THE MAINTENANCE OF TORNADOES OBSERVED WITH
HIGH-RESOLUTION MOBILE DOPPLER RADARS

A Dissertation in
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by
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High-resolution dual-Doppler wind syntheses and ensemble Kalman filter data assimilation experiments performed using Doppler on Wheels velocity data collected in four tornadic supercells are examined in order to determine how the evolution of storm structure affects the maintenance of tornadoes. Tornado maintenance can occur as long as the tornado remains connected to the horizontal convergence located along the rear-flank gust front. The tornadoes dissipate when this connection is severed. The strongest and longest-lived tornado examined in this study maintains a connection to the primary rear-flank gust front for a considerable time. The weakest and shortest-lived tornado maintains only brief contact with the gust front. Rather than perishing while completely surrounded by outflow air (as is shown in past conceptual models of tornadic supercell storms), the two tornadoes with a moderate intensity and duration are sustained by convergence and vertical vorticity provided by secondary rear-flank gust fronts. It is speculated that these tornadoes do not achieve a greater intensity or a longer duration because of the transient nature of the secondary gust fronts to which they are connected.

The strongest and longest-lived tornado dissipates when it becomes far removed from the mid-level updraft. At this time, the mid-level updraft is fueled by near-surface inflow that rises straight upward. However, the weakest and shortest-lived tornado dissipates beneath the mid-level updraft that primarily ingests near-surface inflow located several kilometers to its east. This observation suggests that having a tornado located beneath the primary updraft is not always a sufficient
condition for maintenance, particularly in the presence of strongly surging outflow winds.

The three-dimensional structure of vortex lines along the rear-flank gust fronts is consistent with the tilting of baroclinically-generated horizontal vortex rings surrounding a negatively buoyant downdraft. When the tornadoes are completely surrounded by outflow air, tilting along the primary rear-flank gust front does not generate cyclonic vertical vorticity at the tornado. Instead, tilting of the horizontal vortex lines by the secondary gust fronts more directly supplies the mature tornadoes with low-level rotation. This observation suggests that relatively cold secondary surges of outflow may assist with tornado maintenance despite a greater negative buoyancy. Thermodynamic data retrieved in the two data assimilation cases suggest that a warming of the outflow appears to be symptomatic of changes in the rear-flank downdraft that adversely affect tornado maintenance.
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Introduction and Motivation

Despite their obvious societal importance, little is known about the dynamics governing several aspects of tornado behavior, particularly their maintenance. This is due to the dangers and logistical complexity involved with data collection and the computer power required to simulate both the parent storm and the small-scale tornado. Despite these difficulties, several studies have deduced information about conditions supporting tornadogenesis and/or tornado maintenance by analyzing storm-scale properties of supercells using coarse-resolution numerical simulations or radar observations. For example, Browning (1965) and Brandes (1977, 1978, 1981, 1984) showed the position of tornadoes relative to other features of their parent supercell storms, such as updrafts, downdrafts, precipitation regions, gust fronts, and the low-level mesocyclone, throughout the evolution of the storm using single- and dual-Doppler radar data. Other studies have used idealized numerical simulations in order to study the importance of a rear-flank downdraft (e.g., Davies-Jones and Brooks 1993, Straka et al. 2007), low-level baroclinity (e.g., Rotunno and Klemp 1985, Trapp and Fiedler 1995, Wicker and Wilhelmson 1995), the orientation of low-level wind shear (Wicker 1996), and dynamically-induced
vertical perturbation pressure gradient forces (Klemp and Rotunno 1983, Wicker and Wilhelmson 1995, Trapp and Davies-Jones 1997) on the evolution of a low-level mesocyclone, and possibly also a tornado. More recent studies have analyzed the possible relationships between tornadogenesis and

1. the thermodynamic characteristics of the outflow air (Markowski et al. 2002, Markowski et al. 2003, Shabbott and Markowski 2006, Grzych et al. 2007, Majcen 2009),

2. specific processes resulting in the formation and subsequent tilting or stretching of three-dimensional vortex lines (Straka et al. 2007, Markowski et al. 2008),

3. the presence of horizontal shear vortices along gust fronts (Finley et al. 2002, Lee and Wilhelmson 1997),

4. cyclic mesocyclogenesis (Dowell and Bluestein 2002a),

5. the transition of a one-celled mesocyclone into a two-celled structure (Wakimoto and Liu 1998),

6. descending radar reflectivity cores (Rasmussen et al. 2006, Byko et al. 2009),

7. and corner-flow collapse of the low-level mesocyclone or tornado-cyclone (Lewellen and Lewellen 2007a,b).

These and other studies have resulted in a vast advancement of our understanding of tornado behavior over the last several decades. A comprehensive review of tornado-related literature suggests a strong focus on the understanding of processes involved with the formation of tornadoes; informally tallied, at least a dozen different, albeit not always unrelated, hypotheses spanning dozens of studies have been posed for methods of tornadogenesis (Davies-Jones 2006). In comparison, a relative paucity of hypotheses exist in the literature regarding what processes cause certain tornadoes to persist for longer periods of time than others. In fact, the author is aware of only one study posing a hypothesis that predicts the conditions for tornado maintenance (Dowell and Bluestein 2002b). A brief description of tornado demise is offered in a few studies, in which it is speculated that cold outflow air wraps around the low-level circulation, which weakens the updraft aloft (because
it ingests negatively buoyant air) and subsequently precludes the contraction of angular momentum near the ground (e.g., Lemon and Doswell 1979, Wurman et al. 2007). These simplified accounts of tornado dissipation lack details regarding the transition from a steady to a weakening mesocyclone and/or tornado, which would likely prove valuable for the assessment of tornado maintenance within a storm. It has been shown in a recent study by Marquis et al. (2008) that, in some cases, a tornado may persist for a long period of time despite being completely surrounded by outflow air, suggesting a more complex structure in the rear-flank downdraft region of a supercell than exists in the simplified accounts of a decaying supercell. Indeed, Marquis et al. and other studies (Adlerman 2003, Finley and Lee 2004, Lee and Finley 2004, Hirth et al. 2008, Wurman et al. 2007, Lee et al. 2008, Finley and Lee 2008), have shown evidence of horizontal winds shifts and sometimes horizontal thermodynamic gradients leading what are assumed to be discrete secondary pulses of rear-flank downdraft. Having only been documented in recent studies, the role that these secondary surges of rear-flank downdraft and their attendant gust fronts play in tornado maintenance is not understood.

Over the last decade, the Doppler on Wheels (DOW; Wurman et al. 1997) radars have collected fine-scale resolution data in close proximity to many tornadoes. Some of the storms were observed with two DOW radars simultaneously, providing the highest temporal (50–90 second) and spatial (100–300 m) resolution dual-Doppler wind analyses surrounding tornadoes ever achieved in the lowest few kilometers of a storm. In this study, single- and dual-Doppler data collected by the DOW radars in four tornadic supercells are analyzed in order to assess the influence of the mesocyclone-scale flow on the maintenance of tornadoes. The objective analysis needed to produce mesocyclone-scale fields necessarily degrades our ability to accurately assess the intensity of the tornado itself. For this reason, the unsmoothed single-Doppler data are used to document the resolved structure and intensity of each tornado and to relate their trends to the evolution of the mesocyclone-scale flow. Single-Doppler data from two cases are assimilated into a numerical cloud model using the ensemble Kalman filter method (e.g., Snyder and Zhang 2003) in order to expand the availability of three-dimensional wind fields and to provide estimates of unobserved thermodynamic variables surrounding the tornadoes. The resulting combination of dual-Doppler wind syntheses and
simulated cases permits a detailed assessment of the roles that mesocyclone-scale processes play in tornado maintenance, which may not have been possible in past studies. The characteristics of the mesocyclone-scale flow that are examined in our four cases include

1. the structure of the rear-flank gust fronts and their motion relative to the tornadoes,
2. the intensity and evolution of the rear-flank downdraft,
3. the relative magnitudes of terms in the vertical vorticity tendency equation surrounding the tornadoes,
4. the temperature of the air surrounding the tornadoes (retrieved through data assimilation),
5. and the structure of baroclinically-generated vortex lines surrounding the tornadoes and outflow air,

all of which have been hypothesized to be important to tornado formation and/or demise in past studies. A comparison of these topics among the four storms, along with comparisons with others described in the past literature, allows us to determine which aspects of tornado maintenance are apparently common to many supercells and which are unique to particular storms.

Chapter 2 provides a brief review of the past literature that is relevant to various aspects of supercell and tornado behavior described in this study. Chapter 3 outlines the dual-Doppler and data assimilation methods employed in the retrieval of the 3D wind fields used in the analysis. Chapter 4 documents the evolution of the single-Doppler-resolved structure and intensity of the tornadoes observed in each case. All of the data are analyzed in chapter 5, and conclusions and suggestions for future work are provided in chapter 6.
A tornado is a violently rotating column of air in contact with both the ground and the base of a convective cloud. Given this definition, many vortices can be considered tornadoes, including those occurring beneath non-rotating updrafts along a wind shift associated with an atmospheric boundary (e.g., cold front, thunderstorm gust front, etc.). These vortices are commonly called non-supercell tornadoes (e.g., Wakimoto and Wilson 1989 and Lee and Wilhelmson 1997). However, the most intense, longest-lived, and most damaging tornadoes typically occur beneath the rotating updraft of a supercell storm. For this reason, tornadoes occurring within supercell storms are the focus of this study. In this chapter, we describe the observed structure and evolution of typical supercell storms and provide a review of past hypotheses regarding tornado formation and dissipation that highlights the uncertainty in the link between supercell storm evolution and the maintenance of a tornado.
2.1 Structure of a typical supercell storm

A supercell thunderstorm differs from more common (and generally less severe) thunderstorms in that the positively buoyant supercell updraft contains significant rotation over much of its depth. The origin of the mesocyclone-scale rotation at middle to upper levels of the updraft is well understood. The evolution of vertical vorticity in an Eulerian reference frame is governed by,

$$\frac{\partial \zeta}{\partial t} = - \left( u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} \right) - w \frac{\partial \zeta}{\partial z} - \left( \frac{\partial \alpha \partial p}{\partial x \partial z} - \frac{\partial \alpha \partial p}{\partial y \partial z} \right) - \left( \frac{\partial w \partial v}{\partial x \partial z} - \frac{\partial w \partial u}{\partial y \partial z} \right)$$

$$- (\zeta + f) \delta - v \frac{\partial f}{\partial y} + \left( \frac{\partial F}{\partial x} - \frac{\partial F}{\partial y} \right),$$

