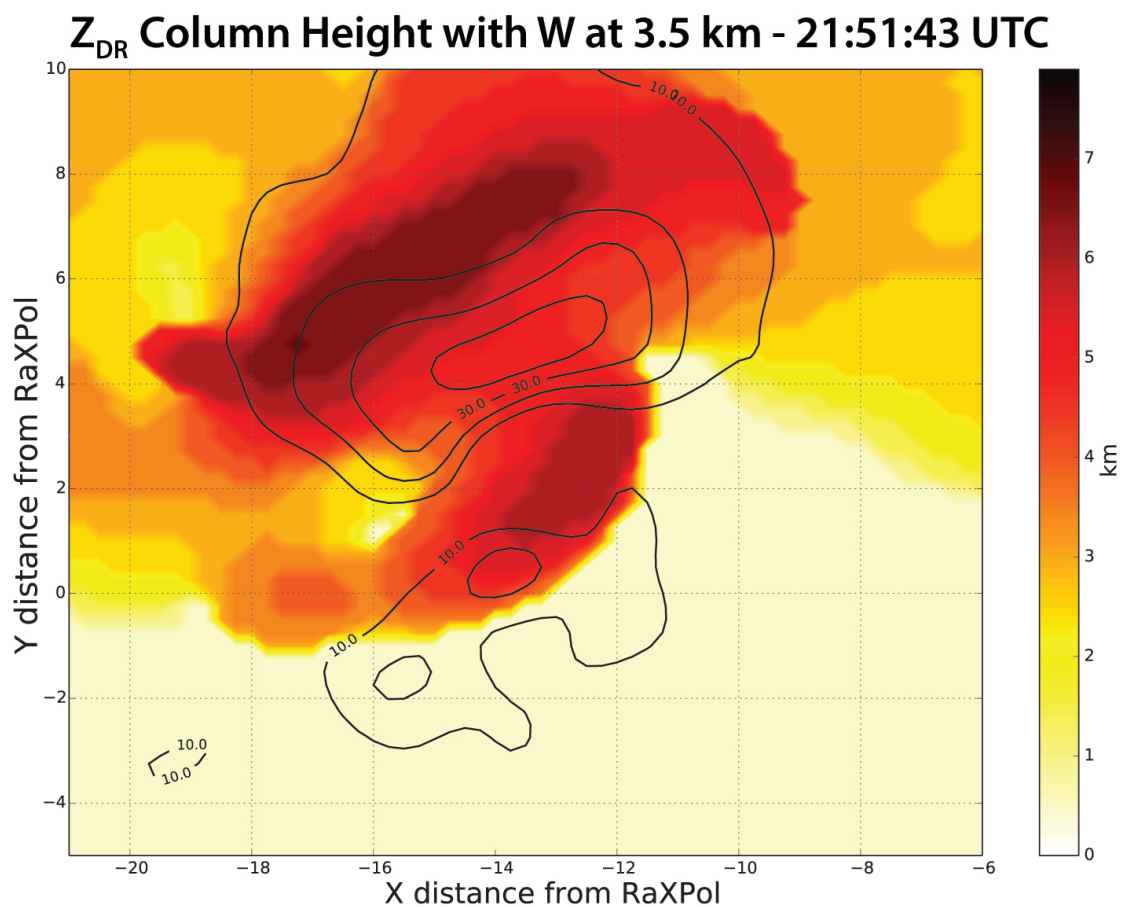


Figure 5.4: Analysis of KTLX Z_{DR} column depth at 21:51:43 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below 1 dB and is plotted in km AGL.

updraft and associated Z_{DR} columns in that three separate Z_{DR} maxima correspond to three distinct maxima in vertical velocity. The Z_{DR} column tops exhibit little change despite a weakening of the updraft. This behavior could be due to a number of reasons. First and foremost, the coarse resolution of WSR-88D data from which the Z_{DR} columns are obtained only allows for data every 500 m in the vertical. It is possible that a Z_{DR} column top hovering very near 7.5 km was pushed just above in the data, resulting in what appeared to be a very large jump in the Z_{DR} column despite much real change at all. Additionally, vertical velocity data could experience the same effect, where vertical velocities very near 40 m s^{-1} could suffer a slight decline in intensity (perhaps even due to data noise) which ultimately have little effect on the analysis.

As discussed earlier, quantitative vertical velocity data obtained by dual-Doppler synthesis should not be trusted given the numerous sources of error. Timeseries data exhibit the Z_{DR} column height as a function of time with respect to the both the maximum and mean vertical velocity within the updraft of the supercell (Fig. 5.7). Coarseness of the Z_{DR} column height line is due to the interpolation of data in between KTLX volume scans. That is, because we interpolate between volume scans in order to obtain higher temporal resolution of our analyses, the Z_{DR} data are also interpolated in the same fashion. Thus, changes in the maximum height of the Z_{DR} column can only occur at the times in which each volume was obtained. Additionally, fewer radar elevations (especially at upper levels) requires that a vertical grid spacing of 500 m be used. These factors combine to give us a rather coarse looking plot of Z_{DR} column height (Fig. 5.7; red line). The blue line represents the maximum vertical velocity within the updraft derived from the dual-Doppler analysis. The black line represents a mean vertical velocity on the domain. The mean updraft was calculated by considering the updraft as **only**

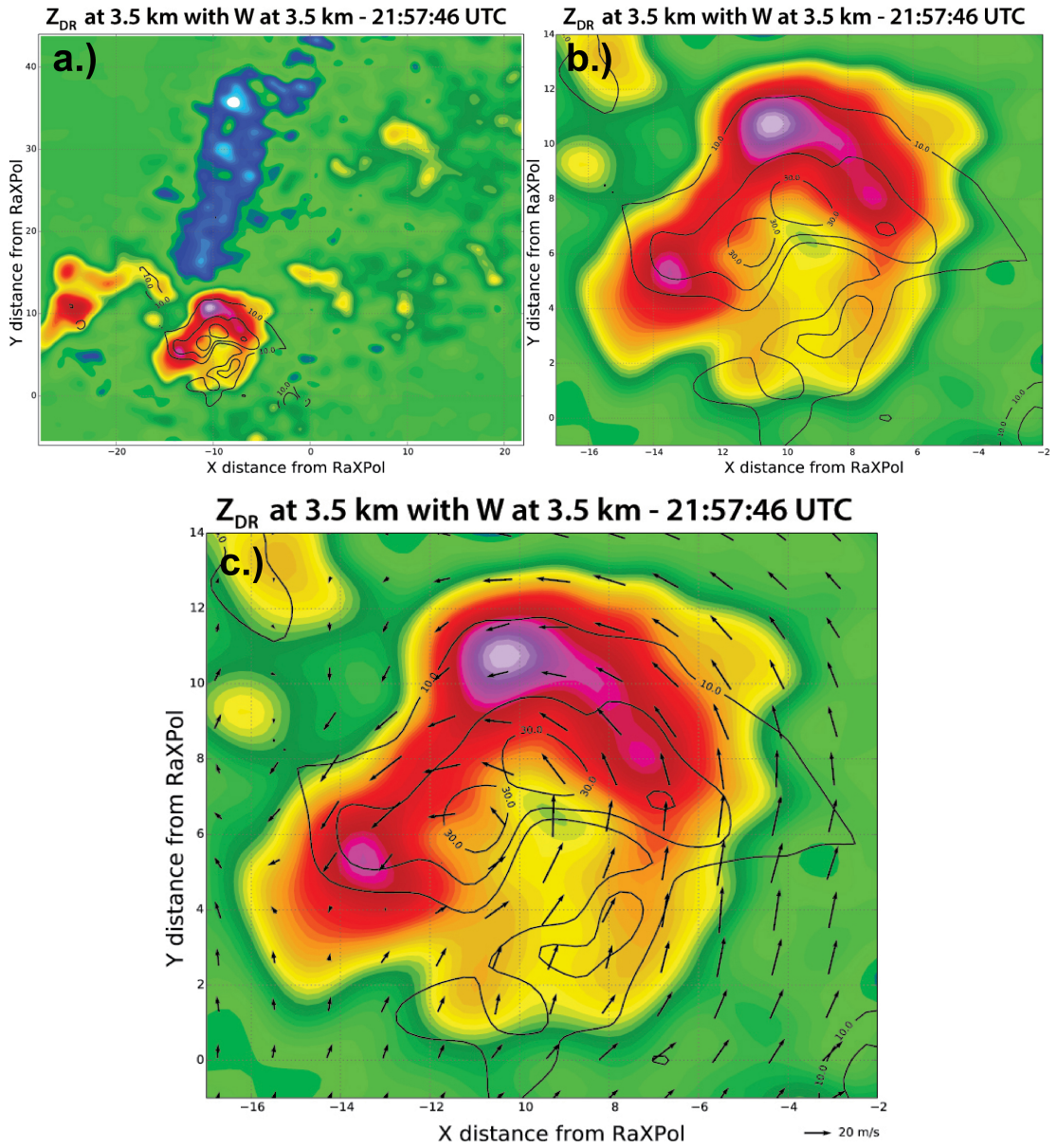


Figure 5.5: Analysis of KTLX Z_{DR} valid at 21:57:46 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in panel a, but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid.

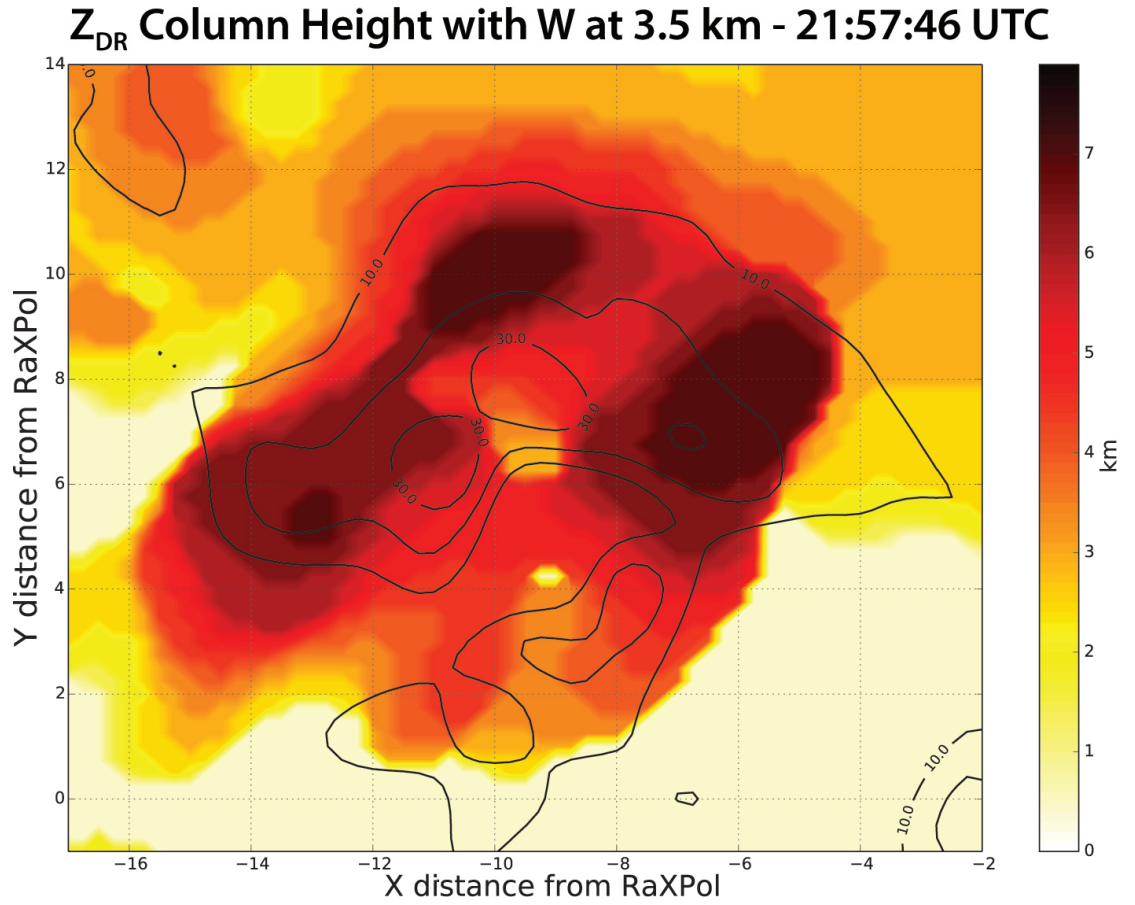


Figure 5.6: Analysis of KTLX Z_{DR} column depth at 21:57:46 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below 1 dB and is plotted in km AGL.

vertical velocities greater than 10 m s^{-1} . Calculation of the mean allows us to isolate potential significant errors in the maximum vertical velocity data, and instead gives us an idea of the general intensity of the entire updraft.

The maximum and mean updrafts decrease in velocity initially, which are followed by a sharp increase and then a roughly maintained updraft near 45 m s^{-1} over the next ten minutes. In this case, as the updraft intensifies initially, the Z_{DR} column closely follows climbing to 7 km AGL and then maintaining itself for the remainder of the analysis. Although the mean vertical velocity data are substantially smoother than the maximum, both tell the same story. As the updraft first intensifies and then remains relatively steady, the Z_{DR} column top is also steady.

As the Z_{DR} column is maintained by a steady updraft intensity, maximum Z_{DR} at 5 and 6 km are observed to behave differently. Throughout the bulk of the analysis, Z_{DR} at 5 km (solid red line) and 6 km (dashed red line) steadily increase with time while the maximum updraft (solid black line) and mean updraft (dashed black line) first increase and then remain relatively steady (Fig. 5.8). Thus, it is inferred that as the updraft is maintained, Z_{DR} is able to continue growing with time. Steady vertical velocities are able to keep larger drops lofted longer, allowing them to grow with time. Another possibility is that drops are continually recirculated into the updraft, growing as they do. Because Z_{DR} grows with time in this case, the data suggest that maximum Z_{DR} is not a good indicator of the updraft intensity. However, it is possible that maximum Z_{DR} could potentially be used as an indicator of storm behavior where Z_{DR} column data is not available (e.g., very near the radar). This usefulness hypothesis of this hypothesis must be investigated further.