(2.1)

where $\vec{v} = (u, v, w)$ is the three-dimensional wind velocity vector, $\alpha$ is the specific volume, $p$ is the atmospheric pressure, $f$ is the Earth’s vertical vorticity (Coriolis factor), $\delta$ is the divergence of the horizontal wind, and $F$ is friction. The terms on the rhs are as follows (from left to right): horizontal and vertical advection of vertical vorticity, solenoidal generation, tilting of horizontal vorticity into the vertical, stretching of existing absolute vertical vorticity, the horizontal advection of Earth’s vorticity, and frictional effects. The advection and stretching of $f$ in equation 2.1 typically are less significant to the production of vertical vorticity in supercells than the remaining terms ($\partial f/\partial y \sim 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$ at mid-latitudes, therefore $v \frac{\partial f}{\partial y} \sim 10^{-10} \text{ s}^{-2} << \text{mesocyclone- or tornado-scale vorticity} \sim 10^{-2} - 10^0 \text{ s}^{-1}; f \sim 10^{-4} \text{ s}^{-1} << 10^{-2} - 10^0 \text{ s}^{-1}$, therefore, the stretching of Earth’s vorticity is insignificant on time scales less than about 3 hours). Frictional effects also are assumed to be insignificant away from the surface and the solenoidal term typically is small (Heymsfield 1978). Vertical shear of the environmental horizontal wind is tilted into the vertical by the initially non-rotating updraft, generating a pair of counter rotating vortices that straddle it (Wilhelmson and Klemp 1978, Klemp and Wilhelmson 1978, Rotunno 1981, Davies-Jones 1984). In the case of streamwise environmental vorticity, storm-relative horizontal winds advect these vorticity maxima relative to the updraft such that net rotation is achieved (Davies-Jones 1984).

Figure 2.1 illustrates the schematic structure of the precipitation and kinematic
fields that typify a supercell storm. Downstream (with respect to the storm-relative wind) of the rotating updraft is a large area of rainy downdraft that is produced by a combination of the evaporation and melting of the primary rain and hail core and through precipitation loading. As a result, near-surface air within the forward flank region of the storm is cooler than in the ambient environment, and an air mass boundary called the forward-flank gust front defines its leading edge. The near-surface horizontal vorticity vector, which is baroclinically-generated within the temperature gradient along the forward-flank frontal zone, points toward the updraft. A hook-like appendage in radar reflectivity (typically called a “hook echo”) often is seen on the right rear flank of the storm. This appendage is interpreted in many studies to be a tracer of hail and rain drawn away from the main precipitation core to the rear of the updraft by the mesocyclone (e.g., Fujita 1958, Browning 1965), although, in some storms, the hook echo could be a result of precipitation that falls from aloft on the rear flank of the updraft (e.g., Forbes 1981, Rasmussen et al. 2006, Byko et al. 2009).

Following the production of a rotating updraft and forward-flank downdraft, a
second downdraft forms on the rear flank of the storm. The formation mechanism of this downdraft is not well understood. However, because it is composed of negatively buoyant air, evaporative cooling of rain and cloud water on the rear flank of the updraft by intruding dry mid-level environmental air is assumed to play a role (e.g., Browning and Ludlam 1962, Browning and Donaldson 1963, Browning 1964, Nelson 1977, Brandes 1981, 1984, Klemp et al. 1981, Markowski et al. 2002). Lemon and Doswell provide a schematic of the airflow involved with this manner of rear-flank downdraft production (Fig 2.2), although certain aspects of the flow included in their diagram are not completely understood, such as the depth of the rear-flank downdraft. Other possible factors leading to the formation of downdraft include dynamically-induced vertical perturbation pressure gradients, described in Markowski et al. (2002) for Boussenesq flow,

\[
\frac{\partial p'}{\partial z} \propto \frac{\partial}{\partial z} \left[ \left( \frac{\partial u'}{\partial x} \right)^2 + \left( \frac{\partial v'}{\partial y} \right)^2 + \left( \frac{\partial w'}{\partial z} \right)^2 \right] - \frac{1}{2} \frac{\partial}{\partial z} \left[ \zeta'^2 \right] \\
+ 2 \frac{\partial}{\partial z} \left[ \frac{\partial \vec{v}_h}{\partial z} \cdot \nabla w' \right] - \frac{\partial}{\partial z} \left[ \frac{\partial B}{\partial z} \right],
\]

where \( \partial \vec{v}_h / \partial z \) is the vertical shear of the horizontal wind and \( B = -g \frac{\rho'}{\rho} \) is the buoyancy. This equation is developed using the linearized equations of motion, where the overbar denotes mean or environmental variables, and \( ' \) indicates departures from the mean. The rhs of equation 2.2 are vertical gradients of (from left to right): the fluid extension terms, the perturbation vertical vorticity, the alignment of the mean shear vector with the horizontal gradient of updraft, and the vertical gradient of the buoyancy acceleration.

Several studies (e.g., Klemp and Rotunno 1983, Wicker and Wilhelmson 1995, Wakimoto and Liu 1998) have shown a local intensification of the downdraft on the rear flank of the storm close to the center of rotation. The placement of this downdraft relative to other storm features is illustrated in Fig 2.3. Klemp and Rotunno (1983) called this the ‘occlusion downdraft’ and attributed it to dynamically driven vertical perturbation pressure gradients; perturbation pressure within the strong low-level vertical vorticity is lower than that within the less intense mesocyclone aloft, promoting the downward acceleration of air by the second term
on the rhs of equation 2.1. The shear-updraft interaction term in equation 2.2 could contribute to the downward forcing of air in the rear-flank downdraft if the hodograph is curved, such that \( \frac{\partial}{\partial z} \left[ \frac{\partial \bar{\mathbf{v}}_h}{\partial z} \cdot \nabla_h w' \right] > 0 \) on the rear flank of the storm.

At the ground, the cool outflow air originating from within the rear-flank downdraft is led by a line of low-level horizontal convergence, vertical motion, and horizontal temperature gradient, called a rear-flank gust front. Several studies (e.g., Lemon and Doswell 1979, Brandes 1984) have noted that vertical vorticity at low levels dramatically increases after the rear-flank downdraft reaches the ground. This observation suggests that this downdraft is instrumental to the production of low-level rotation; however, the exact processes involved with the generation of this vorticity are not well known. Several hypotheses from past studies regarding the production of low-level vertical vorticity are discussed in section 2.3 below.

As the cool air within the rear-flank downdraft reaches the ground, it diverges horizontally and begins to spiral around the low-level circulation. After several minutes, the rear-flank outflow wraps around the low-level circulation, and the rear-flank gust front leading it contacts the forward-flank gust front, forming an occlusion point. This period marks the mature stage of the supercell lifecycle, and if a tornado is going to form, it will usually do so along the occluded portion of the gust front.

**Figure 2.2.** Schematic diagram of the three-dimensional flow in a supercell storm. Surface gust fronts are illustrated with thin black lines. Adapted from Lemon and Doswell (1979).
Figure 2.3. Vertical velocity (thin contours; \( w < 0 \) is dashed), the \( 0.5 \text{ g kg}^{-1} \) rainwater contour (the heavy contour), storm-relative horizontal winds (vectors), and cloud water > \( 0.4 \text{ g kg}^{-1} \) (shaded) at \( z = 1 \text{ km} \) from the numerical simulations from Klemp and Rotunno (1983). The location of the vertical vorticity maximum is indicated with a black dot and the rear-flank gust front is traced with a heavy black line. The location of the occlusion downdraft is indicated.

### 2.2 Structure of tornadoes

Before reviewing hypotheses posed in the past literature regarding the production of vertical vorticity near the ground, and to assist with the interpretation of the evolution of the tornadoes examined in this study, it will be helpful to discuss the generally accepted structure of tornadoes as deduced by observations, theory, and idealized numerical modeling studies.

The airflow within and immediately surrounding a tornado often is classified with five regions, the core, outer flow, upper flow, corner flow, and boundary layer flow, illustrated in Fig 2.4. The core flow typically extends outward a few tens to hundreds of meters from the axis of rotation. A pressure deficit consistent with strong rotation is found within the core region. In the outer flow region, the conservation of angular momentum results in tangential winds that increase as air is drawn inward toward the axis of rotation due to the convergence associated with the primary storm updraft (upper flow). The radial profile of tangential velocity approximately mimics a Rankine vortex, with near solid-body rotation
Figure 2.4. A map of the five different flow regimes surrounding a tornado. The regimes are as follows: I) outer flow, II) core flow, III) corner flow, IV) boundary layer flow, and V) upper flow. From Markowski and Richardson (2010).

(i.e., a tangential wind that increases linearly with radius) within the core region and tangential winds in the outer region that decrease according to a 1/radius profile. An example of a similar profile measured in the Spencer, South Dakota tornado on 30 May 1998 (Wurman and Alexander 2005) is shown in Fig 2.5.

Cyclostrophic balance, a balance between an inward-directed (with respect to the center of the tornado) pressure gradient force and an outward-directed centrifugal force, exists in the core region, where frictional affects are assumed negligible. Such a balance is not possible in the boundary layer where surface drag decreases the tangential wind (and the centrifugal force), resulting in an acceleration of the flow inward toward the axis of rotation. The flux of angular momentum toward the axis of rotation in this shallow depth produces intense tangential winds in the corner flow region, where the core and boundary layer regions meet. The inward directed winds from the boundary layer converge in the corner region and erupt upward, resulting in an intense jet of vertical motion near the ground.

Laboratory and numerically simulated tornado chamber experiments have been able to reproduce many aspects of tornado vortex structure observed in nature (e.g., Dessens 1972, Ward 1972, Davies-Jones 1973, Rotunno 1984, Lewellen et al. 2000, Kosiba 2009). These studies are typically designed to have a cylindrical chamber with an inflow area along the sides near the bottom through which rotat-
Figure 2.5. Observed magnitude of the tangential wind profile (red line) and a modeled Rankine profile of tangential winds (blue line) calculated using single-Doppler DOW velocities as a function of radar azimuthal distance across the Spencer, South Dakota tornado. The velocity profile is collected slightly off-axis, explaining its slight asymmetry. From Wurman and Alexander (2005).

\[
S = \frac{RM}{2Q} = \frac{\bar{v}_{\text{tan}}}{\bar{w}},
\]

where \( R \) is the updraft radius, \( 2\pi M \) is the circulation in the chamber, \( 2\pi Q \) is the vertical mass flux through the chamber, \( \bar{v}_{\text{tan}} \) is the mean tangential velocity at \( R \), and \( \bar{w} \) is the mean vertical velocity at the top of the chamber. Figure 2.6 illustrates the changes in the radial and vertical winds within the vortex core flow observed in tornado chambers with varying magnitudes of \( S \) (Davies-Jones 1986). When \( S \) is small (near zero), one-celled vortices may develop, in which vertical motion within the core flow region is upward (Fig 2.6a). As \( S \) increases, negative vertical motion develops in the upper levels of the core flow and builds downward to the top of the corner flow region, where it meets the intense jet of upward motion from
Figure 2.6. Vertical cross sections of the radial and vertical wind through the axis of vertical rotation when $S \approx 0$ (a), $S > 0$ (b), and $S_{\text{critical}} > S$ (c). A multiple vortex structure can be found when $S > S_{\text{critical}}$ (d). Adapted from Davies-Jones (1986).

the boundary layer (Fig 2.6b). If $S$ exceeds a critical value, the jet of downward motion penetrates into the corner region, possibly contacting the ground (Fig 2.6c). When this happens, the tornado is called a two-celled vortex, and smaller vortices can sometimes develop in a cylindrical vortex sheet near the radius of maximum tangential winds (Ward 1972, Lewellen et al. 1997, Wurman 2002). The result is a multiple-vortex system, with a large parent vortex possessing two or more sub-vortices (Fig 2.6d).