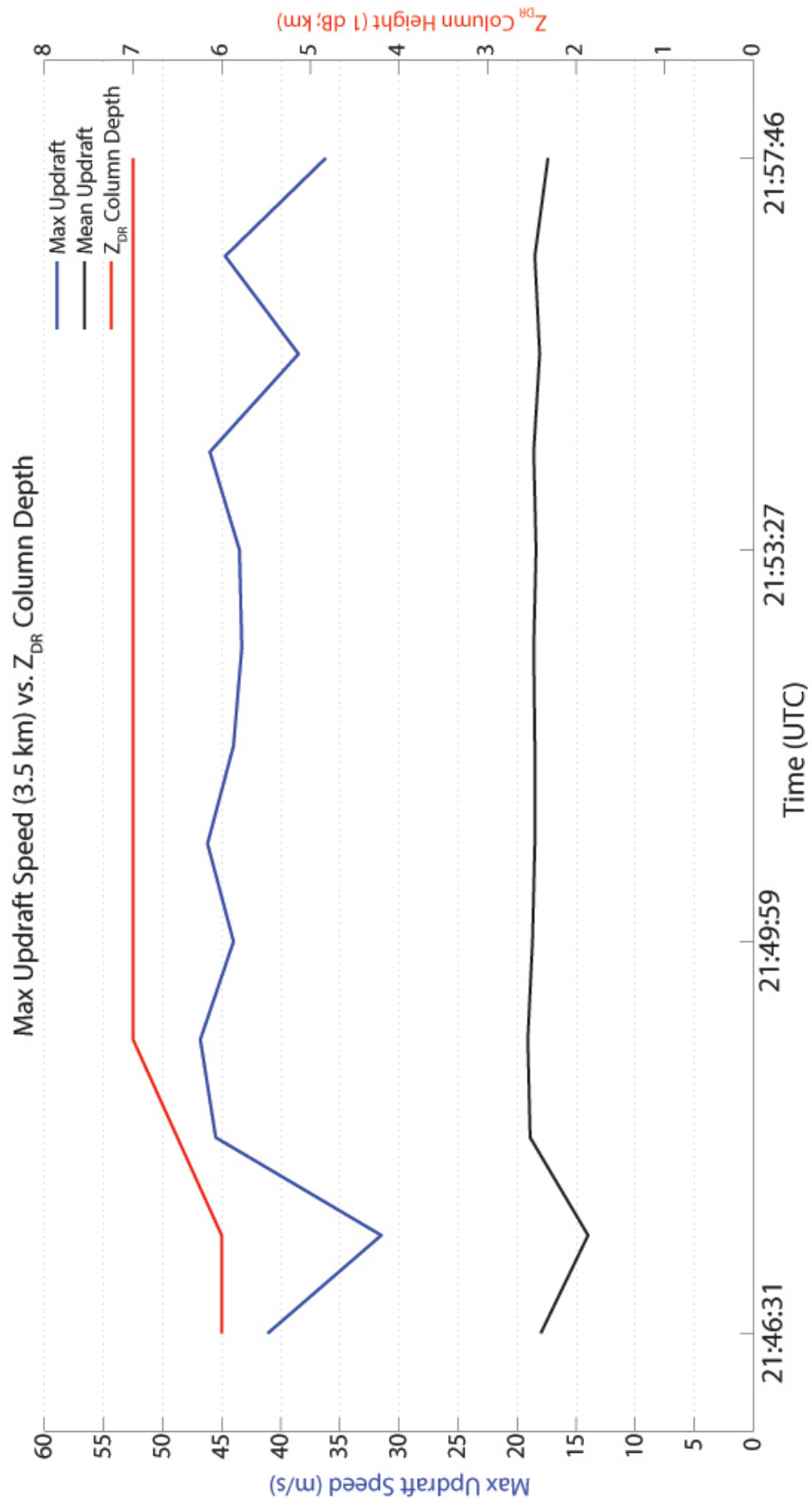


Figure 5.7: Time series of the maximum Z_{DR} column depth (red line), and mean (black) and maximum (blue) dual-Doppler derived updraft intensities for the Edmond-Carney tornado. Z_{DR} column depth is in km, and vertical velocities are in $m\ s^{-1}$.

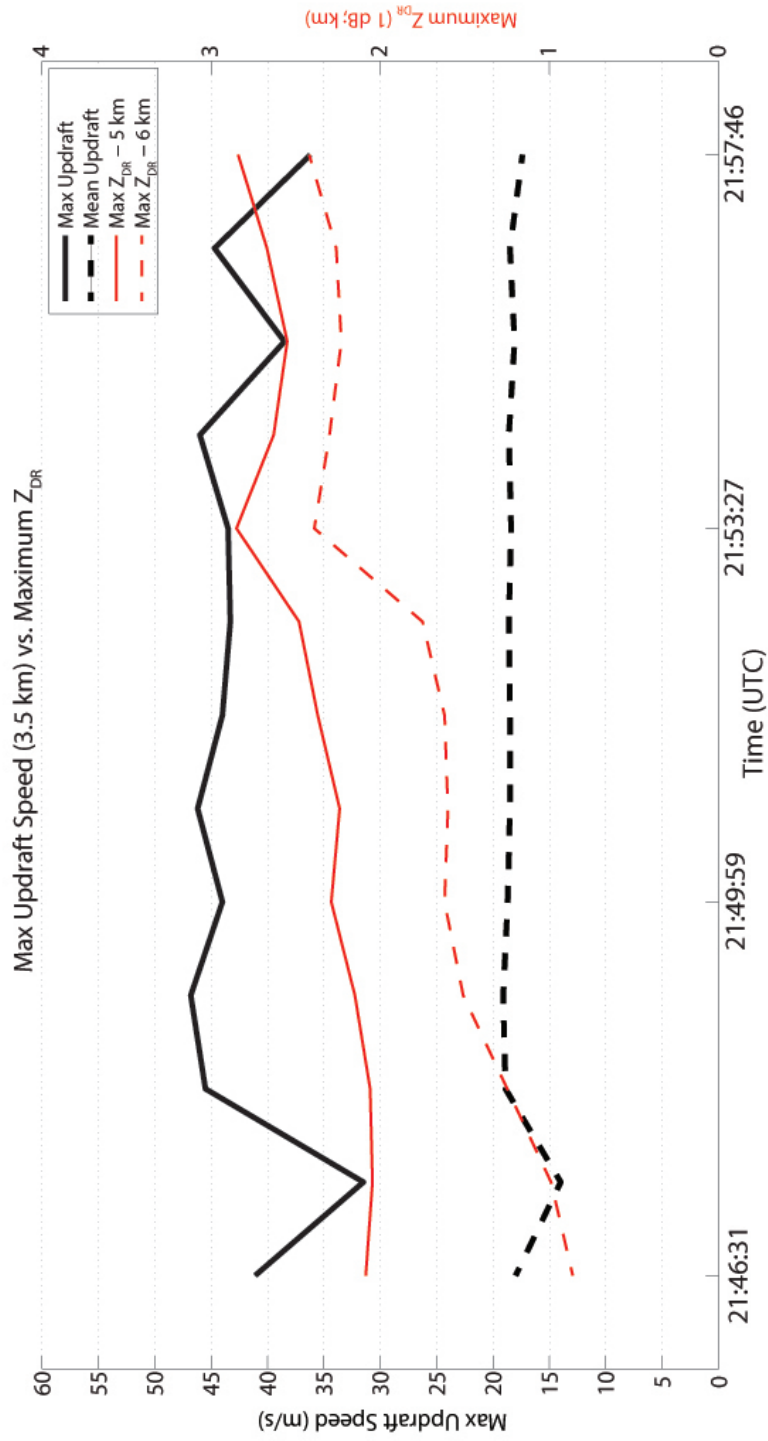


Figure 5.8: Time series of the maximum Z_{DR} at 5 km (solid red) and 6 km (dashed red), and mean (dashed black) and maximum (solid black) dual-Doppler derived updraft intensities for the Edmond-Carney tornado. Z_{DR} is in dB, and vertical velocities are in $m s^{-1}$.

5.2 The Norman-Shawnee Supercell

As in the preceding discussion, an analysis of Z_{DR} with respect to the supercell updraft was also done for the Norman-Shawnee supercell. Unlike the Edmond-Carney case, proximity of the supercell to the radars made analysis of the Z_{DR} column impossible at times where the Z_{DR} column top extended into the cone of silence.

As the supercell propagated northeastward, dual-Doppler data points near the updraft began to move across the baseline causing them to be unavailable/unusable. Regardless, the results still provide enough information about the updraft and Z_{DR} column to make this data set useful. At 23:15:38 UTC, the analysis shows enhanced Z_{DR} at 3.5 km coincident with the updraft of the supercell (Fig. 5.9). Unlike in the Edmond-Carney case, the Z_{DR} column and strongest vertical velocities are vertically stacked on top of one another throughout the course of observation. The edge of the dual-Doppler lobe can be seen on the northern side of the domain where vertical velocity data and horizontal wind vectors are no longer shown. At this particular time, maximum w in excess of 50 m s^{-1} is located on the eastern side of the updraft. A secondary maximum is observed just east of the primary maximum, with vertical velocities exceeding 40 m s^{-1} . At the location of the secondary maximum we do not observe Z_{DRs} nearly as high as those collocated with the primary maximum.

One possible explanation is that development of the enhanced vertical velocities on the western side of the updraft occurs first, preceding the development of larger drops and enhanced Z_{DR} . By 23:18:20 UTC, both the primary, eastern maximum and the secondary, western maximum have grown spatially and in intensity (Fig. 5.11). Vertical velocities in both maxima exceed 50 m s^{-1} . At this time, the second updraft maximum is now associated within a region of relatively high Z_{DR} as hypothesized. Thus, it is inferred that the development of enhanced Z_{DR} lags

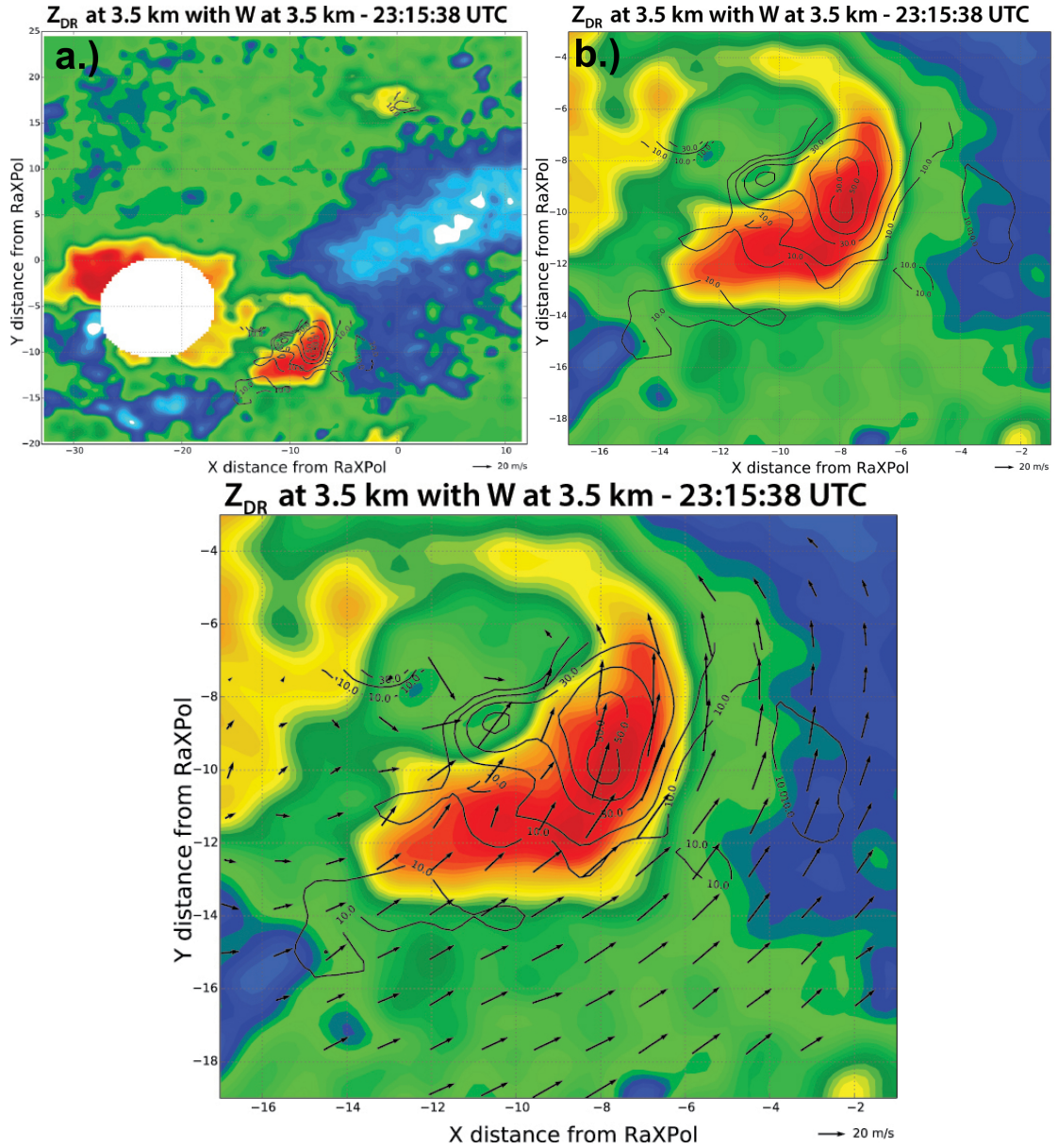


Figure 5.9: Analysis of KTLX Z_{DR} valid at 23:15:38 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in a but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid.

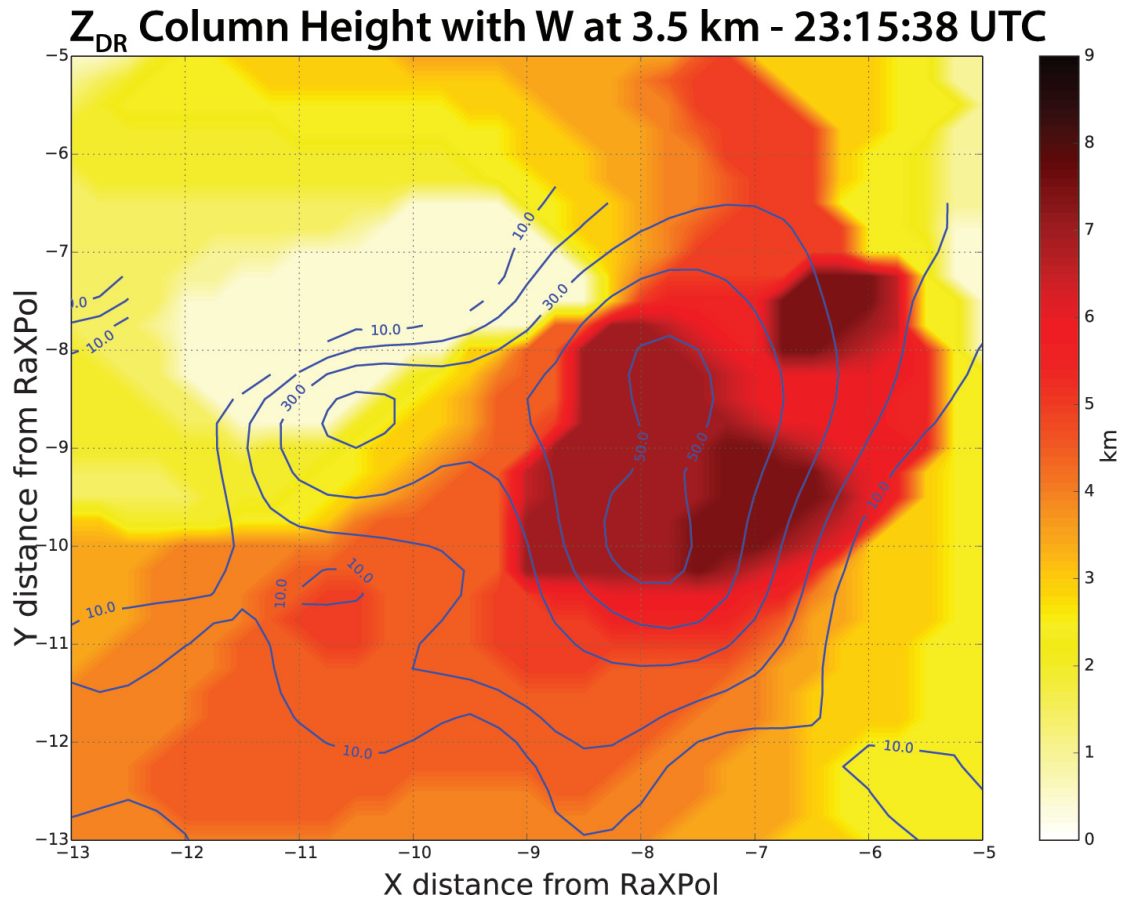


Figure 5.10: Analysis of KTLX Z_{DR} column depth at 23:15:38 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every $10 m s^{-1}$). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below $1 dB$ and is plotted in km AGL.

changes in the updraft at a particular height AGL (i.e., this may not be the case at the Z_{DR} column top).

In a comparison of Z_{DR} column height to updraft intensity at both 23:15:38 UTC and 23:18:20 UTC (Figs. 5.10 and 5.12), it is seen that the updraft maximum generally coincides with the Z_{DR} column top, with the exception again being a 23:15:38 UTC, where the secondary maximum develops ahead of observed enhancements in Z_{DR} . At both times, northwest of the maximum Z_{DR} column top, the column height is observed to step down rapidly, quickly becoming zero northwest of the updraft. As mentioned earlier, because of the proximity of the storm to KTLX at this time, the Z_{DR} column top falls above the maximum beam height within the cone of silence. In this case, KTLX is located to the northwest of the domain; thus, the cone of silence cuts off data at lower and lower elevations closer to the radar. The result of this is the substantial drop in the Z_{DR} column top caused by a lack of upper-level data.

At 23:21:02 UTC, the contours of Z_{DR} at 3.5 km again match the general shape of the updraft; however, the maximum in Z_{DR} is located to the southwest of the most intense updraft (Fig. 5.13). In comparison with the Z_{DR} column, the updraft closely matches not only the maximum in Z_{DR} column height, but the shape and structure of the updraft itself. A maximum Z_{DR} column height of 8.5 km is collocated with updraft speeds greater than 50 m s^{-1} , and the southwestward extension of the updraft closely matches the structure of the Z_{DR} column. These observations suggest that the analysis is not only providing a better understanding of the strength of the updraft by way of the Z_{DR} column height, but that it may also be possible to better understand the structure of the Z_{DR} column as well. Both of these features could lead to a better understanding of the structure of a supercell, and more importantly, how that structure is changing on a more rapid time scale.

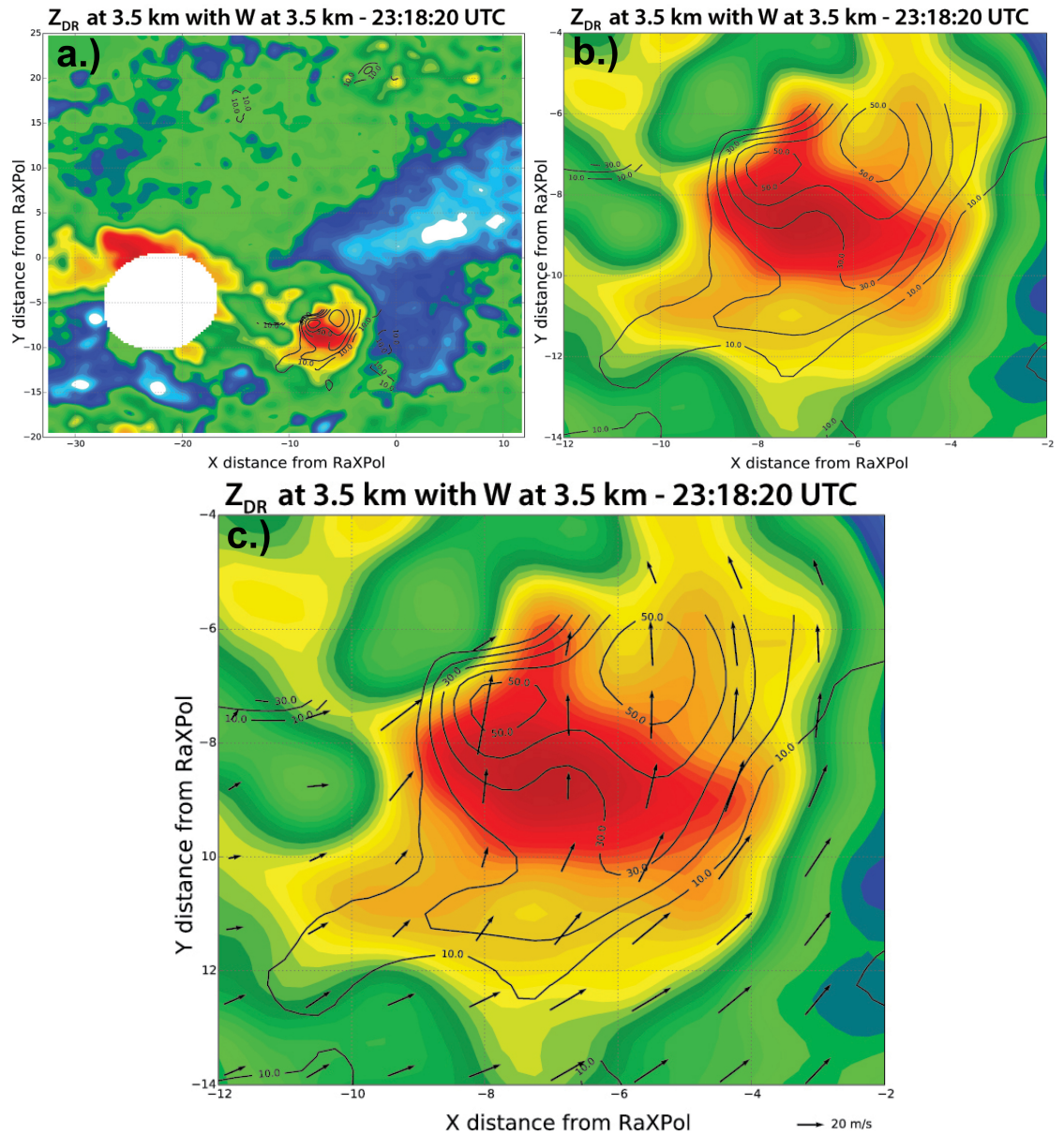


Figure 5.11: Analysis of KTLX Z_{DR} valid at 23:18:20 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in a but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid.

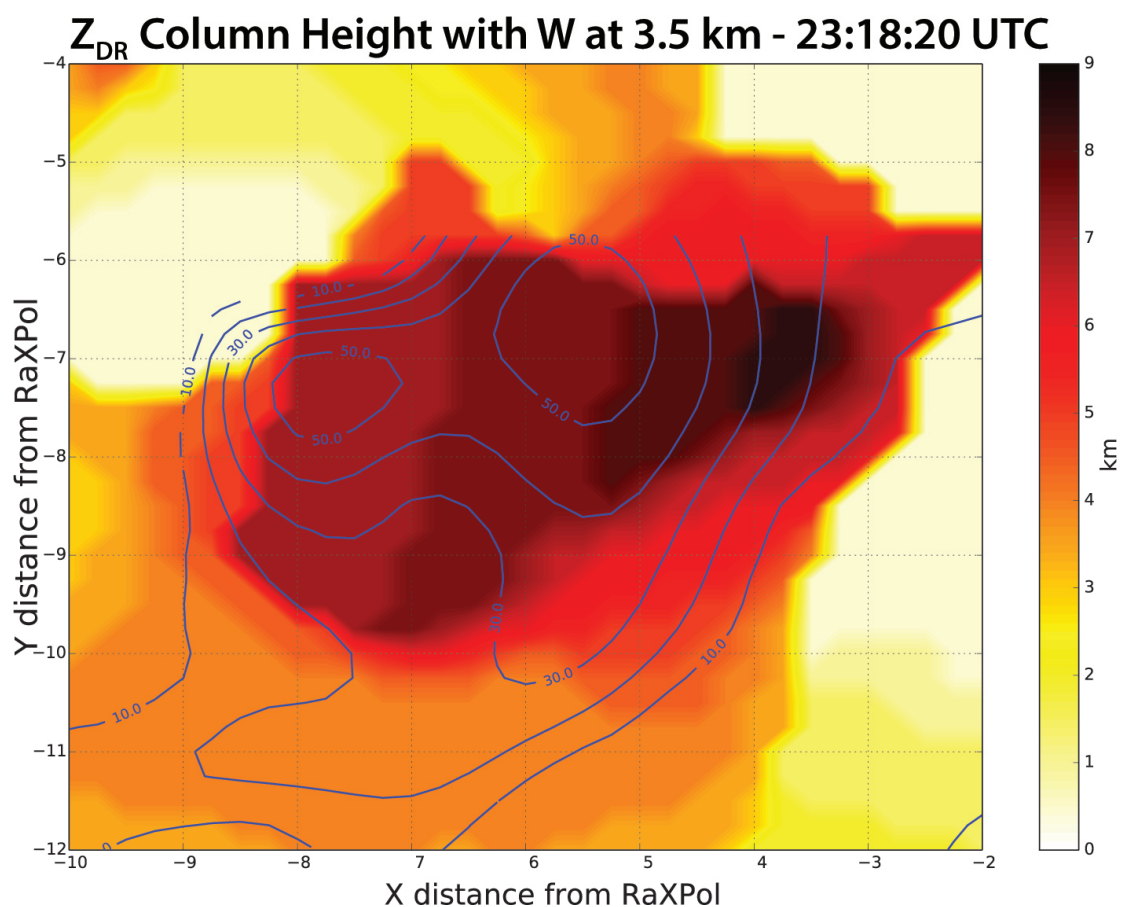


Figure 5.12: Analysis of KTLX Z_{DR} column depth at 23:18:20 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below 1 dB and is plotted in km AGL.

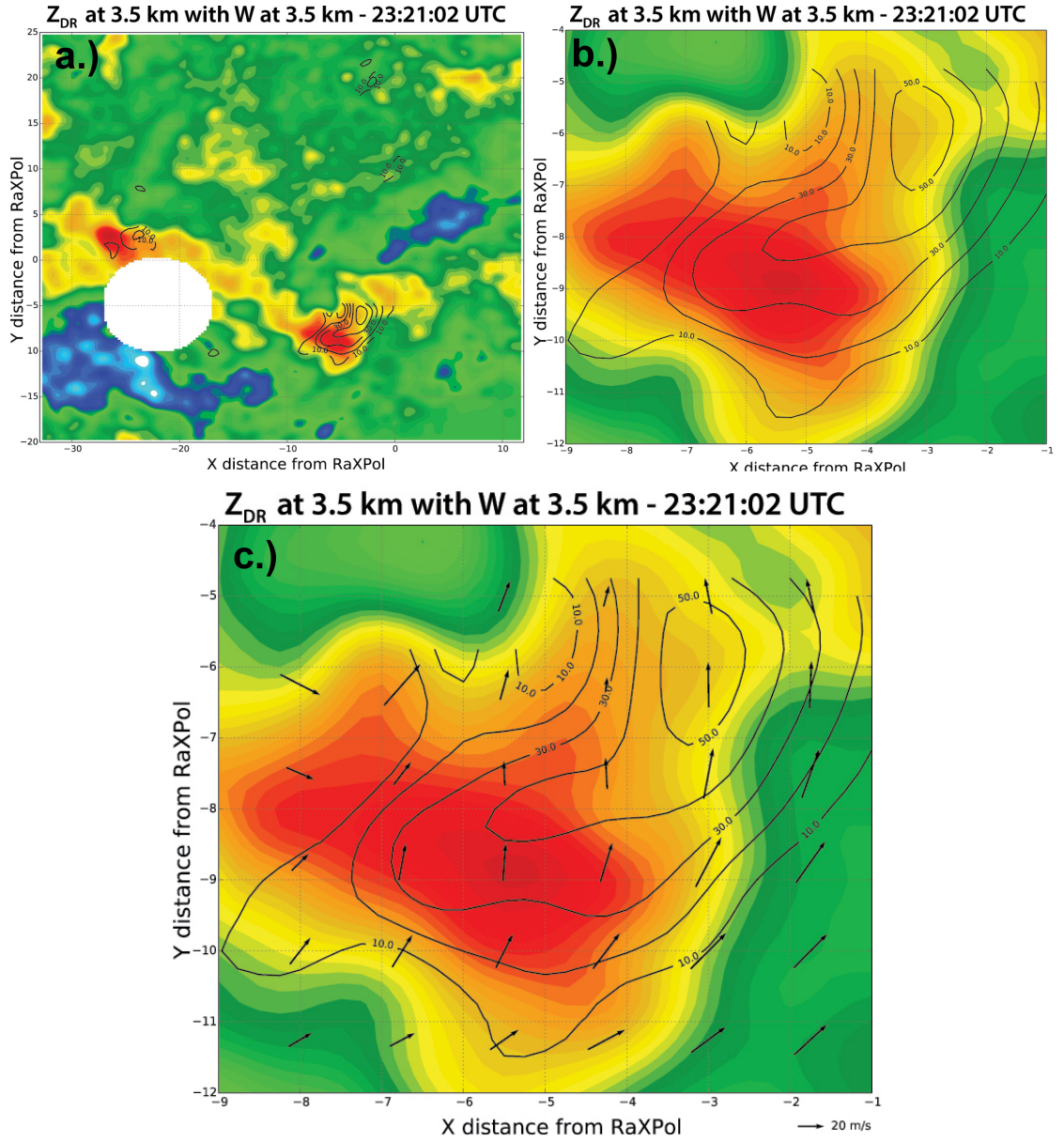


Figure 5.13: Analysis of KTLX Z_{DR} valid at 23:21:02 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). Panel a is a wide view of Z_{DR} , and panel b is the same as in a but zoomed in on the region of interest. Panel c is the same as in b, but with horizontal velocity vectors overlaid.