Recent high-resolution Doppler radar observations have shown that multiple concentric scales of rotation can sometimes exist (e.g., Wurman and Alexander 2004, Rasmussen and Straka 2007, Marquis et al. 2008). In these cases, a small single-Doppler velocity couplet is found near the center of a larger velocity couplet, both of which are smaller than spatial scales typically associated with mesocyclones (diameter $> 4$ km, Fujita 1981). Figure 2.7 shows one such example of concentric Doppler velocity couplets. It is not known exactly how several concentric vortices form. When a discussion of concentric vortices is necessary in this study, we will loosely refer to the multiple spatial scales of rotation as mesocyclones (diameter $> 4$ km), tornado-cyclones ($4$ km $> \text{diameter} > 2$ km), and tornadoes (diameter $< 4$ km).
2 km). Certainly, dynamical criteria composed of more than merely diameter are important to discerning the origins and physical relevance of multiple concentric scales of rotation; however, determining such dynamic relationships is beyond the scope of this study.

### 2.3 Origins of low-level rotation

Many past studies discuss the possible origins of low-level rotation through the use of numerical models or radar observations that lack the spatial resolution required to capture tornado-scale features. Therefore, distinguishing the processes important for the generation of tornado-scale vortices from those relevant at the mesocyclone-scale is problematic. It is possible that the generation of tornado-scale vorticity first requires the generation of a broader rotation at low-levels that is subsequently intensified via stretching by small-scale updrafts along the rear-flank gust front, (e.g. Brandes 1984 and Rasmussen and Staka 2007). Several studies have put forth hypotheses regarding the generation of low-level mesocyclone-scale and/or tornado-scale vertical vorticity in supercells. The following is a brief list of the prominent hypotheses in the literature:

1. *Influence of the forward-flank gust front*: In numerical simulations of supercells performed by Klemp and Rotunno (1983), Rotunno and Klemp (1985),
Wicker and Wilhelmson (1995), and Wicker (1996), the temperature gradient found along the forward-flank gust front was instrumental to the generation of low-level vertical vorticity. In these studies, air parcels composing the developing and mature low-level mesocyclone travel along the forward flank gust front. While residing in this baroclinic zone, parcels acquire horizontal vorticity at the rate of

\[
\frac{\partial \vec{u}_h}{\partial t} = \nabla \times \vec{B} \hat{k}
\]  

(2.4)

(neglecting frictional effects, the tilting of vertical vorticity into the horizontal, and horizontal stretching). The baroclinically-generated horizontal vorticity acquired by the parcels is mostly streamwise because the parcels travel roughly parallel to the gust front en route to the mesocyclone. Vertical vorticity is generated through tilting as the parcels reach the primary updraft. Subsequent stretching of this vertical vorticity yields more intense rotation near the ground. Wicker (1996) found that low-level rotation was greater in simulated storms when the near-ground environmental horizontal vorticity vector was aligned with the baroclinically-generated vorticity vector along the forward-flank gust front. In this case, parcels acquire greater streamwise horizontal vorticity before reaching the updraft than they do if the environmental horizontal vorticity vector points perpendicular to or opposite the horizontal vorticity vector along the forward-flank gust front.

2. Influence of the rear-flank downdraft: Klemp and Rotunno, Rotunno and Klemp, and Wicker and Wilhelmson particularly emphasize the role that parcels passing through the forward-flank baroclinic zone play in generating the observed intense low-level rotation. However, the vertical tilting and subsequent stretching by the primary updraft of the horizontal streamwise vorticity attained by parcels within the forward-flank gust front cannot by itself create vertical vorticity very close to the ground because the vertical vorticity is being advected upward as it is created through tilting (Davies-Jones and Brooks 1993). Therefore, the presence of a downdraft, which provides a mechanism for parcel descent during tilting, seems to be a necessary condition to produce vertical vorticity close to the ground. From visual ob-
Figure 2.8. A vertical cross section in the north-south direction of parcel trajectories (thick black lines with arrowheads) and vorticity vectors (black vectors) in a hypothetical rear-flank downdraft. The temperature gradient points toward the east (into the page). The vertical motion and the sign of the tilting of horizontal vorticity into the vertical that is experienced by parcels along the trajectories are indicated at bottom. Adapted from Davies-Jones and Brooks (1993).

Observations, radar observations, and simulations, it is known that tornadoes tend to form after the rear-flank downdraft reaches the ground; therefore, it is assumed that the rear-flank downdraft is critical to their formation. Wicker and Wilhelmson (1995) showed that parcels entering the low-level circulation after residing within the rear-flank downdraft contained negative vertical vorticity (albeit, weak negative vertical vorticity), while parcels traveling through the forward-flank baroclinic zone are responsible for the supply of positive vertical vorticity. Davies-Jones and Brooks (1993) show one scenario in which the tilting of horizontal vorticity within a column of air in the rear-flank downdraft, along with baroclinic production of vorticity along the path of descent through the cold pool, could theoretically generate positive vertical vorticity near the ground (Fig 2.8).

Walko (1993) shows how a rear-flank downdraft could generate positive vertical vorticity in the area typically containing a mesocyclone and possibly a tornado by tilting environmental horizontal vortex lines. When a downdraft
Figure 2.9. Three-dimensional schematic of the a) Walko (1993) and b) Straka et al. (2007) hypotheses of the generation of low-level vertical vorticity by tilting of horizontal vorticity in the rear-flank downdraft. The placement and orientation of initially horizontal vortex lines are shown with black lines and arrow-rings. The three-dimensional volume of air occupied by the downdraft is shown with a thin black contour. The placement of a hypothetical rear-flank gust front is illustrated in the bottom panels, along with the locations of positive and negative vertical vorticity that is generated by tilting. Adapted from Straka et al. (2007).

Impinges on westerly shear (i.e., a northward-pointing vorticity vector), vortex lines arch downward, yielding positive vertical vorticity on the northern edge of the downdraft and negative vertical vorticity on the southern edge. The resulting pattern of vertical vorticity, illustrated in Fig 2.9a, is consistent with the frequently observed cyclonic and anti-cyclonic vortex pairs in tornadic storms (e.g., Brandes 1984, Wakimoto et al. 1998, Wakimoto and Cai 2000, Bluestein and Gaddy 2001, Dowell and Bluestein 2002). The positive vorticity generated by this process would reach a larger magnitude than
the negative vorticity on the southern edge of the downdraft because it is located within the low-level convergence beneath the southern-western edge of the primary updraft, although anticyclonic vertical vorticity on the southern edge of the downdraft could be intense when stretched by horizontal convergence along the rear-flank gust front. This mechanism is consistent with that proposed by Brandes (1978, 1981), who, using dual-Doppler data, concludes that low-level vertical vorticity is generated along the rear-flank gust front due to the downward advection of horizontal momentum south of the tornado. Brandes speculates that a sheet of vertical vorticity associated with the horizontal wind shift along the rear-flank gust front could have rolled into discrete vortices owing to a horizontal shearing instability, which are subsequently stretched into tornado-strength circulations. However, he does not attempt to reconcile this mechanism with the common observation of a positive-negative vortex pair.

In idealized numerical simulations, Straka et al. (2007) demonstrate how the tilting of baroclinically-generated vortex lines surrounding a downdraft could explain the creation of vertical vorticity at low-levels. In their model, horizontal vortex rings surround an isolated downdraft that is composed of negatively buoyant air. These rings, advected downward by the downdraft, spread out horizontally when they approach the ground. A low-level updraft, like that typically found along the rear-flank gust front of a supercell storm, can lift the vortex lines upward into an arched shape, producing a pair of oppositely signed vertical vorticity maxima along the edges of the updraft band (Fig 2.9b). The resulting pattern of vertical vorticity resembles that produced by the Walko process of vorticity generation. However, in the Straka et al. mechanism, the vortex lines arch upward between the vertical vorticity pair, rather than downward as in the Walko mechanism. The observation of vortex line arches in some storms (e.g., Markowski et al. 2008) may support the process of baroclinic generation of vortex lines and subsequent arches. This theory presents a conundrum when coupled with observations from Markowski et al. (2002), who found that rear-flank downdrafts that are composed of air with large temperature deficits relative to the ambient environment ($\theta' < -2$ K) tend to be associated with non- or
weakly-tornadic storms, while smaller deficits ($\theta'_\rho > -2$ K) are associated with significantly tornadic storms (i.e., storms that generate intense and long-lived tornadoes). The favoring of warm downdrafts possibly is due to the inability of the updraft to contract angular momentum at low levels when excessively dense cold air is present (Markowski et al. 2003). However, it also is possible that if the rear-flank downdraft air is quite warm, then insufficient horizontal vorticity may be produced. Therefore, an intermediate range of outflow temperatures is implied in order to optimize both the production of vorticity and the ability of the updraft to stretch the vorticity to tornadic strength (Markowski et al. 2008).

3. Enhancement of vertical vorticity along the rear-flank gust front: In some storms, vertical vorticity maxima form along the rear-flank gust front outside of the region in which tornadoes typically are observed (e.g., Dowell et al. 2002, Marquis et al. 2006). These maxima could form by way of a variety of mechanisms, including the tilting and stretching of ambient horizontal vorticity or the tilting and stretching of baroclinic horizontal vorticity created along the gust front. These vorticity maxima have been observed in some storms to travel along the gust front toward the region of the storm in which tornadoes are frequently observed (Fig 2.10). Such vorticity maxima could potentially seed tornadogenesis when they are vertically stretched by convergence beneath the primary updraft. This mechanism for the production of low-level rotation has been observed in storms undergoing cyclic mesocyclogenesis and/or tornadogenesis (e.g., Dowell and Bluestein 2002, Beck et al. 2006) and is assumed to be instrumental in the production of non-supercell tornadoes (e.g., Wakimoto and Wilson 1989, Lee and Wilhelmson 1997).