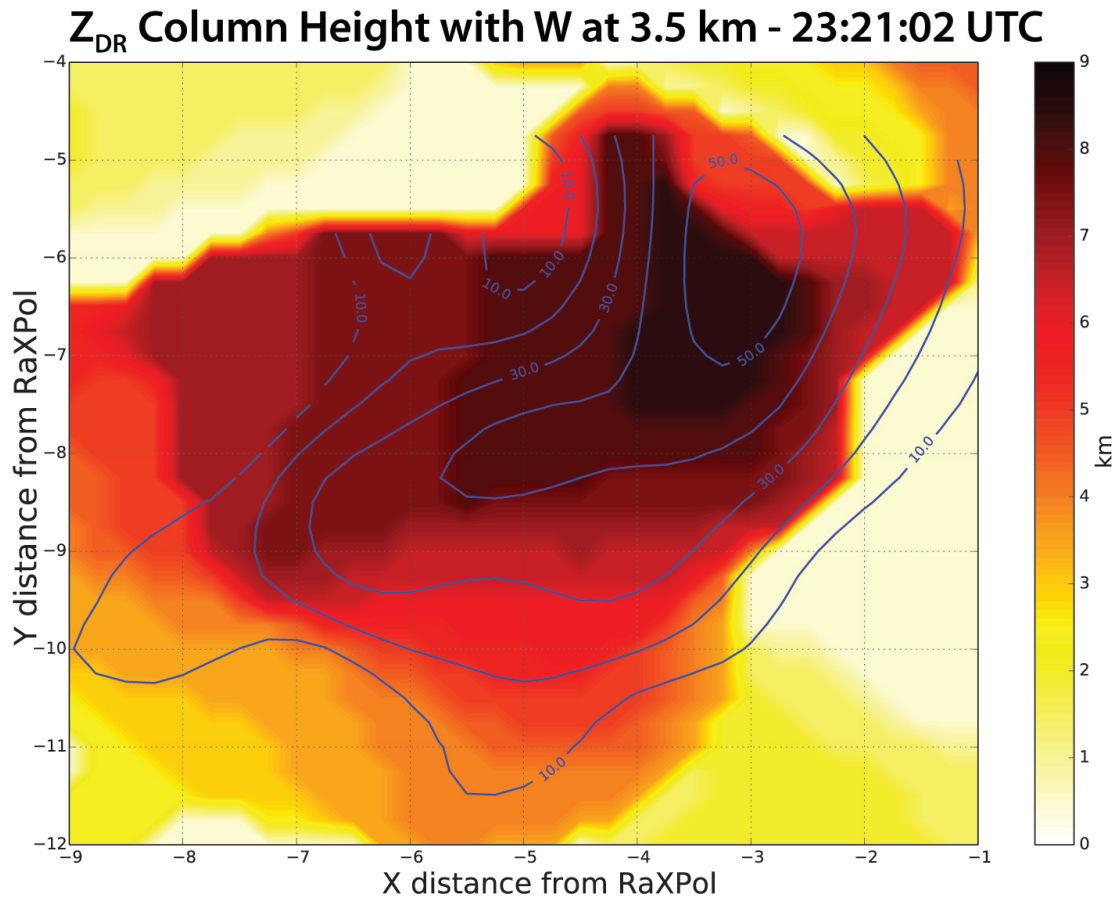


Figure 5.14: Analysis of KTLX Z_{DR} column depth at 23:21:02 UTC overlaid with dual-Doppler derived vertical velocities (solid black; every 10 m s^{-1}). In this analysis, the Z_{DR} column top is defined to be where Z_{DR} drops below 1 dB and is plotted in km AGL.

As in the Edmond-Carney case, timeseries data of the Z_{DR} analysis reflect how the maximum Z_{DR} column top changed with respect to both the maximum and the mean updraft (Fig. 5.15). For clarity, the mean updraft was calculated as before, considering the updraft to be only vertical velocities greater than 10 m s^{-1} .

For the first two minutes of the analysis, the maximum updraft intensity appears relatively steady (albeit noisy) while the mean updraft begins to slowly increase over time. During this period, the maximum Z_{DR} column height also remains largely steady until it begins to increase in height around 23:16:30 UTC. It was at this time that the maximum updraft intensity also began to increase and the mean velocity data generally continued to slowly increase throughout the analysis. After 23:18 UTC, the Z_{DR} column top remains steady at 8.5 km while the maximum updraft appears to begin to rapidly decrease. It is worth noting that this behavior is likely due to the cone of silence in the RaXPol data, causing the maximum updraft at 3.5 km to occur outside of the dual-Doppler domain. Although both maximum and mean updrafts reflect an updraft intensification, these results support the conclusion that the maximum updraft intensity most closely reflects changes in the Z_{DR} column itself.

Maximum Z_{DR} s at 5 and 6 km are similar to those in the Carney-Edmond case, except that in this case the updraft is generally strengthening with time, not remaining steady. The maximum Z_{DR} at each level both closely follow the trends in the maximum updraft, and less closely to trends in the mean. As the maximum updraft approaches its peak intensity at approximately 23:19:30 UTC, maximum Z_{DR} at 6 km actually exceeds the maximum Z_{DR} at 5 km. This behavior is likely due to the updraft lofting larger drops higher into the storm as it continued to intensify. Although the relationship between the updraft and location and growth of maximum Z_{DR} is not well understood at this time, comparison of the two may prove to be a useful proxy for better understanding changes in storm intensity.

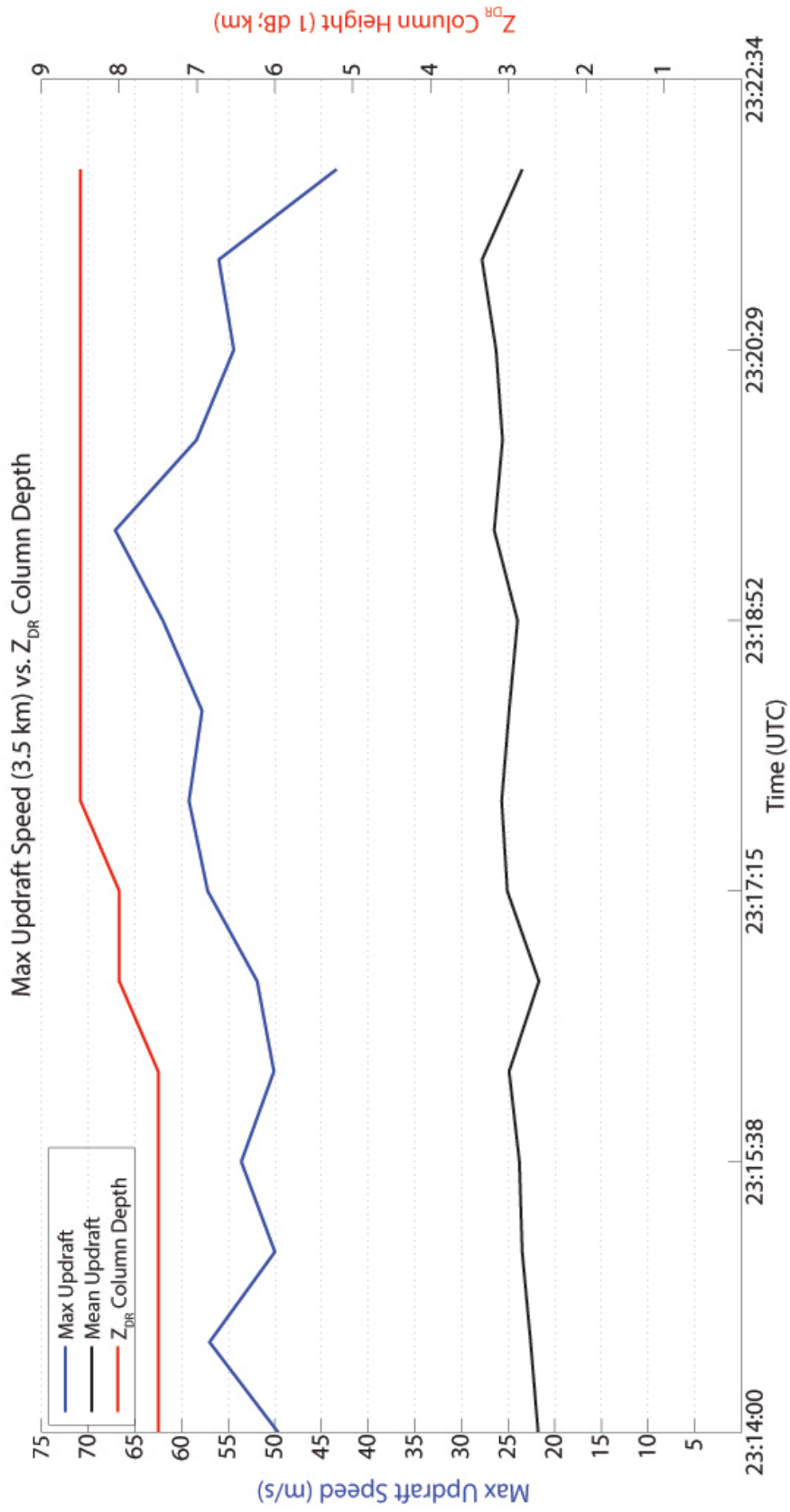


Figure 5.15: Time series of the maximum Z_{DR} column depth (red line), and mean (black) and maximum (blue) dual-Doppler derived updraft intensities for the Norman-Shawnee tornado. Z_{DR} column depth is in km, and vertical velocities are in $m s^{-1}$.

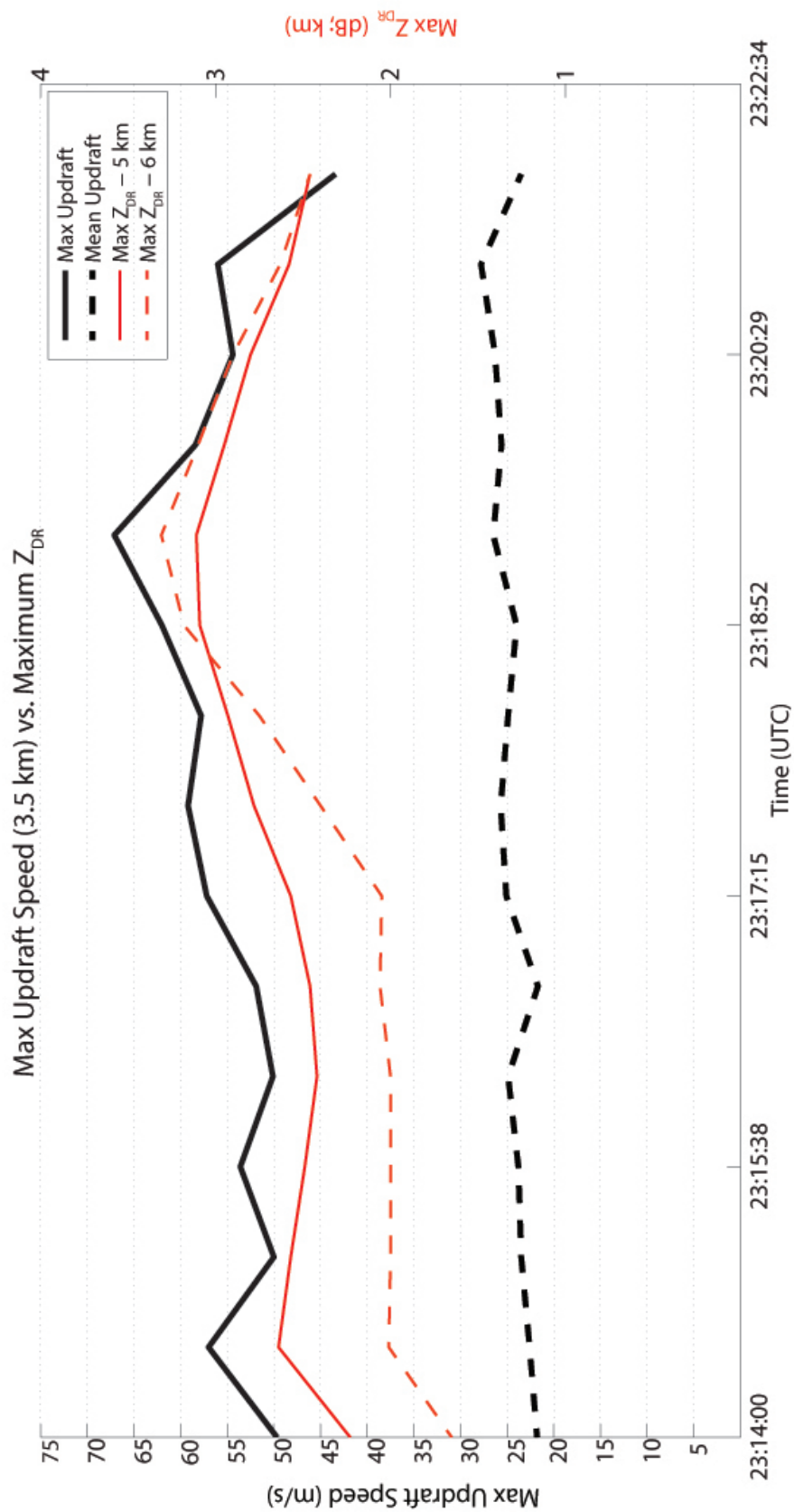


Figure 5.16: Time series of the maximum Z_{DR} at 5 km (solid red) and 6km (dashed red), and mean (dashed black) and maximum (solid black) dual-Doppler derived updraft intensities for the Norman-Shawnee tornado. Z_{DR} is in dB , and vertical velocities are in $m s^{-1}$.

Further research could help provide a better understanding to the relationship between the two, which could serve as a useful tool for nowcasting in severe storm events.

Chapter 6

Summary and Conclusions

On 19 May 2013, RaXPol and KTLX documented two violent tornadoes in central Oklahoma, one near Edmond and Carney, OK, and the second near Shawnee, OK. The motivation behind this study was to investigate a method to usefully combine two data sets with drastically different temporal resolutions. In general, a dual-Doppler synthesis between two radars with drastically different temporal resolutions is challenging because the analysis will technically be valid only over a short period of time. The goal was to utilize both data sets in a way that the would preserve the temporal resolution of RaXPol, and given that these two radars collected data of two separate supercells, the data could be compared to evaluate the technique.

To best accomplish this task, KTLX analyses from two consecutive radar volumes were interpolated linearly in a Lagrangian sense using spatially variable advection fields. This procedure allowed for the data to be interpolated to times valid at the same time as individual RaXPol volumes. Once this procedure was accomplished, a dual-Doppler synthesis was completed for both the Edmond-Carney and Norman-Shawnee tornadoes. In both cases, dual-Doppler syntheses were performed and the results of the analyses were compared to previous studies of tornadic supercells in an attempt to assess both the feasibility and potential usefulness of implementing this technique.

In the Edmond-Carney case, ζ was observed to slowly intensify throughout the analysis, which occurred between tornadogenesis and the maximum radar-observed radial velocity greater than 75 m s^{-1} . Individual analyses showed that the most intense vorticity maximum was collocated with the tornado and mesocyclone throughout the lifetime of the tornado, as expected. The magnitude of vertical vorticity in both storms were on the order of those found in strong low-level mesocyclones, but given that the analysis took place while the storm was producing a strong tornado, the values did not initially reflect typical vertical vorticity values in a tornado. Additionally, the strongest vertical vorticities in this case occurred more than a kilometer AGL. While this is consistent with observational and numerical findings utilizing grid spacings of approximately 250 m (e.g., Wicker and Wilhelmson, 1995; Dowell and Bluestein, 2002a), it is not consistent with high-resolution radar observations tornadoes in which the vertical vorticity generally increase towards the ground (e.g., Burgess et al., 2002; Alexander and Wurman, 2005; Wurman et al., 2007b; French et al., 2014). However, a number of key differences separate this case from findings of high-resolution radar studies. Data in this case were prepared for dual-Doppler syntheses by first interpolating them to a Cartesian grid with 250 m horizontal grid spacing, and subjecting them to a Barnes’ smoothing parameter. With a grid resolution of 250 m, features less than 500 m are not resolved. Throughout this analysis, the Edmond-Carney tornado’s estimated core radius grew from 290 m to 350 m, well short of the necessary 500 m needed to be resolved. If it is assumed that the vortex grows radially with height to the low-level mesocyclone, it is likely that we are seeing the vorticity maximum occurring at the lowest level AGL that the vortex is capable of being resolved.