4. Restructuring of mesocyclone-scale vorticity: Using pseudo-dual-Doppler wind syntheses, Wakimoto and Liu (1998) observe a restructuring of the mesocyclone in the Garden City, Kansas, supercell by an axial downdraft within the core of rotation. The mesocyclone transitions from a one-celled to a two-celled vortex structure with smaller vortices orbiting it, analogous to the development of multiple-vortex tornadoes. One of the smaller vortices
Figure 2.10. Vertical vorticity (shaded and black contours) and horizontal ground-relative velocity (vectors) at $z = 300$ m in a tornadic storm on 5 June 2001 at 0028:22 (left) and 0031:54 UTC (right). Individual vertical vorticity maxima along the rear-flank gust front are tracked in time with letters. Adapted from Marquis et al. (2006)

resulting from this transition intensifies into the observed tornado. They identify the axial downdraft within the mesocyclone as an occlusion downdraft, resulting from $-\partial \zeta / \partial z$. Brandes (1977, 1978) suggests the possibility of similar a restructuring of the storm-scale circulation in the Harrah, Oklahoma, supercell, except that the tornado-scale vortices develop within the core of the mesocyclone. Brandes (1978) somewhat ambiguously hypothesizes that the downdraft causing the restructuring of the Harrah mesocyclone is a result of “unbalanced pressure gradient forces associated with accelerated outflow,” therefore, it is unclear exactly what process is leading to tornado-genesis.

Another way that intense tornado-scale vorticity could be generated from a broader scale of rotation is through corner-flow collapse. Lewellen and Lewellen (2007a,b) performed idealized numerically-simulated tornado chamber experiments in order to investigate the dynamical relationships between the intensification of a tornado whose axis of rotation is collocated with that of a larger circulation, possibly a tornado-cyclone or a mesocyclone. They found that, through sudden changes in the low-level swirl flow, rotation within the corner flow of the outer circulation “collapses,” resulting in a rapid intensification of wind speeds closer to the axis of rotation. While
this theory may seem to be a plausible mechanism for the intensification of tornado-scale vorticity within a broader rotation, it awaits observational verification.

2.4 Tornado dissipation

In many past studies, the demise of tornadoes does not receive as much attention as does their formation. For example, Lemon and Doswell (1979) emphasize the evolution of the storm structure in relation to tornadogenesis, but describe the processes involved with tornado dissipation in a mere two sentences. They state that tornado dissipation occurs when downdraft behind the rear-flank gust front, which already formed an occlusion with the forward-flank gust front near the time of tornadogenesis, continues to wrap around the low-level circulation. This continued occlusion is stated to have the effect of cutting off the low-level inflow to the main updraft. However, no further detail is provided. Brandes (1978) mirrors this observation, stating that tornado dissipation occurs as the low-level mesocirculation chokes the supply of warm inflow air and separates the tornado from the parent updraft. Several observers have noted that, visually, tornadoes often “rope-out,” transitioning from a nearly vertically-oriented vortex, to one that is tilted out of the vertical and narrows as it weakens (e.g., Wakimoto and Martner 1992). It is usually assumed that the tilt of some dissipating tornadoes is caused by the strong outflow winds that advect their bases horizontally relative to cloud base, however, few observational data sets exist to detail this process.

From these accounts, it is implied that tornado dissipation results from the removal of low-level convergence due to a loss of buoyancy available to the updraft as the cold outflow air behind the rear-flank gust front completely surrounds the tornado. However, recent studies have shown evidence in some storms of significant heterogeneity in the outflow air due to successive pulses of rear-flank downdraft. For example, using mobile mesonet data, Finley and Lee (2004, 2008), Lee and Finley (2004), Hirth et al. (2008), and Lee et al. (2008) have shown distinct wind shift lines in the outflow of at least five storms, which they believe are boundaries between multiple rear-flank downdraft surges. In some of these cases, thermodynamic heterogeneity exists across the multiple downdraft surges, with $\theta_v$ differences...
across wind shift boundaries as high as 2–3 K. Finley and Lee (2008) show one example (Fig. 2.11) in which $\Delta \theta'_v$ and $\Delta \theta'_e$ (where $\Delta$ denotes the difference between an outflow surge and a subsequent outflow surge) are negative across two observed wind shift lines, but become positive after a third line, indicating that waves of outflow air were not always colder than the previous wave. However, in some multiple RFD surge events, $\Delta \theta'_v$ and $\Delta \theta'_e$ were small (< 1 K), indicating a mostly kinematic variability within the outflow. It is difficult to connect these observed features to the behavior of a tornado because of the low spatial density of mobile mesonet observations. Wurman et al. (2007a) and Marquis et al. (2008) show a more complete two-dimensional structure of possible multiple rear-flank downdrafts in the Kiefer, Oklahoma, and Crowell, Texas, storms, respectively, using dual-Doppler data. In both cases, secondary gust fronts, evidenced by a clear shift in the horizontal wind with a band of horizontal convergence leading strong divergence and horizontal winds, are seen near the surface behind a preceding rear-flank gust front (Fig. 2.12). In the Crowell case, the secondary gust front wraps around the tornado throughout the observation period similar to the expected evolution of the gust

Figure 2.11. Mobile mesonet observations of $\theta'_e$ (shaded) and ground-relative horizontal winds (barbs) collected in the 23 May 2008 Quinter, Kansas supercell storm plotted every 30 seconds. White lines represent possible gust fronts leading multiple rear-flank downdraft surges. From Finley and Lee (2008).
Figure 2.12. Horizontal convergence (shaded) and horizontal velocity (vectors) near the
ground in the Crowell, Texas (left), and Kiefer, Oklahoma (right), supercells. Vertical
vorticity is contoured in the left panel. The location of the primary and secondary rear-
flank gust fronts and the tornadoes are labeled in each panel. Adapted from Marquis et
al. (2008) and Wurman et al. (2007), left and right, respectively.

front separating the initial surge of outflow from the environment. The secondary
gust front in the Kiefer storm does not directly contact the tornado during the
observation period, a possible consequence of the short dual-Doppler deployment.
The role that the secondary downdraft surges and accompanying gust fronts play
in tornado maintenance and dissipation has not been detailed in the literature.

2.5 Tornado maintenance

To the author’s knowledge, there have been only two studies that have detailed
the roles that storm-scale processes may play in tornado maintenance. Wicker and
Wilhelmson (1995) describe the lifecycle of two tornadoes within a numerically
simulated supercell. The first tornado, which lasts approximately 10 minutes, dis-
sipates at a time when the near-surface flow becomes strongly westerly, causing
the lowest several hundred meters of the weakening tornado to become located
southeast of the primary thunderstorm updraft. A second tornado, which lasts approximately 15 minutes, dissipates in a similar manner to the first when strong westerly winds develop in the southern half of the tornado circulation. From these accounts of tornado demise, it is not immediately clear what causes the strong near-surface winds that dislocate the lowest portions of the tornado from the influence of the thunderstorm updraft. They briefly mention the development of an occlusion downdraft that surrounds one of the tornadoes with divergence at low-levels at the time of its demise and erodes the updrafts surrounding it. However, they do not clearly relate the strong near-surface westerly winds with the occlusion downdraft, which they imply could be important to tornado demise. Though Wicker and Wilhelmson perform a trajectory analysis showing that the air entering the tornadoes at their most intense stages originates within the ambient air mass (with subsequent travel along the low-level thermal gradients along the forward-flank gust front) and from within the outflow air (Fig 2.13), they do not show trajectory calculations throughout the lifecycle of the tornadoes, which would have illustrated possible changes in the flow entering the tornadoes as they transition from a strengthening to a weakening state. Therefore, though Wicker and Wilhelmson do suggest some processes relevant to tornado maintenance, they do not do so with sufficient detail to isolate the factors controlling tornado duration.

Dowell and Bluestein (2002b; hereafter DB02b) examine the lifecycles of three tornadoes that occurred within a supercell undergoing repeated cyclic mesocyclogenesis and tornadogenesis near McLean, Texas. In such cyclic storms, the mesocyclone and primary updraft at low-levels frequently dissipate in rear-flank outflow and are replaced by new circulations that develop along the rear-flank gust front (e.g., Adlerman et al. 1999, Dowell and Bluestein 2002a, Beck et al. 2006). Two of the McLean, Texas tornadoes caused F2 damage (associated with peak winds of 50–69 m s\(^{-1}\)) and existed for approximately 12 and 23 minutes, while a third tornado survived for at least 70 minutes and reached peak wind speeds consistent with F4 damage (93–116 m s\(^{-1}\)). To demonstrate the possible reasons why one tornado within the storm had a significantly greater longevity than the others, DB02b showed 1) the horizontal advection and vertical stretching of vertical vorticity terms near the tornadoes dominate the other terms in equation 2.1 in the lowest 1 km, 2) the direction of the ground-relative motion of each tor-
Figure 2.13. Vertical velocity (shaded; $w > 2 \text{ m s}^{-1}$ is dark gray, $w < -2 \text{ m s}^{-1}$ is light gray) and parcel trajectory paths projected into a horizontal plane (a) at $z = 100$ m AGL in the Wicker and Wilhelmson (1995) simulated supercell. The position of the tornado is labeled with a ‘T’. Three-dimensional vortex lines within the dashed box in panel a are shown in panel b. From Wicker and Wilhelmson (1995).

Tornado closely matched the direction of the mean wind surrounding it, and 3) the weakening of the tornadoes occurred while their motion vectors were pointed in a direction drastically different than the motion of the primary updraft. The mean flow surrounding the tornado was influenced by both gusting outflow winds and the environmental inflow to the storm. When the magnitude of the outflow velocity was similar to that of the opposing inflow winds, the updraft-relative motion of the tornado was near zero, keeping it beneath the updraft and supplying it with a steady source of convergence and vorticity. Decreased outflow winds mark the beginning of the dissipation stage of each tornado. At these times, the updraft-relative motion of each tornado points to the left-rear-flank of the storm owing to the influence of the strong southeasterly low-level inflow (Fig 2.14). This causes each tornado to become dislocated from its greatest source of low-level buoyancy, convergence, and vertical vorticity, ultimately leading to its demise. Therefore, DB02b conclude that long-lived tornadoes are possible when their motions closely match those of the primary storm updraft so that they remain attached to the storm structures that supply them with the best possible convergence and verti-
Figure 2.14. Ground-relative track (gray lines) and the average ground-relative wind within 4 km of the three tornadoes observed in the McLean, TX supercell. The times corresponding to tornado locations and wind averages are labeled in each panel. From Dowell and Bluestein (2002).

cal vorticity. In their storm, discrete pulses of rear-flank outflow afforded certain tornadoes longer periods of near-zero updraft-relative motion, while shorter-lived tornadoes occurred during periods of weak outflow winds.
Three-dimensional wind fields are produced at certain analysis times in each case and allow us to evaluate the structure and evolution of the storm-scale flow surrounding each tornado. When both radars were simultaneously deployed with a minimum beam-crossing angle of $30^\circ$, the three-dimensional wind data are retrieved via a dual-Doppler wind synthesis. In two out of the four cases analyzed, dual-Doppler data are supplemented with three-dimensional wind and temperature fields calculated using the Ensemble Kalman filter (hereafter EnKF) data assimilation technique. This chapter describes the steps taken to produce three-dimensional wind data for each storm using the dual-Doppler wind synthesis and EnKF data assimilation technique.