As the analysis advanced, vertical profiles of ζ showed the maximum *descend* towards the surface. This is likely due to intensification of vertical vorticity at low

levels, but can also likely be attributed to widening of the core radius. As the core radius of the tornado grew, the analysis was capable of resolving the vortex lower to the ground.

Vertical velocity data were also analyzed from this case and compared to expectations of the traditional supercell model, and quantitative results were compared to previous dual-Doppler studies. In a qualitative sense, vertical velocity analyses compared well with theoretical models of supercell thunderstorms, specifically revealing an updraft on the eastern periphery of the hook echo curving toward the northwest into the notch north of the hook. One unexpected result from this analysis was the slight downstream displacement of the BWER. Generally speaking, because the BWER is associated with strong vertical velocities evacuating hydrometeors out of the updraft, it was expected that the two would be collocated. In this case, we hypothesize that updraft tilt perhaps may have caused the downstream displacement that was also observed in analyses of the Z_{DR} column. Previous work has presented similar results (e.g., Markowski et al., 2012a; Calhoun et al., 2013) with the BWER displaced from updraft especially early in the lifetime of a severe storm. Quantitatively, analyses show reasonable maximum vertical velocities at 3.5 km also consistent with previous studies (Tanamachi et al., 2012; Markowski et al., 2012a; Calhoun et al., 2013). Because dual-Doppler derived vertical velocities are frequently error prone, one concern with this method (considering one of the datasets was interpolated) was the errors that may arise in analyses of vertical velocity. These results suggest that while the updrafts are qualitatively reasonable, quantitative results from the study must be considered with extreme caution until further work can be performed to assess the magnitude of errors associated with this method. Ultimately, *trends* in the updraft strength with time and height are likely more reliable than the magnitude of each.

Analyses were then performed on the Norman-Shawnee tornado and compared to those seen in the Edmond-Carney case. The Norman-Shawnee tornado was stronger than Edmond-Carney tornado both in damage rating (EF-4 to Edmond-Carney’s EF-3 rating) and in RaXPol observed radial velocities ($> 88 \text{ m s}^{-1}$). Although recorded wind speeds in the Norman-Shawnee tornado were stronger, it is worth noting that the distance of the radar to the tornado at the time of the maximum radial velocity was less than 4 km, allowing the low-level measurements to occur much closer to the surface. Additionally, core radius estimations of the Norman-Shawnee tornado appear smaller at the lowest elevation, but again, estimates were generally taken much closer to the tornado and to the surface.

For a number of reasons, including orientation of the supercell in the dual-Doppler lobe and initial data quality of D3, vertical vorticity and velocity estimates early in the deployment are not particularly trustworthy, but in general show intensification of vertical vorticity with time up to 3 km. Results later in D3 show intensification of the vertical vorticity at low levels while the mid-level vertical vorticity decreased. It is concluded that “shrinking” of vertical vorticity by a downward-directed perturbation pressure gradient force above the level of maximum vorticity likely caused the vorticity in the mid levels to fall while continued low-level stretching intensified the vortex below. Subsequently, as the vorticity maximum aloft descended towards the surface, mid-level vorticity began to increase as buoyancy dominated well above the level of maximum vorticity.

Comparatively, the maximum vertical vorticity observed in the Norman-Shawnee case was higher than was observed in the Edmond-Carney, as expected based on the general intensity of the storm. Although the Norman-Shawnee dual-Doppler analysis did not encompass the storm during its most intense period (D4 *may have* observed the storm at peak intensity, but dual-Doppler was not possible at that time), the relative maximum in vertical vorticity observed between 23:16 UTC

and 23:21 UTC coincides approximately with the time in which EF-3 damage was observed. The analysis also captured the demise and dissipation of the tornado during D5 as it weakened and ultimately lifted.

Vertical profiles of vertical vorticity were nearly qualitatively identical to those in the Edmond-Carney case. As the storm intensified, a vorticity maximum descending towards the ground was observed as the vortex strengthened. As before, this is likely due to both intensification of the vortex near the surface and widening of the vortex as the tornado intensified. Because the horizontal grid spacing used in this case was the same as Edmond-Carney (250 m horizontal grid spacing), the same reasoning applies to the lack of a vorticity maximum near the surface. However, the higher spatial resolution achieved in D4 resulted in ζ increasing towards the surface associated with the strong tornado. The use of D4 data despite not being dual-Doppler capable allowed us to find the missing piece of the puzzle. This technique *did provide* adequate results everywhere where high resolution analyses were not necessarily required. Once the data was supplemented with high-resolution, single-Doppler data, the results matched our expectations.

Analyses of Z_{DR} were performed to observe how changes in the Z_{DR} column were related, if at all, to changes in dual-Doppler derived updrafts. Results from the Edmond-Carney case showed good agreement between maximum Z_{DR} and the most intense updraft regions at 3.5 km, although the maximum Z_{DR} was characterized by a slight downstream displacement similar to those observed in the BWER of the storm. Additionally, the updraft and regions of relatively high Z_{DR} had the same relative shape, further linking their relationship. Similarly, the Norman-Shawnee case showed similar results in that the updrafts were generally collocated with Z_{DR} columns, growing spatially as the updraft intensified.

Estimates of the Z_{DR} column top were examined with respect to the most intense updrafts to investigate if change in updraft intensity altered the maximum

height of the Z_{DR} column. Consistent with previous studies (Snyder et al., 2015), intensification of the updraft maximum was associated with increases in Z_{DR} column height. In the Edmond-Carney case, early intensification followed by a relative steadiness in updraft intensity was characterized by an initial increase in the Z_{DR} column height followed by little-to-no change. In addition to the maximum updraft, a mean updraft was calculated in attempt to mitigate errors, and the results showed little change as well. In the Norman-Shawnee case, intensification of the maximum updraft was characterized by an increase in the Z_{DR} column depth, as expected. Results from the mean updraft, however, revealed a steady increase in updraft intensity throughout virtually the entire analysis period. Although the mean quantity may be a useful measure of updraft size and general intensity, further work must be done to better understand any significance it may have with respect to the Z_{DR} column.

Maximum Z_{DR} at 5 and 6 km was also considered with respect to vertical velocity to determine how it changed with respect to the updraft. In both cases, Z_{DR} increased with time while the updraft in the Norman-Shawnee case increased, but the Edmond-Carney case remained steady. It is hypothesized that increases in Z_{DR} are not necessarily good indicators of storm intensification, but could provide useful information about storm maintenance, maturity, and demise of tornadic supercells.

The focus of this study was to develop a way to combine rapid-scan radar data with additional radar data of generally lesser temporal resolution. This work has shown that not only does this method provide reasonable results, but it allows us to, for the first time, produce dual-Doppler syntheses with much more rapid updates than ever before. This study demonstrates that it is possible to use time interpolated radar data to supplement rapid-scan radar data where a second

radar can be used to provide additional insight on the physical processes within a convective storm.

In order to improve upon the technique, more work must be done to begin to assess the accuracy of the quantitative measurements. To accomplish this, comparisons to traditional dual-Doppler with temporally and spatially-synchronized volumes as a control group would be extremely useful. In the future, perhaps a field project near the ARM site in northern OK could be proposed. This project could make use of multiple mobile radars, as well the Vance Air Force Base WSR-88D near Lamont, OK, to perform traditional dual- or multiple-Doppler analyses and use this technique to interpolate WSR-88D data to the temporal resolution of RaXPoL. In addition, vertical profilers at the ARM site would be useful in determining the true vertical velocities associated with convective storms and comparing them to those obtained in the dual-Doppler analyses.

A project such as this would not only provide useful comparisons to data sets like the one presented in this paper, but it also has the potential to increase our understanding of severe storms. Advances in technology, such as the increase of polarimetric radars, have advanced our understanding of storm microphysics. In concert with advances in phased-array technology, radar data is continuing to become more abundant than ever before. Implementing a technique such as this could give us the ability to use these high-temporal resolution radars to produce rapid-dual- and rapid-multiple-Doppler data sets. Tornadoes can evolve in seconds; thus, improvements in the temporal resolution of analyses used to study these events are vital to better understanding these phenomena.

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Appendix A

Variation Tracking of Rapid-Scan Doppler Radar Data

This thesis focused on a technique for dual-Doppler synthesis capable of combining data with drastically different temporal resolutions. This was accomplished by interpolating data using spatially variable u and v fields, as was discussed in Section 3.3. Separate from the study above which focused on 19 May 2013 data, another case was analyzed using the spatially variable advection correction algorithm to attempt to resolve the horizontal wind field using only a single Doppler radar. Data from 24 May 2011, were used in which RaXPol collected several minutes of single-elevation (1.0°) data on a violent, EF-5 tornado near El Reno, OK (Houser et al., 2015, 2016). Previous studies have explored different methods of obtaining the horizontal wind field from tracking reflectivity elements in supercells using single-Doppler data, with varying degrees of failure (e.g., Kramar et al., 2005; Houser, 2013). This case is unique because high spatial resolution data were available every two seconds, allowing us to attempt to track data with very rapid updates.

Numerous studies have explored this technique in the past, beginning with Rinehart (1979), all with the idea that tracking single-Doppler data could provide a useful means of resolving the horizontal wind field without employing dual- or multiple-Doppler techniques (e.g., Rinehart, 1979; Smythe and Zrnic, 1983; Tuttle and Foote, 1990; Sun et al., 1991; Qiu and Xu, 1992; Laroche and Zawadzki, 1994;

Shapiro et al., 1995; Xu et al., 1995; Zhang and Gal-Chen, 1996; Liou et al., 1999; Liou, 1999; Gao et al., 2001; Liou, 2002; Weygandt et al., 2002; Harasti et al., 2004; Kramar et al., 2005; Houser, 2013). A technique to resolve the 3-D wind field from single-Doppler data was desired because even with the implementation of the WSR-88D network, the 300+ km spacing of individual radars made regular dual-Doppler analyses impractical. In attempt to accomplish this task, two methods of performing single Doppler wind retrievals have been focused on. The first, called the adjoint retrieval method, makes use of prognostic equations in a dynamical model to act as a background to, in essence, guide the wind retrieval (e.g., Laroche and Zawadzki, 1994; Gao et al., 2001). These methods make use of cost function minimization, and the conservation equation for reflectivity, but have the added value of model data to support their results. For example, Gao et al. (2001) used analyzed forecast and background fields to help support their data, to which they were met with some success.

The second type of single-Doppler wind retrieval technique involves variational analysis of reflectivity and radial velocity to obtain the horizontal wind field without the help of model data. Shapiro et al. (1995) created a retrieval using the reflectivity conservation equation, flow incompressibility, and the frozen turbulence hypothesis on the radial velocity field to evaluate the wind field. The least squares method was then used with boundary conditions that could be satisfied based on single-Doppler data. While the study had some success, the authors concluded that many additional cases were needed before the reliability of the procedure could be determined. Zhang and Gal-Chen (1996) performed similar experiments to those of Shapiro et al. (1995), except that they used a reference frame moving at the mean motion of the storms by which the frozen turbulence hypothesis was least violated. A study by Liou (2002) also used a least squares

velocity retrieval method similar to the one done here, which will be discussed in comparison to this dataset.

Although single-Doppler wind retrievals had been attempted in the past, never had it been applied to data with such high spatial and temporal resolution in the vicinity of a strong tornado. RaXPol was located north of the 24 May 2011, EF-5 tornado by roughly 4 km (Fig. A.1) and collected approximately six minutes of single-elevation (1.0°) data. The data were processed were mapped to a Cartesian grid with 75 m by 75 m horizontal grid spacing, and methodology consistent with that listed in Section 3.2.

The spatially variable advection correction technique was then applied to the data as before. However, it was anticipated that with two second updates, the lack of a sufficient decorrelation time (i.e., not enough time for individual reflectivity “blobs” to be advected) may result in random, chaotic data as the data were not trackable on such short time scales. To examine this problem, reflectivity was first tracked using between-scan intervals from two seconds to twenty-two seconds at every two second interval, and the mean storm motion was subtracted to make the horizontal wind field storm relative. For the sake of brevity, only between sweep intervals from two to twelve seconds are included.

At a two second sweep interval (Fig. A.2), the retrieved wind speeds are rather large, with a chaotic wind field resulting in multiple areas. This behavior is especially clear in the northern half of the storm, where attenuation has cause reductions in reflectivity, and due west of the radar where the beam suffers from blocking. Additionally, in the southwestern quadrant of the domain, a lack of data also results in poor data quality. Longer sweep intervals result in more smoothly varying data and the wind field retains the same relative directions at each location in each analysis (Figs. A.3-A.7). Sharp backing of the wind field in the forward

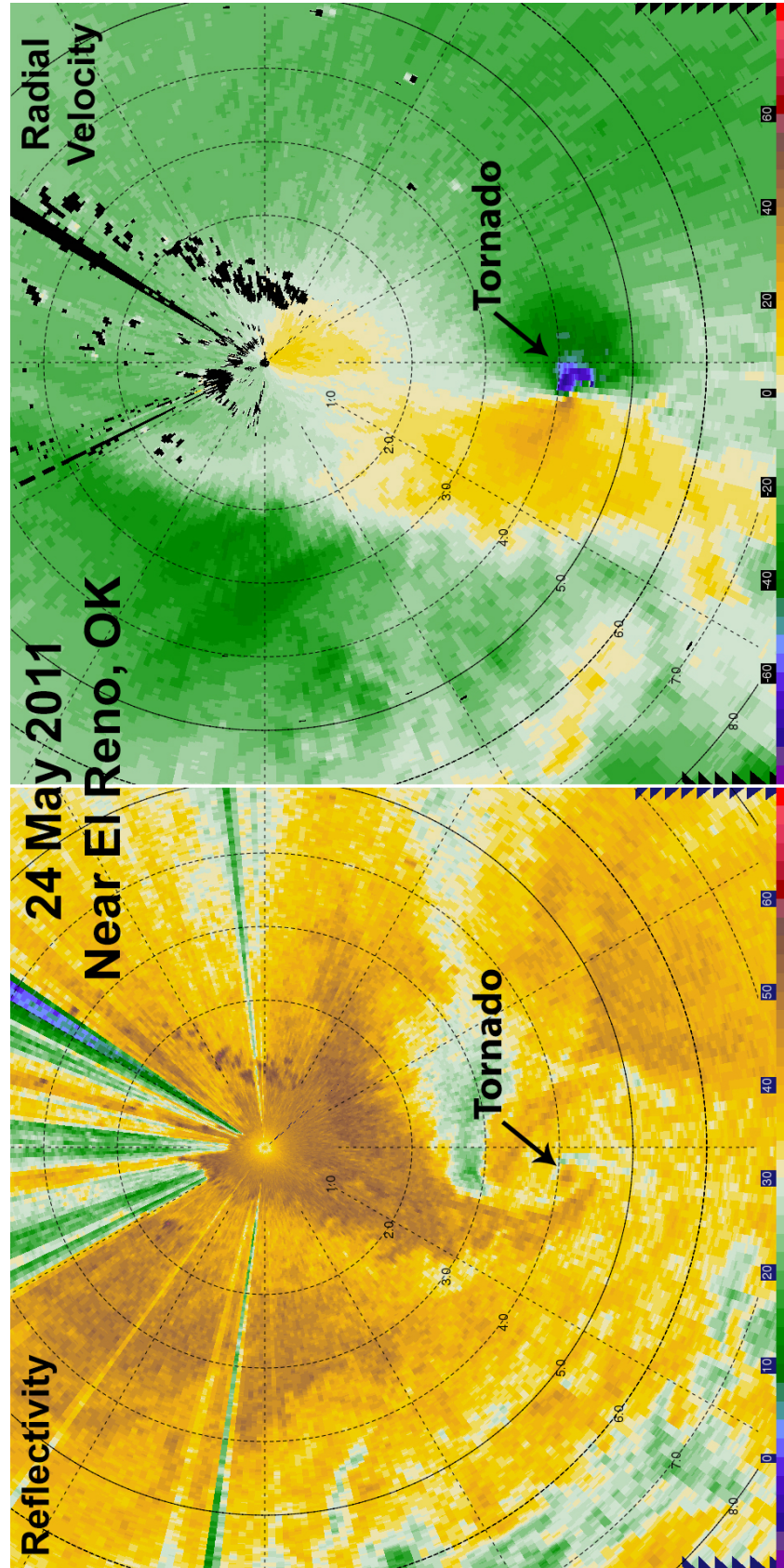


Figure A.1: Reflectivity (left) and radial velocity (right) of the 24 May 2011, El Reno, OK, tornado. RaXPol was located approximately 4 km due north of the tornado at this time.

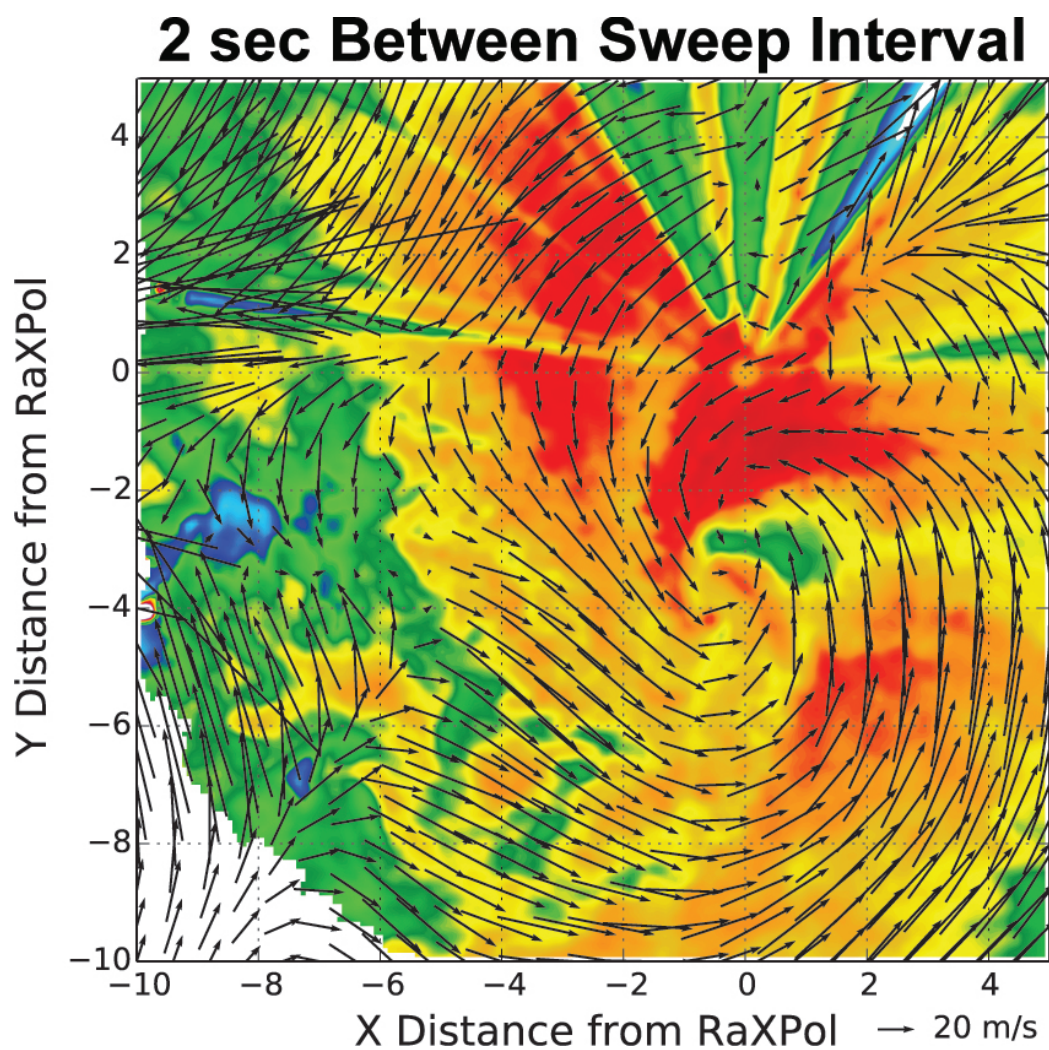


Figure A.2: Analysis of the horizontal wind field at a 2 second sweep interval using the single-Doppler wind retrieval method. RaXPol reflectivity is color contoured, and the horizontal wind field is represented by the black vectors.

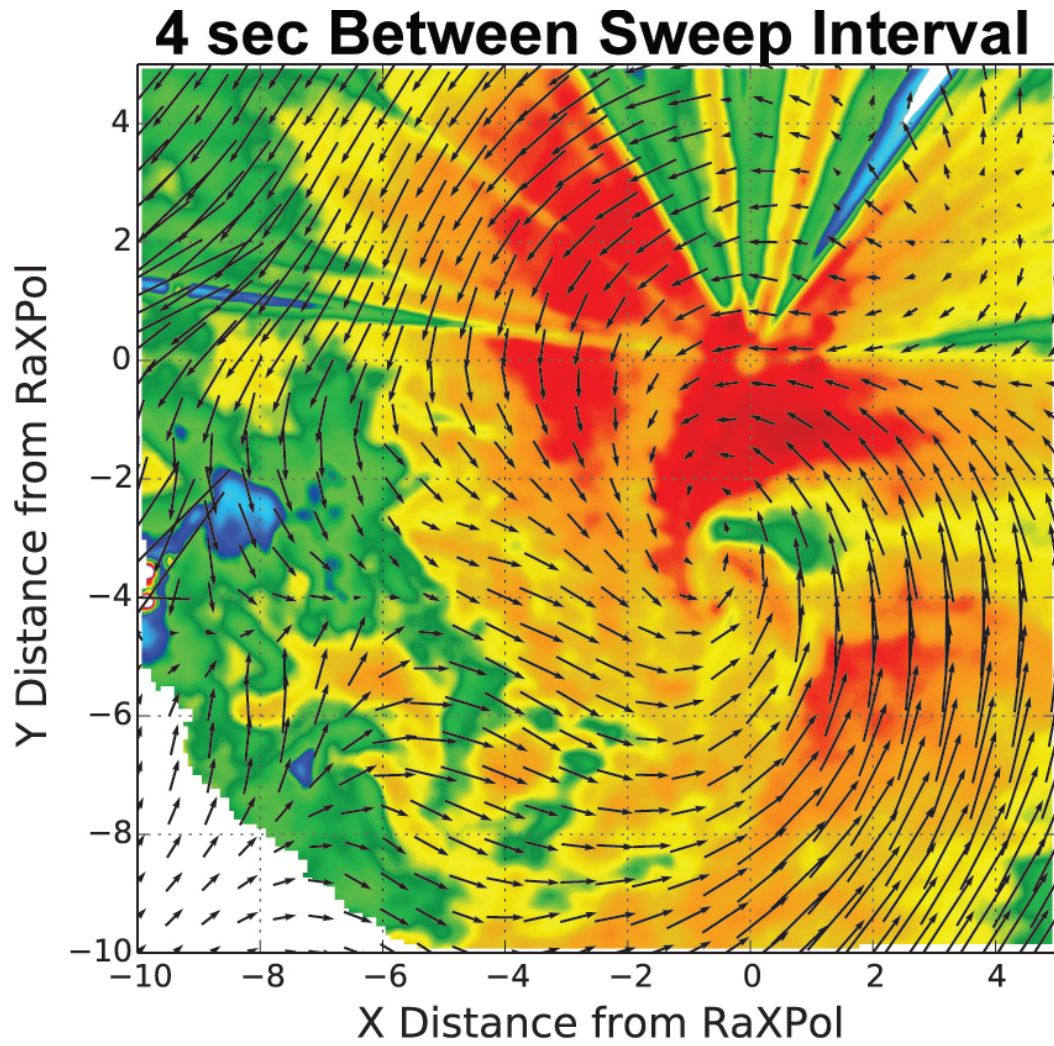


Figure A.3: As in Fig. A.2, but with a 4 second sweep interval.

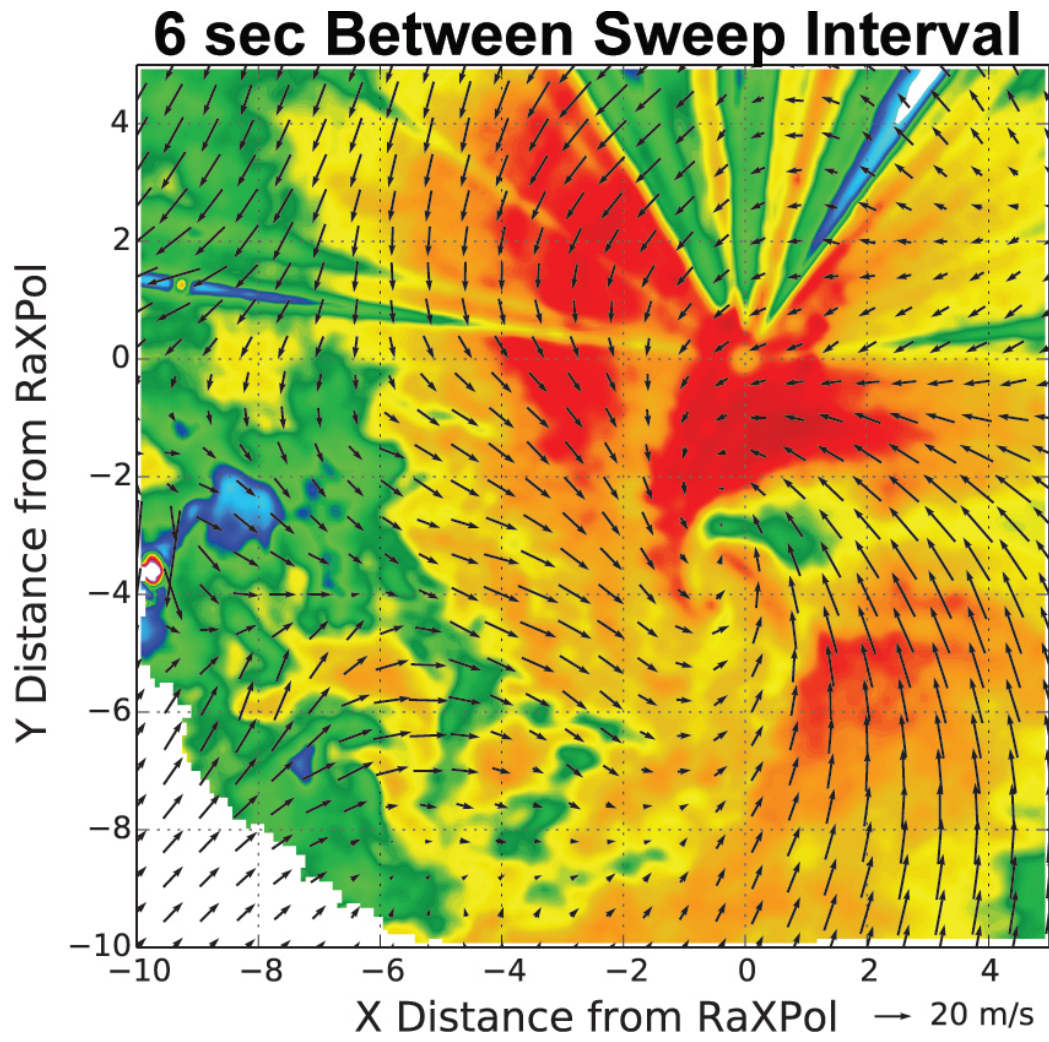


Figure A.4: As in Fig. A.2, but with a 6 second sweep interval.

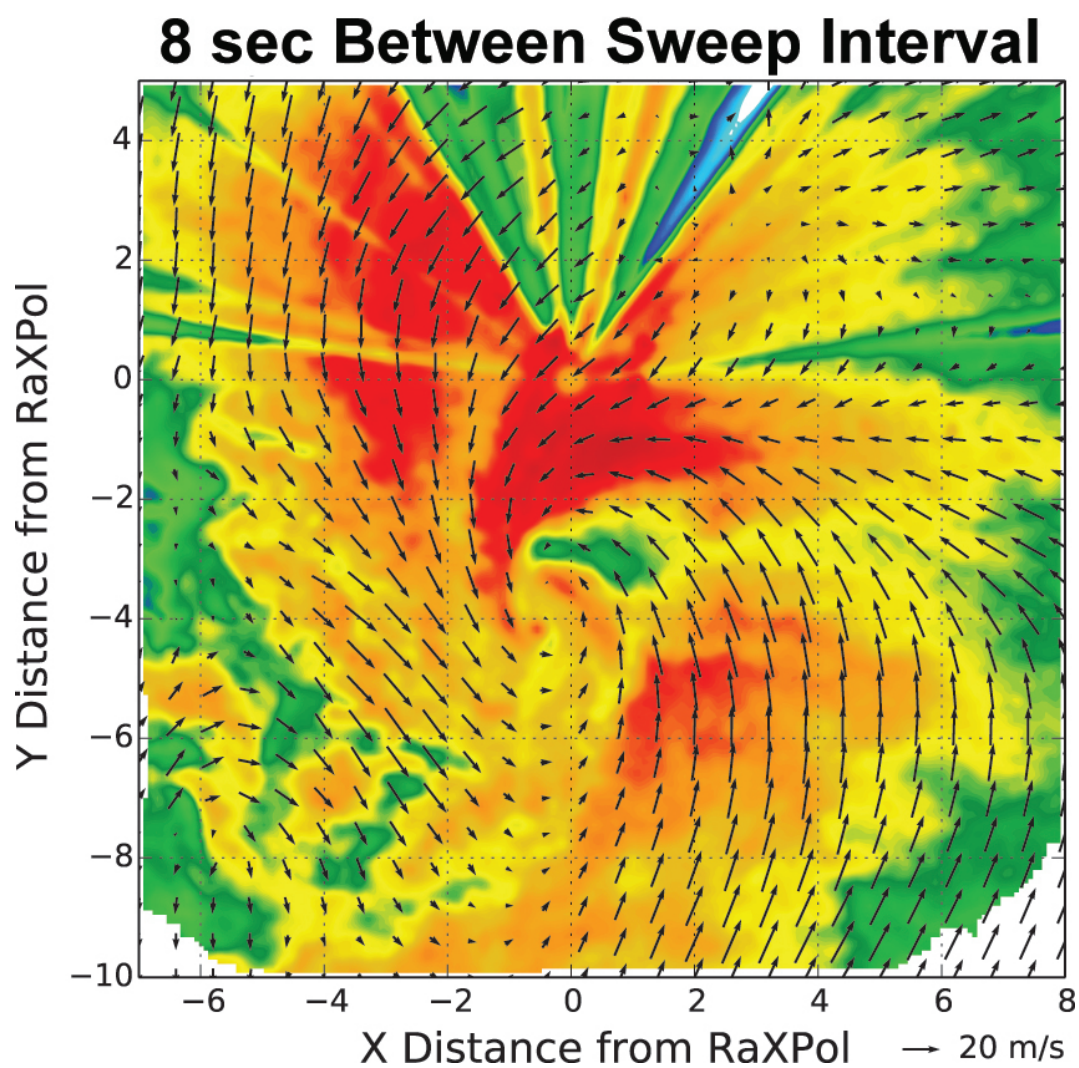


Figure A.5: As in Fig. A.2, but with a 8 second sweep interval.