3.1 Dual-Doppler Wind Synthesis

3.1.1 Processing of radar data

The single-Doppler data are collected in a truck-relative framework; therefore, prior to the retrieval of any three-dimensional wind fields, the data must be aligned to
Figure 3.1. A sample of the aerial photography data (top left), and DOW reflectivity at the 0.5° elevation angle (bottom left) used to determine the ground-relative azimuth-offset angle of the first DOW2 deployment of the 30 April 2000 case. The positions of power poles and the radar are indicated in each panel. The difference between the dashed vector and due west in the overlay of these two panels (right) is the magnitude of the azimuth-offset angle.

Once rotated to an Earth-relative framework, the quality of the raw DOW data
are improved by the following steps:

1. Data with a high noise-to-signal ratio are removed by deleting radar gates with a normalized coherent power value < 0.2. Values of normalized coherent power range from 0.0 to 1.0, with a value of 0.0 indicating no coherent Doppler signal and 1.0 indicating a perfect Doppler signal. Removing these data eliminates most range-folded echoes, noisy clear-air returns, and badly attenuated radar signals.

2. Ground clutter contamination is eliminated from the radar data. Ground clutter signals are identified as having relatively high values of radar reflectivity, unusually low radial velocity, and no motion in successive radar volumes. All of these symptoms usually occur in elevation angles less than 2.0°.

3. Values of radial velocity that are above or below the positive or negative Nyquist frequency are subjectively dealiased. The Nyquist velocity,

\[
V_{Nyq} = \pm \frac{PRF \cdot C}{4f},
\]  

where \(PRF\) is the pulse repetition frequency of the radar, \(C\) is the speed of light, and \(f\) is the frequency of the radar.

After these modifications are performed, an artifact of data collection called “jitter” is corrected. This artifact is realized as an oscillation of successive clockwise and counterclockwise radar sweeps about a central azimuth angle, arising due to imperfections in the radar antenna motor and a time stamp for each radar ray that does not correspond to the time the antenna is pointing at the central angle of each radar beam. The dashed line in Fig 3.2 demonstrates the azimuthal oscillation with height of the center of a tornado found in a radar volume containing jitter collected by DOW2 in the 30 April 2000 storm intercept. An azimuthal correction of the jitter in each radar deployment is estimated during the data alignment step. This correction is determined by taking the difference between the average offsets of at least two clockwise sweeps and two counterclockwise sweeps. A typical azimuthal correction of radar jitter across all four storms is \(\pm 0.18°\). The solid line in Fig 3.2 illustrates the radar volume with jitter removed.
Figure 3.2. Azimuth angle (degrees) vs. height (km) of the center of the 30 April 2000 tornado in subsequent clockwise and counterclockwise sweeps from one three-dimensional volume with jitter (dashed line) collected by DOW2. The solid line indicates the position of the tornado center after a jitter correction of ± 0.2° (arrows) is applied.

3.1.2 Objective analysis of radar data

Prior to the dual-Doppler wind synthesis, the edited DOW data are objectively analyzed to remove noise that arises due to beam spreading and possible instrument error. A multi-pass Barnes objective analysis technique is used in order to minimize the over-smoothing of well-resolved spatial scales (Majcen et al. 2008). Such a technique has the effect of steepening the response function, $\beta$,

$$\beta_n = \beta_{n-1} + (1 - \beta_{n-1}) \exp \left[ -\kappa \cdot \gamma^{n-1} \left( \frac{\pi}{\lambda} \right)^2 \right],$$  \hspace{1cm} (3.2)

compared to that for a one-pass analysis,

$$\beta_1 = \exp \left[ -\kappa \left( \frac{\pi}{\lambda} \right)^2 \right],$$  \hspace{1cm} (3.3)

where $n$ is the number of passes, $\lambda$ is the spatial wavelength of a meteorological phenomenon, and $\gamma$ is the convergence parameter. A two-pass method with $\gamma =
0.3 is chosen for all analyses based on the results of a comparison of one-, two-, and three-pass objective analyses used in dual-Doppler wind syntheses of synthetic radar data performed by Majcen et al. (2008). The Barnes smoothing parameter (Barnes 1964), $\kappa = (1.33 \cdot \mu)^2$, as recommended by Pauley and Wu (1990), where, $\mu = \phi \cdot R$ is the observed data spacing, $\phi$ is the beam width of the DOWs (0.93°), and $R$ is the longest distance between a radar and the farthest edge of the desired portion of the dual-Doppler domain. This method of using the coarsest observed data resolution to smooth all radar volumes in a dual-Doppler deployment is recommended by Trapp and Doswell (2000) to “avoid under-smoothing the data in regions of the most coarse data-point spacing”. For all four cases, $\kappa = 0.187$ km$^2$, ensuring that similar spatial scales are retained across all dual-Doppler analyses. Figure 3.3 compares one- and two-pass Barnes response functions, showing the increased retention of smaller scales with a two-pass analysis over a one-pass analysis.

The radial velocity data are interpolated to a Cartesian grid in order to facilitate a more intuitive implementation of the dual-Doppler wind synthesis. Extrapolation of data onto the grid (i.e., assigning values to grid points that are beyond the edge of the radar data) is limited by requiring that each objectively analyzed data point
be influenced by observations in all surrounding octants. From recommendations made by Koch et al. (1983), the Cartesian grid spacing, $\Delta$, has a value between $\mu/3$ and $\mu/2$ in order to represent the resolved wavelengths while minimizing noise in spatial derivatives of the resulting wind fields. For all dual-Doppler syntheses, $\Delta = 150$ m. During the objective analysis, the horizontal positions of data are adjusted in order to correct an artificial tilt with height of meteorological features that arises due to their motion during the time it takes each radar to complete a full volumetric scan. The data locations are adjusted to a central time representative of each radar volume using a reference velocity that is estimated by tracking the motion of each tornado during the observation period.

### 3.1.3 Dual-Doppler wind syntheses

Once the single-Doppler data are interpolated to the Cartesian grids, the dual-Doppler wind syntheses are performed using the calculations outlined in the Appendix. Briefly summarized, the three-dimensional wind field is synthesized by an iterative upward integration of the anelastic mass continuity equation with the lower boundary condition of $w = 0$ m s$^{-1}$ at $z = 0$ km. Iterations are performed to adjust the $u$, $v$, and $w$ fields until the change in the density-weighted $w$ between iterations is less than 0.01 kg m$^2$ s$^{-1}$. Corrections for the centrifuging and falling of debris and hydrometeors are not performed because of power attenuation along beams in the heavy precipitation, an uncalibrated DOW reflectivity factor, and an
**Figure 3.4.** Radar deployment maps for the a) 5 June 2001, b) 3 June 1999, c) 30 April 2000, and d) 22 May 2004 DOW intercepts. Ground-relative tornado and DOW positions are indicated in each panel. The times associated with the DOW and tornado locations also are indicated. Dual-Doppler lobes are indicated with black circles. Shaded areas indicate base-line regions in which no dual-Doppler data are available.

unknown scatterer type. Fall speed errors are assumed to be small because of the small antenna elevation angles used ($\leq 15.0^\circ$).

Table 3.1 lists the dual-Doppler analysis times, table 3.2 lists the antenna elevation angles that compose the radar volumes, and Fig 3.4 shows the radar deployment geometry in each storm intercept. At certain times in the 30 April 2000 and 3 June 1999 storm intercepts, only $0.5^\circ$ elevation angle sweeps are collected (in order to achieve high-temporal-resolution data near the ground). At these times, 2D, quasi-horizontal dual-Doppler wind syntheses are performed, retrieving the horizontal wind fields and their horizontal derivatives (e.g., horizontal convergence and vertical vorticity) analyzed at heights AGL where the radar sweeps intersect the tornado. At other analysis times, three-dimensional dual-Doppler syntheses are produced. In the field, efforts were made to synchronize elevation angle sweeps collected by both radars. However, in some cases, data collection became asynchronized, but by no worse than 15-20 seconds. For the quasi-two-dimensional dual-Doppler analyses, an average asynchronization is approximately 5 seconds.
Due to the radar geometries of each storm intercept, the highest height AGL at which the synthesized $u$, $v$, and $w$ fields are available near each tornado is less than 2.5 km. A downward extrapolation of wind data is necessary in order to apply the lower boundary condition because the lowest radar elevation angle is $0.5^\circ$. This extrapolation is performed by setting the missing near-ground $u$ and $v$ wind components equal to those at the lowest level at which both radars collect data. For the 22 May 2004, 30 April 2000, and 5 June 2001 dual-Doppler analyses, the lowest level for which both radars provide data is one grid point (150 m) above the ground; therefore, the necessary extrapolation is minimal. The leftmost panels in Fig 3.5 shows an example of the kinematic fields produced by the dual-Doppler technique described above from the 5 June 2001 storm.