10 sec Between Sweep Interval

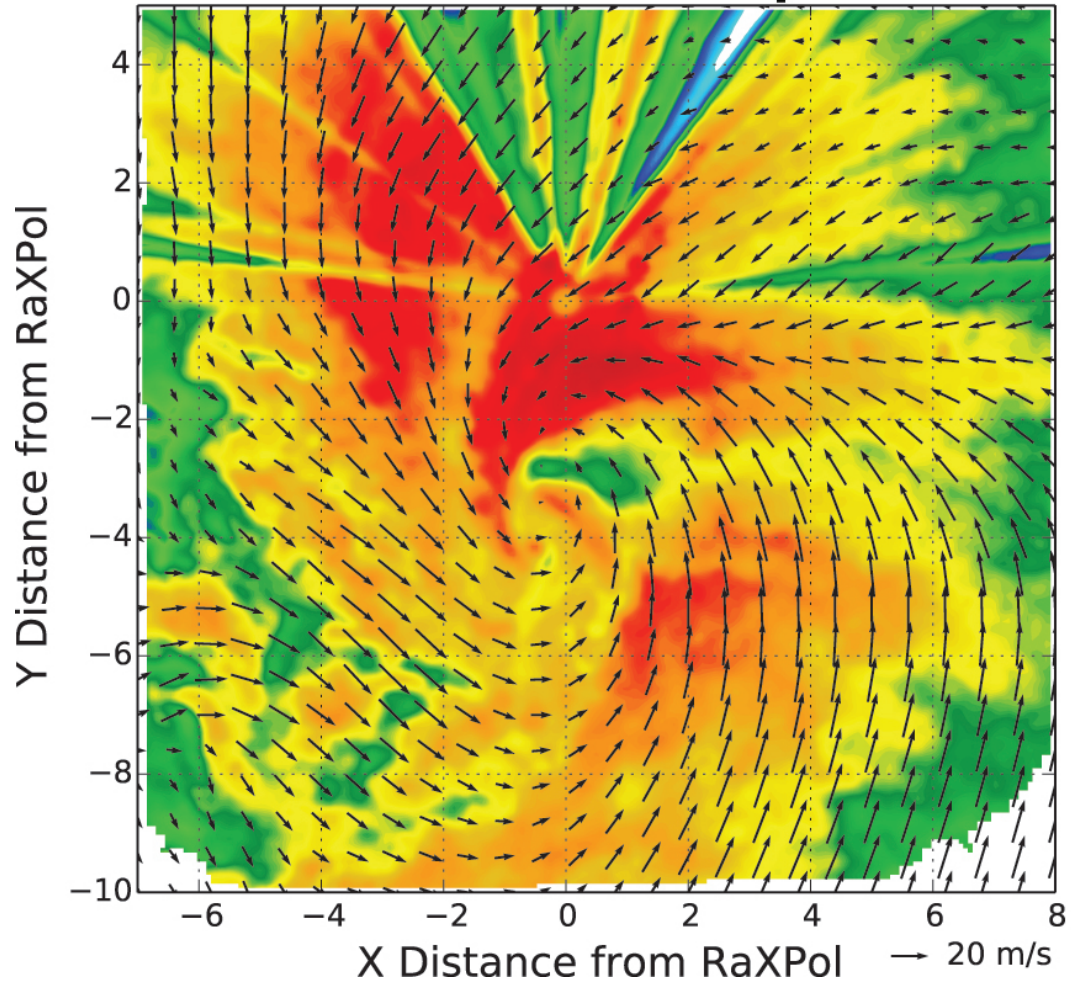


Figure A.6: As in Fig. A.2, but with a 10 second sweep interval.

12 sec Between Sweep Interval

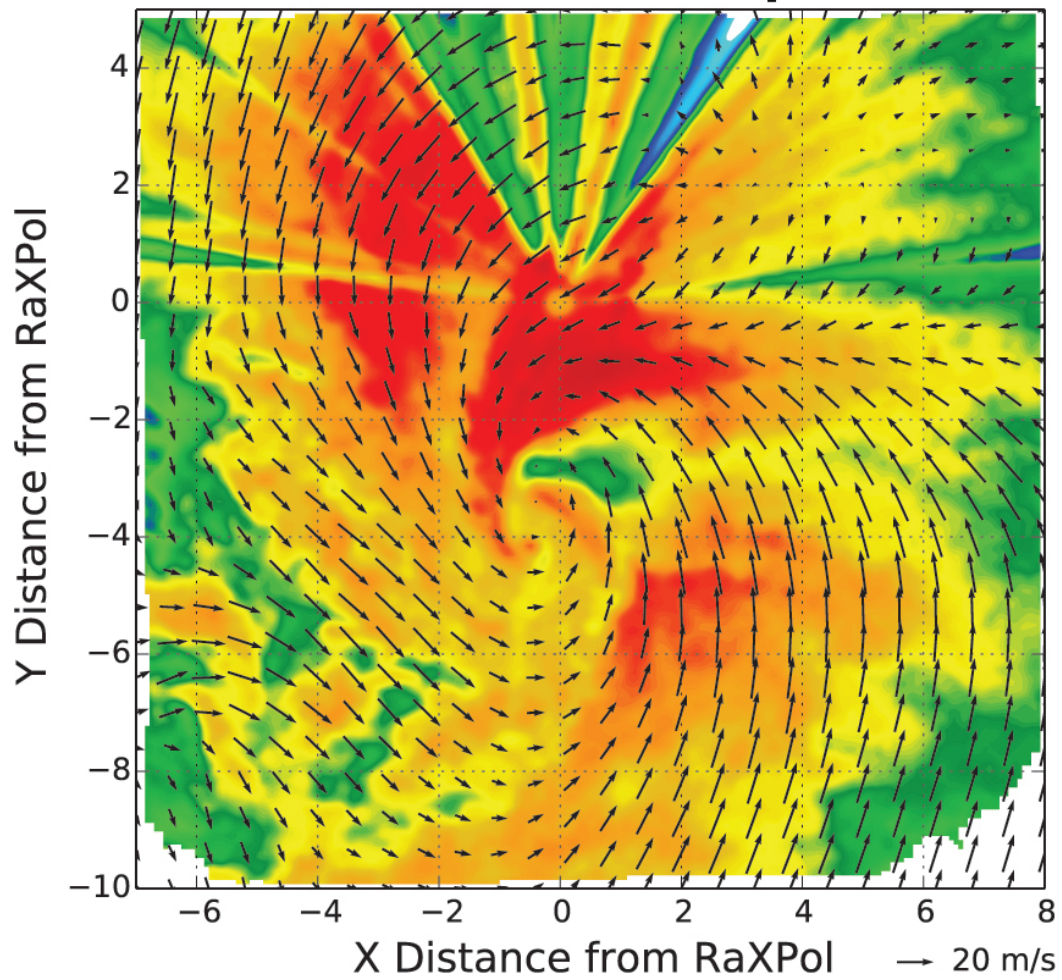


Figure A.7: As in Fig. A.2, but with a 12 second sweep interval.

flank is consistent with a confluence zone observed in dual-Doppler observations (see Fig. 4.4), and slowing and backing winds south of the hook echo are consistent with flow characteristics along the rear-flank gust front. One major concern in all six analyses, though, is the lack of any significant wind speeds near the tornado. The general wind field in each case shows rotation around the tornado; however, the horizontal motion around the tornado, and more specifically the magnitudes of the wind field, are not realized in these analyses. It can be seen in a timeseries of the maximum and mean winds that as the between sweep interval was increased, from two seconds to twenty-two seconds, the mean and maximum wind speeds given by the retrieval converge (Figs. A.8 and A.9). The convergence of the wind speeds beginning at approximately 8-10 second sweep intervals argues that there is value in utilizing this length of time between scans. More importantly, the algorithm appears to converge to roughly the mean storm motion at this time. With major errors evident in the near-tornado wind speeds, further investigation into the cause is needed.

A number of issues could be in play causing large discrepancies in the observed wind field compared to expectations. As described by Liou (2002), the azimuthal wind component is often underestimated. In this case, the region where we would expect the tornado to have an eastward component of motion (north of the vortex) has little to no wind speed at all. Not only is it underestimated, but it is essentially non-existent. Given that this is a violent tornado, we would expect some substantial eastward component even if the technique is known to underestimate the wind field. Another potential reason for the weak winds near the tornado is that perhaps the tornado is *too* strong. One issue with this technique and rapid-scan data is the potential for temporal aliasing in the data. Given that the algorithm works by essentially tracking “blobs” of reflectivity, we must consider how far a blob will

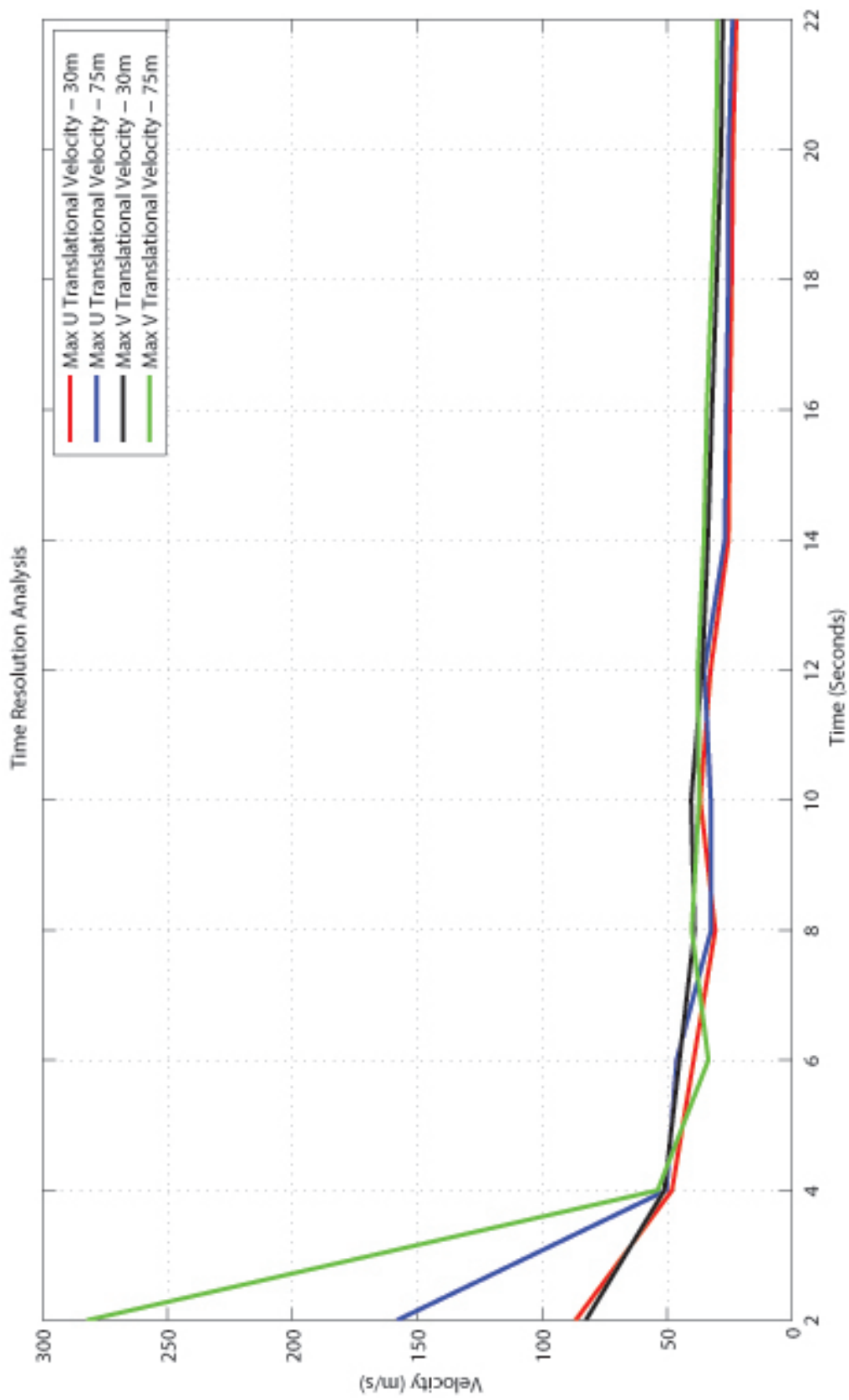


Figure A.8: Domain-calculated maximum U and V retrieved wind speeds for varying sweep intervals at 30 m and 75 m horizontal grid spacing. Maximum velocities are in $m.s^{-1}$.

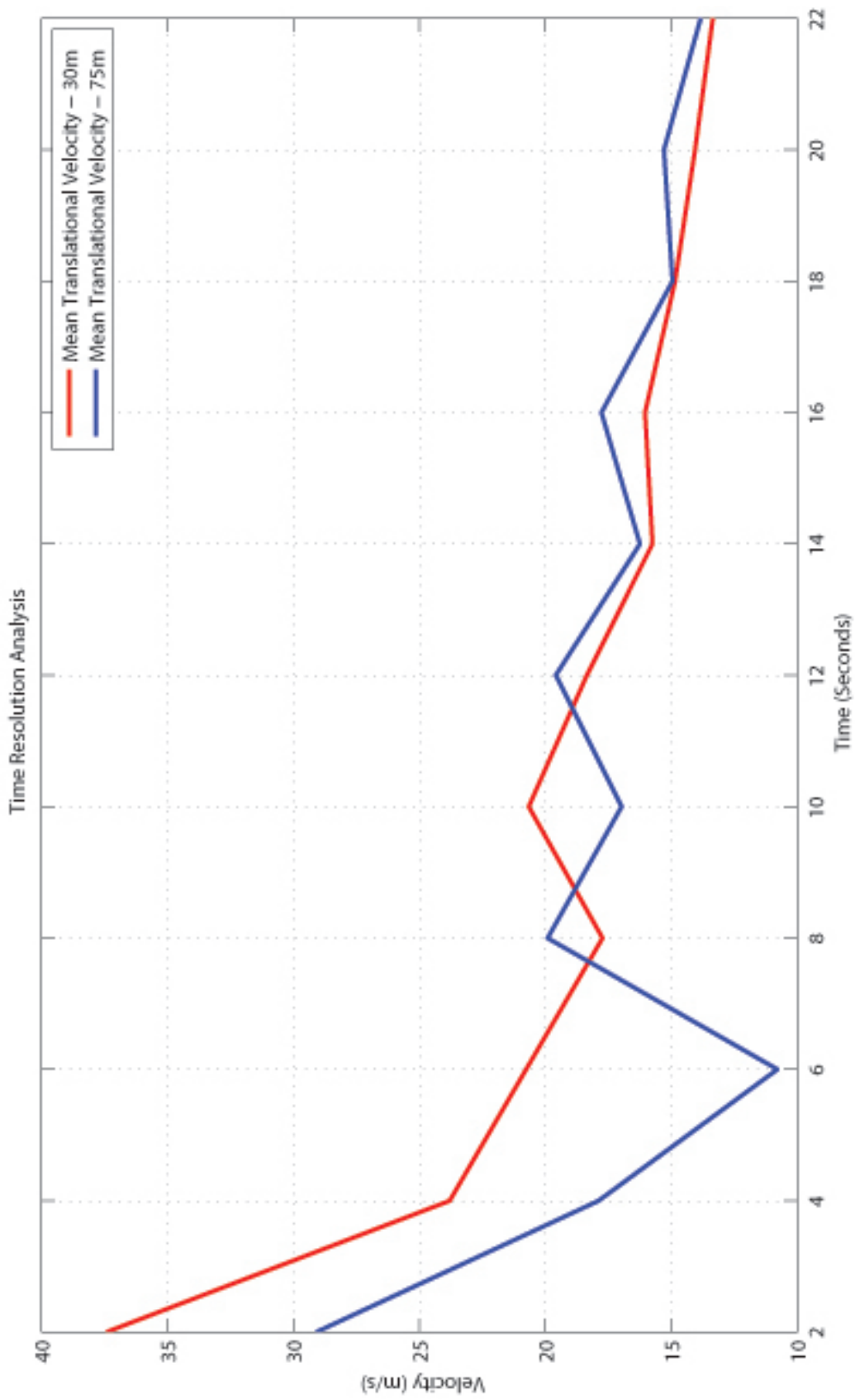


Figure A.9: Domain-averaged retrieved wind speeds calculated at 30 m and 75 m horizontal grid spacing. Maximum velocities are in $m s^{-1}$.

move in a period of time and match it with the proper scanning interval. Recall that, in this case, the grid spacing in the horizontal is 75 m. An average order of magnitude for the wind speed around this supercell is 10 m s^{-1} . At that speed, it would take 7.5 s for a blob to move a full grid point. If the scanning interval is too short, the blob will not be advected a full grid point, resulting in a non-unique solution. This is a likely explanation for some of the noisy analyses associated with a short sweep interval. However, the issue in this case is that 10 m s^{-1} is not a sufficient order of magnitude to satisfy the whole storm. Associated with the strong tornado are wind speeds closer to an order of magnitude of 100 m s^{-1} . In this region, an update time of 8-10 seconds means that a blob will be advected nearly a kilometer. This posed a major issue for this analysis because choosing only one sweep interval was not capable of satisfying all of the data. However, even at 2 seconds, the wind speeds near the tornado remained drastically underestimated, suggesting that they were also not useful.

Efforts were made to utilize the polarimetric variables ρ_{HV} and Z_{DR} as well, in particular as an attempt to track the tornado. If the use of polarimetric variables could help track the velocities in and near the tornado, then perhaps a composite of tracked variables could be combined in a case like this. Tracking Z_{DR} turned out to be useless as the fields were far too noisy to produce any useful results. The tracking of ρ_{HV} looked more promising, especially when restricted only to $\rho_{HV} < 0.8$. With temporal aliasing concerns, attempts were made to produce useful results again using two and ten second sweep intervals (Fig. A.10). Once again, the lack of any appreciable tracer near the tornado resulted in the lack of realistic results near the tornado. First attempts to produce a reasonable wind field, qualitatively or quantitatively, failed likely due to the inability of the algorithm to track blobs around the tornado. The only hope for every being able to resolve a vortex like this

would be to drastically increase the resolution, which means moving substantially closer than the 4 km range in this case. While this may produce better results, it is unclear how much better the results could be, and logistically this strategy would be difficult to safely accomplish.

Once the velocities were tracked, and despite the proper retrieval near the tornado, wind speeds were then broken down into radial and azimuthal components with the true, RaXPol obtained radial velocities inserted into the data as the radial component of motion (Fig. A.11). The idea was that if the tracked velocities could serve as the azimuthal component of the wind, the radial velocity would act as the truth, increasing the accuracy of the analysis. Certain areas of the analysis, including the rear flank and inflow regions of the storm, produced qualitatively and quantitatively reasonable results. However, the same issues remain near the tornado. Strong wind speeds on the eastern and western periphery of the tornado indeed indicate that a strong tornado is ongoing; however, the northern and southern edges of the vortex, where the wind is perpendicular (only azimuthal component) to the beam, are very small. A noticeable kink in the wind field occurs on each side of the vortex. Once again, the algorithm fails to produce reasonable results surrounding the tornado.

As the algorithm essentially tracks reflectivity blobs, consider that the debris ball essentially acts as a solid object in this case with no individual tracers. Instead of tracking the flow around the vortex, it is simply tracking the vortex itself. Although we were unsuccessful in tracking the wind field around the tornado in this case, the technique was successful in tracking the motion of the storm over multiple periods of time. Tornado motion is generally understood to be very non-linear, which makes tracking features in a tornado-relative sense difficult. This technique could be applied to algorithms to better calculate the motion of the tornado itself,

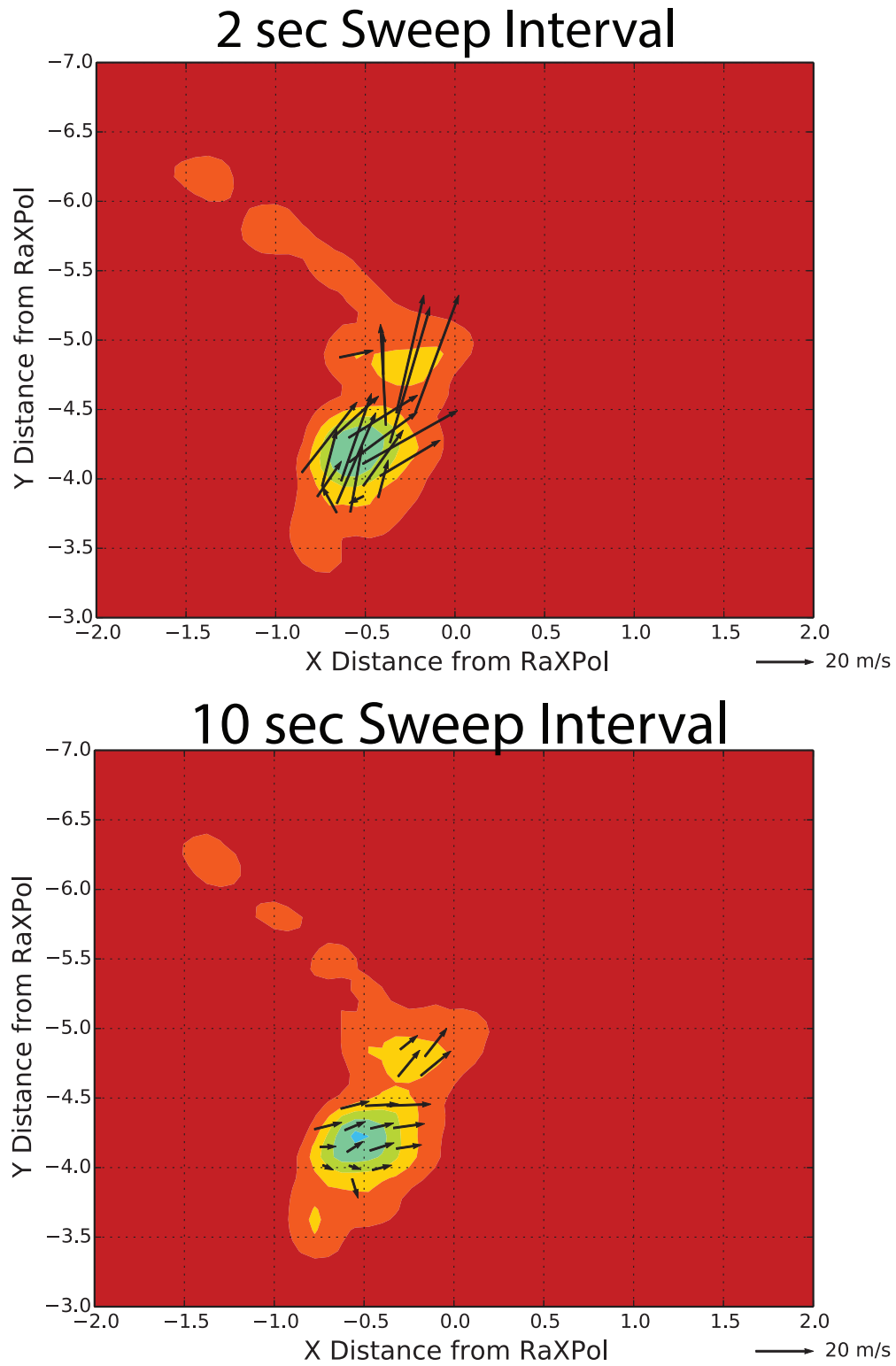


Figure A.10: Single-Doppler retrieval applied to $\rho_{HV} < 0.8$. Color contoured is RaXPol ρ_{HV} , with the horizontal wind field in and near the tornado represented by black vectors.

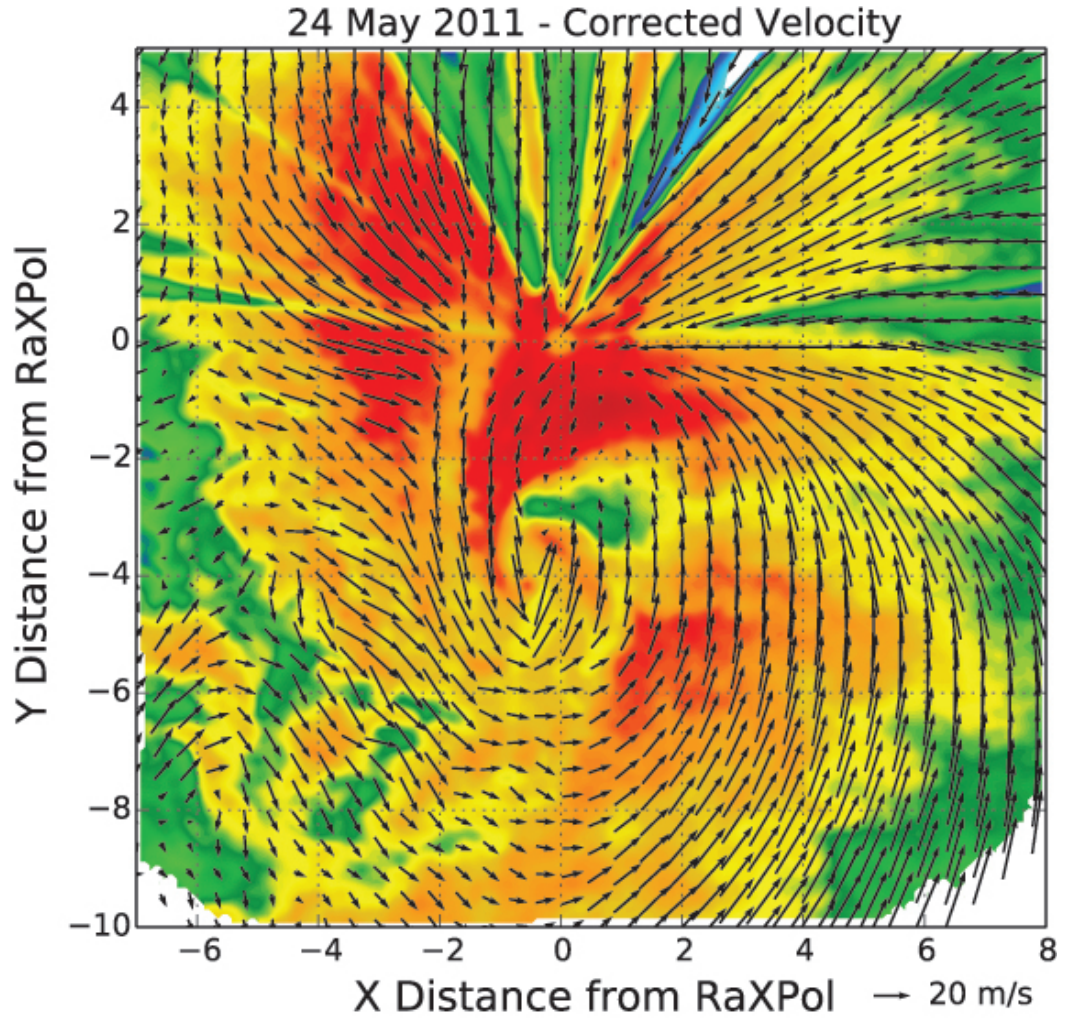


Figure A.11: As in Fig. A.2, but with 10 second scan interval and the RaXPol obtained radial velocity inserted as the radial wind component.

or even just the storm. A second valuable finding, one that has been discussed many times before, is the need for higher resolution data. This may be logistically difficult in the current state of radar meteorology, but it provides a goal for the future. Techniques like this may one day be more successful if we are able to achieve even higher resolution. Finally, with increased implementation of phased-array radars, this study provides some evidence that two second updates at a single elevation may not be necessary. Because radar data is a series of discrete points, there will always be errors associated with finite differencing schemes employed to analyze the data. Furthermore, because the data points are discrete, temporal aliasing will always be an issue if we do not give the data time to decorrelate sufficiently. For this reason, it is important that data scans be spaced long enough apart relative to the horizontal grid spacing to accomplish this task, similar to those findings by Liou (2002). In the case of RaXPol, it makes sense that sweeps be done at multiple levels to not only collect data higher in the storm, but to also give the lower levels time to change. For phased-array radars, scanning rates of 8-12 seconds would be the most beneficial, allowing the radar to collect quality data at multiple elevations without the worry of collecting so much data in such rapid updates.