Marquis et al. (2008) compared the $w$ and vertical vorticity fields synthesized in the 30 April 2000 case using this low-level extrapolation technique and an alternative technique in which the missing low-level $u$ and $v = 0$. Overall, the qualitative differences in the patterns of the $w$ and vertical vorticity fields using these two extrapolation methods are minor. The magnitude of the $w$ field at $z = 1$ km AGL is, on average, about 1 m s$^{-1}$ smaller along the rear-flank gust front when the low-level $u$ and $v = 0$. The greatest differences in the $w$ fields between the two extrapolation techniques are 1 to 4 m s$^{-1}$ in close proximity to the tornado. Marquis et al. (2008) show that the sensitivities of trajectory calculations using the data produced by the two extrapolation techniques are generally minor,

<table>
<thead>
<tr>
<th>Case</th>
<th>obs period (hhmm UTC)</th>
<th>DOW2 elev angle (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3 June 1999</td>
<td>2320-2350, 0029-0043</td>
<td>0.3, 1, 2, 3, 4, 5, 7, 9, 11, 13, 15</td>
</tr>
<tr>
<td></td>
<td>0044-0048</td>
<td></td>
</tr>
<tr>
<td>30 April 2000</td>
<td>2104-2129</td>
<td>1, 1.5, 2.2, 3.5, 4.5, 6.5, 8.5, 10.5, 13.5</td>
</tr>
<tr>
<td>5 June 2001</td>
<td>0014-0035</td>
<td>0.5, 1.2, 2, 3, 4, 5, 6, 7.5, 9, 11, 13, 16</td>
</tr>
<tr>
<td>22 May 2004</td>
<td>2258-2311</td>
<td>1, 1.5, 2, 2.5, 3, 4, 5, 6, 7, 9, 11, 13, 15</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Case</th>
<th>obs period (hhmm UTC)</th>
<th>DOW3 elev angle (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3 June 1999</td>
<td>2350-0004, 0016-0018,</td>
<td>0.3, 1, 2, 3, 4, 5, 7, 9, 11, 13, 15</td>
</tr>
<tr>
<td></td>
<td>0036-0043</td>
<td>0.3, 1, 2, 3, 4, 5, 7, 9, 11, 13, 15</td>
</tr>
<tr>
<td></td>
<td>0019-0021, 0043-0045</td>
<td>1</td>
</tr>
<tr>
<td>30 April 2000</td>
<td>2100-2109</td>
<td>0.5</td>
</tr>
<tr>
<td></td>
<td>2110-2119</td>
<td>0.5, 2, 3.5, 5, 6.5, 8, 9.5, 11, 12.5, 14, 15.5</td>
</tr>
<tr>
<td>5 June 2001</td>
<td>2350-0040</td>
<td>0.5, 1.2, 2, 3, 4, 5, 6, 7.5, 9, 11, 13, 16</td>
</tr>
<tr>
<td>22 May 2004</td>
<td>2257-2311</td>
<td>0.3, 0.6, 1, 1.5, 2, 2.5, 3, 4, 5, 6, 7, 9, 11, 13, 15</td>
</tr>
</tbody>
</table>

Table 3.2. Sweep elevation angle lists composing each DOW2 (top) and DOW3 (bottom) radar volume during the observation periods listed in the middle column.
Figure 3.5. a-c) Vertical velocity (shaded), vertical vorticity (contoured; outermost contour is \(0.02 \, \text{s}^{-1}\), incremented by \(0.01 \, \text{s}^{-1}\)), and ground-relative horizontal wind (vectors); d-f) horizontal convergence (shaded), vertical vorticity (contoured; outermost contour is \(0.02 \, \text{s}^{-1}\), incremented by \(0.01 \, \text{s}^{-1}\)), and ground-relative horizontal wind (vectors); g-i) east-west component of vorticity (shaded), vertical velocity (contoured; outermost contour is \(1 \, \text{ms}^{-1}\), incremented by \(2 \, \text{ms}^{-1}\)); j-l) north-south component of vorticity (shaded), vertical velocity (contoured; outermost contour is \(1 \, \text{ms}^{-1}\), incremented by \(2 \, \text{ms}^{-1}\)); m-o) vertical vorticity (shaded), vertical velocity (contoured; outermost contour is \(1 \, \text{ms}^{-1}\), incremented by \(2 \, \text{ms}^{-1}\)); at 0028 UTC and at \(z = 750 \, \text{m AGL}\) in the 5 June 2001 storm produced using a high-resolution dual-Doppler wind synthesis (with single-Doppler data objectively analyzed with \(\kappa = 0.187 \, \text{km}^2\) and \(\Delta = 150 \, \text{m}\); left column), using a lower-resolution dual-Doppler wind synthesis (\(\kappa = 0.65 \, \text{km}^2\), \(\Delta = 300 \, \text{m}\), \(\gamma = 1.0\); middle column), and the EnKF data assimilation experiment outlined in chapter 3.2 (right column). In the right column, all fields are ensemble averages of the model results immediately after data assimilation has occurred.

Presumably because of the short period of dual-Doppler data availability, which limits the differences in parcel positions at the last integration time step. Marquis et al. (2008) do not demonstrate the sensitivity of the horizontal vorticity fields to the assumed missing \(u\) and \(v\) data. Figure 3.6 shows the differences between the \(\hat{x}\) and \(\hat{y}\) components of vorticity (\(\xi\) and \(\eta\), respectively) synthesized from both
assumed profiles of missing $u$ and $v$ data in the 30 April 2000 case. At $z = 1$ km, $\xi$ is larger (smaller) southeast (northwest) of the occluded gust front north and northwest of the tornado when $u$ and $v$ are assumed to be equal at the lowest two grid levels. On average, these differences are on the order of $10^{-3} \text{s}^{-1}$. Larger differences are found on smaller scales close to the eastern edge of the dual-Doppler domain in the outflow air southeast of the tornado. The differences in $\eta$ are on the order of $10^{-3} \text{s}^{-1}$ in the eastern domain, approaching zero in the western domain,
while the peak values of $\eta$ and $\xi$ along the gust fronts are on the order of $10^{-2}$ to $10^{-1}$ s$^{-1}$; therefore, the two low-level extrapolation methods are believed to produce quantitatively similar vorticity fields.

### 3.2 Ensemble Kalman filter data assimilation

The EnKF technique employed by this study generally follows that of Snyder and Zhang (2003) and Dowell et al. (2004), in which single-Doppler radial velocity observations are assimilated into a numerically simulated convective storm. A
brief outline of the EnKF process is as follows,

1. An ensemble of simulated storms is integrated forward in time from the initial time until a time when radar observations are available to assimilate.

2. Analyses of the ensemble mean state of the simulated storms, \( \bar{x}^a \) (where \( x \) is a model variable, such as \( u, v, w, \theta \), etc.), are produced according to the following expression,

\[
\bar{x}^a = \bar{x}^f + WK \left[ y_o - H(\bar{x}^f) \right],
\]

(3.4)

where \( \bar{x}^f \) is the ensemble mean of \( x \) prior to data assimilation (note, the superscripts ‘f’ and ‘a’ denote a ‘forecast’ and an ‘analysis’ or ‘updated’ value), \( W \) is a weighting factor (also called the “localization”), \( y_o \) is a radar velocity observation, \( H \) is the observation operator, which converts modeled \( u, v, \) and \( w \) fields into radial velocity observations relative to the location of the radar in the model reference frame and performs any necessary interpolation of data [therefore, the \( H(\bar{x}^f) \) term in equation 3.4 is the ensemble mean of the modeled radial velocity field], and \( K \) is the Kalman gain,

\[
K = \frac{1}{N-1} \sum_{n=1}^{N} \left( x_n^f - \bar{x}^f \right) \left( H x_n^f - H(\bar{x}^f) \right) \frac{1}{\sigma_{obs}^2 + \frac{1}{N-1} \sum_{n=1}^{N} \left( H x_n^f - H(\bar{x}^f) \right)^2},
\]

(3.5)

where \( N \) is the ensemble size, the subscript \( n \) denotes an individual ensemble member, and \( \sigma_{obs}^2 \) is the error variance of the radar velocity observations, assumed to be \((2 \text{ m s}^{-1})^2\). Note that in equation 3.5, the numerator corresponds the covariance of \( x_n^f \) and \( H(\bar{x}^f) \) (the ‘background error covariance,’ i.e., how the modeled radar velocities correlate with other model fields, such as \( u, v, w, \theta \), etc.), and the right-most term in the denominator corresponds to the error variance of the ensemble of modeled radar velocity. Thus, the \( K(y_o - H(\bar{x}^f)) \) term in equation 3.4 can be interpreted as a correction to the ensemble mean of each model field equal to a product of the difference between observed radial velocity and modeled radial velocity, and how the modeled radial velocity field correlates with the other model fields. The term,
3. Model variables in individual ensemble members are updated as follows,

$$x_n^a = \bar{x}^a + x_n^I - \bar{x}^I + WK \left[ H(x_I) - H(x_n^I) \right].$$

(3.6)

An “Ensemble Adjustment Filter” method is used to control the effects of a reduced analysis-error covariance based on using an ensemble of a finite size (Anderson 2001),

4. The updated storm ensemble is integrated forward in time until the next set of available observations. Steps 2–4 are repeated until all observations are processed.
This procedure is performed using data assimilation software from the National Center for Atmospheric Research called the Data Assimilation Research Testbed (DART). The Weather Research and Forecasting model (WRF) is used to perform the numerical simulation of the ensemble. The parameters used in both the WRF and DART components of the data assimilation experiments are discussed below.

### 3.2.1 Radar data assimilated

Single-Doppler radial velocity data are assimilated in each of the experiments. Radar reflectivity data are not assimilated for two reasons: 1) reflectivity factors for the DOWs are uncalibrated, and 2) beam attenuation in the heavy precipitation of the storms yields erroneous spatial derivatives of the reflectivity factor. Data from only one DOW radar are assimilated at a given time because of computational constraints, although coarse-resolution EnKF experiments in which data are assimilated from both radars (when available) yield similar results because the additional radar typically had a similar viewing angle as the first radar and/or a comparatively smaller data coverage. The storms analyzed in this study are located at ranges greater than 100 km from the nearest WSR-88D. At such a range, the list of sweep elevation angles employed in a typical WSR-88D volume does not provide a vertical distribution of data in the lowest few kilometers of the storm that is sufficient to improve the assimilation result, similar to the findings of Zhang et al. (2004). The quality of the data assimilation results can be tested by comparing them to available dual-Doppler syntheses, which are an independent estimate of the winds in the storm because data from only one DOW are assimilated during an analysis period. Dual-Doppler velocity data are not assimilated in order to prevent the introduction of any error that might arise due to assumptions in their synthesis, such as the assumed near-surface convergence profile necessary to assume the lower boundary condition.

Assimilated DOW velocities are objectively analyzed in a different manner from that used in the dual-Doppler wind synthesis. As in Dowell et al. (2004), radar sweeps are objectively analyzed separately to conic surfaces; data are analyzed on a rectangular grid on the conic surface, but the height of each observation is located along each radar beam rather than being interpolated to a regular vertical grid.
Figure 3.8. The timelines of the a) 5 June 2001, and b) 3 June 1999 EnKF experiments, including the experiment durations, the times of the assimilated DOW data, and the known lifetimes of the tornadoes. The dashed line in b) indicates the uncertainty of the tornadogenesis time. The gray line in b) indicates the period of synthetic DOW2 data (see text for explanation).

This interpolation method reduces errors that arise owing to irregular observation resolution and preserve the high density of observations in the vertical near the radar. A Cressman weighting function,

$$W_{Cress} = \frac{R_c^2 - r^2}{R_c^2 + r^2},$$  \hspace{1cm} (3.7)

is used to objectively analyze data to the horizontal grid, where $R_c$ is the radius of influence, and $r$ is the distance between an observation and a grid point. The radius of influence is equal to the horizontal grid spacing, which is 1 km. Extrapolation to the horizontal grid is limited by requiring that each grid point be influenced by at least three radar observations with a minimum combined Cressman weighting of 0.1.

Based on the results of some early feasibility EnKF experiments, it is found that at least 40 continuous minutes of DOW observations collected within the low-level mesocyclone, rear-flank outflow, and forward-flank are necessary to produce a storm that is similar to the dual-Doppler result. These criteria are only met by two of the storms that are analyzed in the present study, the 3 June 1999 and 5 June 2001 cases. Figure 3.8 shows the assimilation periods of the DOW data in the final
Figure 3.9. a–c) DOW single-Doppler radial velocity (shaded) and the 20 dBZ_e reflectivity contour at the 0.5° elevation angle at 0016 UTC (left), 0030 UTC (middle), 0048 UTC (right) on 3 June 1999. A blue line denotes the assumed position of the rear-flank gust front at each time and a ‘T’ marks the location of the tornado. d–f) vertical velocity (shaded), vertical vorticity (black contours; outermost contour is 0.002 s\(^{-1}\) incremented by 0.001 s\(^{-1}\)), and radar reflectivity computed from model precipitation fields (green contours; outermost contour is 35 dBZ_e, incremented by 5 dBZ_e) at 0016 UTC, 0030 UTC, and 0048 UTC and at 750 m AGL from the EnKF experiment of the 3 June 1999 storm in which the 0029–0030 UTC volume collected by DOW2 is used to fill the data gap between 0004–0029 UTC. g–i) same as panels d–f, but using the results of an EnKF experiment in which no data interpolation is performed between 0004 and 0029 UTC. The data plotted in panels d–i are the ensemble averages of the model fields immediately after data assimilation has occurred.

3 June 1999 and 5 June 2001 EnKF experiments. During the 5 June 2001 storm intercept, DOW3 collected data continuously for 50 minutes at a range sufficient to observe the entire storm below approximately 4 km AGL. All of these data are assimilated in the EnKF experiment. In the 3 June 1999 EnKF experiment, data collected by DOW2 prior to tornadogenesis between 2320 and 2350 UTC are assimilated, followed by a 14-minute period of DOW3 data between 2350 and 0004 UTC. A data gap exists between 0004 and 0029 UTC because the DOWs were redeploying to keep up with the moving storm, with only a small amount of data...
collected by DOW3 from 0016 to 0022 UTC. Unfortunately, these DOW3 data are limited to a small spatial coverage within 5–10 km of the tornado, and many of the data are collected only in 0.5° elevation angle sweeps. Therefore, assimilating these data does not improve the overall representation of the storm. The second DOW deployment begins at 0029 UTC with data from DOW2 covering the rear-flank and most of the forward-flank precipitation core of the storm. These data are assimilated through 0048 UTC, which marks the time of tornado dissipation. Between 0044 and 0048 UTC, the DOW2 data are collected only in 0.5° elevation angle sweeps. DOW3 collected data between 0036 and 0048 UTC, but, similar to the data collected between 0016-0022 UTC, these DOW3 observations are limited to within a few kilometers of the tornado and assimilating these velocities does not improve the overall storm-scale results. In early feasibility EnKF experiments for 3 June 1999, the low-level mesocyclone and updraft deteriorate between 0004 and 0029 UTC due to insufficient DOW observations during this period combined with apparent model error. To prevent this deterioration, the observation gap between 0004 and 0029 UTC (gray line in Fig 3.8b) is filled with synthetic data that are produced from the DOW2 radar volume observed between 0029–0030 UTC. This observation synthesis is performed by spatially translating the data in the 0029-0030 UTC DOW2 volume to their assumed prior locations every two minutes between 0005 and 0027 UTC using an average storm motion. The synthesis method seems to be reasonable between 0016 and 0029 UTC because the observed structure of the tornado and the rear-flank gust front are fairly steady during this period (based on the limited DOW3 observations available at 0016 UTC). Figure 3.9 compares the results of two EnKF experiments: 1) using the synthetic data between 0004 and 0029 UTC, and 2) using no synthetic DOW2 data between 0004 and 0029 UTC, but using the briefly collected DOW3 data between 0016 and 0022 UTC. A stable mesocyclone and gust front structure that closely resembles the observed storm at 0016 (Fig 3.9a) and 0029 UTC (Fig 3.9b) is produced in experiment 1 after a short spin-up period that begins at 0005 UTC due to the introduction of the synthetic data (Fig 3.9d,e). The mesocyclone and gust front produced at 0016 and 0029 UTC in experiment 2 (Fig 3.9g,h) do not resemble radar observations quite as closely. While the two experiments produce similar results at the final assimilation time (Fig 3.9f,i), the fact that the mesocyclone and
gust front structures from experiment 1 more closely resemble observations than those in experiment 2 during the period of missing DOW observations suggests that assimilating synthetic DOW2 data between 0004 and 0029 UTC produces a better result.

### 3.2.2 Specific WRF implementation

The model grids are mapped to geographical coordinates (latitude, longitude, and altitude above sea level) where the storms are observed so that both model data and radar data are oriented in an Earth-relative frame. The grid dimensions are sufficiently large to encompass the entire volume occupied by each storm, plus an additional 10 km in all horizontal directions in order to avoid any interaction with the lateral boundaries. The model grid dimensions for the 5 June 2001 and 3 June 1999 experiments are $110 \times 80 \times 20 \text{ km}^3$ and $100 \times 80 \times 20 \text{ km}^3$, respectively. There are 42 grid levels in the vertical dimension, which is stretched in the height coordinate such that the vertical resolution, $\Delta z$, is about 150 m near the ground and 2 km at the top of the grid. A stretched $z$-coordinate was chosen over a uniform grid owing to the high vertical density of the DOW data in the lowest 3 km AGL. The horizontal grid spacing is $\Delta x = \Delta y = 500 \text{ m}$, and the model $\Delta t = 1 \text{ s}$.

A Rayleigh damping layer with a damping coefficient of 0.003 is used in the highest 5 km of the model grid. A 1.5-order turbulence closure scheme is used. Fifth-order horizontal and third-order vertical positive definite advection are used for the momentum and scalar fields. Sixth-order numerical diffusion with a reduction factor of 0.12 is used to reduce $2\Delta x$ noise. Open lateral boundary conditions are used on all sides of the domain with an assumed gravity wave phase speed of 25 m s$^{-1}$. A zero-flux condition is imposed at the lower boundary. The Lin et al. (1983) ice microphysics scheme is used in all experiments, with a Marshall-Palmer graupel intercept parameter of $4 \times 10^4 \text{ m}^{-4}$, a graupel density of 900 kg m$^{-3}$, and a rain intercept parameter of $8 \times 10^6 \text{ m}^{-4}$. These parameters were chosen based on microphysics sensitivity tests described in further detail in section 3.3.2 below.

The base-state soundings for each experiment are composed from different
sources of environmental data. The 5 June 2001 environment is based on a proximity sounding collected at Lamont, OK. This sounding was launched about 60 km south of the storm’s location during the DOW intercept, and about 20 minutes prior to the beginning of radar data collection. Unfortunately, no sounding was launched in close proximity to the inflow of the 3 June 1999 storm. Instead, the base-state environment for this case is the Rapid Update Cycle (RUC) operational forecast model 0000 UTC analysis (within 15 minutes of the DOW deployment time), at a model grid point within about 40 km of the DOW deployment location. Thompson et al. (2003) compared similar RUC profiles to observed soundings collected in severe storm environments and showed that RUC soundings can be “a reasonable proxy for direct observations in the regional supercell environment.” The temperature profiles of the soundings used in the assimilation experiments are modified such that the Richardson number $R_i = (g/\bar{\theta})(\partial \bar{\theta}/\partial z)/(\partial \vec{v}_h |/\partial z)^2$ is slightly greater than 0.25, eliminating the possibility of turbulent mixing of the base state environment. The 1D base state is removed from the full fields before turbulent mixing is calculated to further prevent mixing of the environment (idealized simulations using the unmodified soundings shows a mixing of the environment when the base state is not removed prior to the calculation of turbulence, yielding unusual patterns of convection along the lateral boundary conditions). Further details regarding considerations of the base state soundings used in the EnKF experiments are described in section 3.3.1 and the soundings used in the final EnKF experiments are shown in Fig 3.10.

A cluster of ten elliptically shaped warm bubbles is used to initiate the convective updrafts. The ten bubbles are randomly placed within a $25 \times 25 \times 2$ km$^3$ box that encompasses the volume occupied by the storm at low levels near the beginning of radar observations. Each bubble has a $\theta$ perturbation of 4 K, a radius of 10 km in the horizontal, and 1.5 km in the vertical. The ten bubbles are centered at random locations within the box for each ensemble member. Justification for this randomized bubble method of initiating updrafts is given below.
Figure 3.10. Skew-T/log-P diagrams in the homogeneous base-states used in the 5 June 2001 (left) and 3 June 1999 (right) EnKF data assimilation experiments. Hodographs for each case are shown in the upper right of each panel. Heights AGL (km) are indicated on the hodographs.

3.2.3 Specific EnKF implementementation

In both EnKF experiments, analyses are produced every two minutes, with each analysis assimilating DOW data collected 60 seconds before and after the analysis time. The model ensemble comprises 50 members. An important element of the EnKF data assimilation technique is the method by which the ensemble is initialized (Snyder and Zhang 2003, Zhang et al. 2004, Dowell et al. 2004, Tong and Xue 2005), and how ensemble spread is maintained throughout the experiment (Dowell and Wicker 2009). Maintaining ensemble spread is necessary in order to prevent the ensemble from converging on a presumably inaccurate solution due to model biases (Dee and da Silva 1998) and the repeated assimilation of data into an ensemble of a finite size (Hamill et al. 2001). In this study there are two ways by which ensemble members differ at the initial model time. First, as mentioned above, convective updrafts in each ensemble member are initiated with a different configuration of 10 warm bubbles. Second, random perturbations are added to the base-state horizontal wind profile. These random perturbations have a standard deviation of 2 m s$^{-1}$, and are added to the lowest 11 km of the environment. Randomizing the hodographs in this way adds ensemble spread to the inflow of each storm at the initial time, and sustains it throughout the experiment. Figure 3.11
Figure 3.11. a–c) Vertical velocity (shaded), vertical vorticity (black contours; outermost is 0.003 s⁻¹, incremented by 0.002 s⁻¹), radar reflectivity computed from model precipitation fields (green contours; outermost is 40 dBZe, incremented by 5 dBZe), and storm-relative horizontal winds (vectors) at 750 m AGL and at 0028 UTC for the 5 June 2001 EnKF experiments in which the ensemble spread is dictated by a) 10 initial warm bubbles only, b) 10 initial warm bubbles and randomized $\bar{u}$ and $\bar{v}$ profiles, and c) 10 initial warm bubbles, randomized $\bar{u}$ and $\bar{v}$ profiles, and additive noise throughout the duration of experiment. d–f) The standard deviation of the vertical velocities of the model ensemble (shaded) for the experiments shown in panels a–c. The vertical vorticity and computed reflectivity fields in panels d–f are the same as in panels a–c. The $w = 0.5$ m s⁻¹ contour also is shown in gray. All plots show ensemble averages or standard deviations of the model fields immediately after data assimilation has occurred.

shows an example of the results of two 5 June 2001 EnKF experiments: 1) using the same environmental wind profile for each ensemble member (Fig 3.11a), and 2) using randomized environmental $\bar{u}$ and $\bar{v}$ profiles for each ensemble member Fig 3.11b). The structure of the rear-flank gust front, updraft, and forward-flank is more organized on the inflow side of the storm when randomized hodographs are used than they are when the same hodograph is used for all ensemble members. A linear updraft structure develops with no low-level mesocyclone in the experiment without randomized hodographs. Without the randomized base-state horizontal wind, the ensemble spread is small in the vicinity of the mesocyclone (Fig 3.11d), causing the ensemble to converge upon a solution that is not consistent with radar observations.
In addition to the random warm bubbles and the randomized hodographs, a periodic additive noise technique is employed to maintain ensemble spread throughout the duration of the experiment. As in Dowell and Wicker (2009), Gaussian perturbations are added to the model $u$, $v$, $\theta$, and $q_v$ fields every 5 minutes at locations where the DOW radar reflectivity factor is greater than 25 dBZ$_e$. The perturbation fields are smoothed to spatial scales of 4 km in the horizontal and 2 km in the vertical, and are added to the model fields immediately before the updated ensemble is integrated forward in time. The perturbations have a standard deviation of 1 K and 1 m s$^{-1}$ (for $T$ and $T_d$, and $u$ and $v$, respectively) during the first 20 minutes of each experiment, and 0.5 K and 0.5 m s$^{-1}$ for the remainder. The large perturbations early in the experiments are intended to increase the influence of the observations early in the experiment. Perturbations of 0.5 K and 0.5 m s$^{-1}$ are chosen for the remainder of our experiments based on a series of tests performed by Dowell and Wicker, who found that perturbations of these magnitudes carried the closest balance between the mean innovation and the ensemble spread throughout their data assimilation experiments. Figure 3.11c shows the results from a 5 June 2001 EnKF experiment analogous to those with results shown in Fig 3.11a,b, but with the additive noise technique added. The overall storm structure produced using the additive noise technique method resembles observations moreso than those produced using only the randomized hodographs and initial bubbles. This improvement is because the ensemble spread produced using the additive noise technique is not limited to the southern (inflow) side of the storm (Fig 3.11f). More ensemble spread is found north and west of the precipitation core in the additive noise experiment than in the other two experiments because of an increased development and spreading of the cold outflow on the rear and forward flanks of the storm.

3.2.4 Comparison of EnKF and dual-Doppler analyses

Figure 3.5 compares the results of the final EnKF analysis of the 5 June 2001 storm with the data produced by dual-Doppler wind synthesis. The middle column of Fig 3.5 shows dual-Doppler data that are approximately three times smoother than those in the left panels ($\Delta = 300$ m, $\kappa = 0.65$ km$^2$, $\gamma = 1$), a resolution
closer to that of the model resolution in the EnKF experiment. There is strong qualitative agreement between the dual-Doppler and EnKF $u$, $v$, and $w$ fields at $z = 750$ m AGL; both contain well-defined bands of ascent along the gust fronts and downdraft on the rear-flank with strong northwesterly outflow (Fig 3.5a–c). The EnKF solution contains a stronger, more coherent band of updraft along a possible secondary gust front south of the tornado than is present in the dual-Doppler solution. The EnKF and dual-Doppler low-level vertical vorticity fields are quite similar, both wind retrieval methods produce bands of vertical vorticity along the rear-flank gust front that connect to the tornado (Fig 3.5m–o). Overall patterns of the $x$ and $y$ components of the horizontal vorticity vector (Fig 3.5g–l) and the horizontal convergence fields (Fig 3.5d–f) are similar, but there are some differences in the magnitudes of these fields near the gust fronts. It is not clear which of these two wind retrieval methods produces a more accurate solution because both contain uncertainty in their calculations. For example, the missing DOW observations near the surface affect the calculation of the horizontal vorticity field, while imperfections in the model microphysics scheme limit the accuracy of the modeled cold pool, which affects the baroclinic generation of horizontal vorticity in the EnKF solution. Therefore, it is possible that differences in these two solutions may be within the range of expected variability due to the possible sources of error that are inherent in each method.

Figure 3.12 shows an example of kinematic fields aloft and the buoyancy fields that are produced in the EnKF analysis of 5 June 2001. Though these results cannot be verified because these fields were not directly observed, they do appear to be reasonable: a positively buoyant updraft is seen at $z = 5$ km AGL (Fig 3.12a) and cool outflow air is seen on the rear flank of the storm behind the gust front where a horizontal temperature gradient is present at $z = 750$ m AGL (Fig 3.12b).

### 3.3 Data assimilation sensitivity experiments

Two sets of sensitivity experiments are conducted in order to determine how the EnKF results vary owing to uncertainty in the choice of the base-state environment and microphysics parameters, both of which could affect characteristics of the storms that may be important to understanding tornado maintenance.
Figure 3.12. Density potential temperature (shaded), vertical vorticity = 0.015 $s^{-1}$ (green contour), vertical velocity (thin black contours; the outermost is 2 $ms^{-1}$, incremented by 2 $ms^{-1}$), and radar reflectivity computed from model precipitation fields (gray contours; outermost is 30 dBZe, incremented by 5 dBZe) at a) $z = 5$ km AGL and b) $z = 750$ m AGL at 0028 UTC in the 5 June 2001 EnKF experiment. All plotted data are ensemble averages of the model fields immediately after data assimilation has occurred.

### 3.3.1 Sensitivity to choice of base-state environment

To assess the sensitivity of the overall storm structure to the choice of base-states, three identical EnKF experiments assimilating the 5 June 2001 DOW observations use the following sources of environmental data: 1) the Lamont, Oklahoma, sounding launched 60 km south of the storm intercept site approximately 25 minutes prior to tornadogenesis (Fig 3.13a), 2) the 0000 UTC RUC analysis $\bar{T}$, $\bar{T}_d$, $\bar{u}$, and $\bar{v}$ profiles from the grid location nearest the storm intercept site (Fig 3.13b), and 3) the temperature and dew point temperature profiles of the Weisman and Klemp (1982) sounding (used in several past studies of idealized, numerically simulated supercells) combined with the horizontal wind profile observed at Lamont, OK (Fig 3.13c). In idealized numerical simulations with homogeneous environments and no data assimilation, a long-lived supercell storm initiated by a warm bubble is generated only in the environment of sounding 3. Convective updrafts in the other two environments perish within 30 minutes of model initialization.

Despite the lack of a sustained storm in the idealized simulations using soundings 1 and 2, all three EnKF experiments produce similar kinematic structures at 0030 UTC and $z = 750$ m AGL (Fig 3.14a–c). Peak values of vertical vorticity
Figure 3.13. Skew T-log p diagrams of the base-states used in three EnKF sounding sensitivity experiments of the 5 June 2001 storm. The environments are composed of data collected from 1) a balloon launch at Lamont, Oklahoma, at 0000 UTC, 2) the 0000 UTC analysis of the RUC model at the grid point nearest the DOW deployment site, and 3) the Weisman and Klemp (1982) idealized sounding. Hodographs for each environment are included in the top right of each panel.

differ by only about 0.003 s\(^{-1}\) and the overall magnitudes of vertical velocity along each rear-flank gust front and in each rear-flank downdraft differ by 1–3 m s\(^{-1}\). The greatest difference in the low-level kinematic fields among the three experiments is the placement of the rear-flank gust front at ranges > 5 km south-southwest of the tornado, where radar observations are not available. The low-level horizontal temperature differential between the rear-flank outflow and the ambient inflow differs by approximately 3 K among the three experiments (Fig 3.14d–f). The temperature differential between the ambient environment and the forward flank outflow differs by a similar amount. Soundings 1 and 3 produce similar high-precipitation storms with wide rainy hook echoes present in the computed radar reflectivity. Similar features are present in the DOW reflectivity data. The RUC sounding produces a much smaller precipitation field with a reduced hook echo.

The differences in the low-level kinematic fields are slightly greater among the three experiments at 0040 UTC. The values of peak vertical vorticity differ by up to 0.006 s\(^{-1}\) (Fig 3.15a–c). All three experiments develop a band of upward motion south of the tornado with a peak magnitude that varies by 3–4 m s\(^{-1}\). The temperature of the air near this secondary band of updraft in experiment 1 and 3 is warmer than the preceding outflow air, and is greater than or equal to the environmental temperature (Fig 3.15d, f), but is cooler than the environment.
Figure 3.14.  a-c) Vertical velocity (shaded), vertical vorticity (black contours; outermost is 0.003 s\(^{-1}\), incremented by 0.002 s\(^{-1}\)), radar reflectivity computed from model precipitation fields (green contours; outermost is 40 dBZ\(_e\), incremented by 5 dBZ\(_e\)), and storm-relative horizontal velocity (vectored) at \(z = 750\) m AGL at 0030 UTC in the 5 June 2001 sounding sensitivity experiments.  d–f) Perturbation density potential temperature (shaded), vertical vorticity (black contours; same contour intervals as in panels a–c), computed radar reflectivity (green contours; same contour intervals as panels a–c), the \(w = 1\) m s\(^{-1}\) contour (gray contour), and horizontal vorticity (vectors). All plotted fields are ensemble averages of the model variables immediately after data assimilation has occurred. Data plotted in panels a and d are from the experiment using sounding 1, panels b and e are from the experiment using sounding 2, and panels c and f are from the experiment using sounding 3 in Fig 3.13.

in experiment 2 (Fig 3.15e). Experiment 2 produces a forward-flank downdraft with a magnitude that is about 2 m s\(^{-1}\) greater and a cold pool that is about 2 K colder than in the other two experiments. While the continuity of the gust fronts varies slightly in each experiment, the qualitative updraft/downdraft and thermodynamic structure on the forward and rear flanks of the storm is generally robust, suggesting a mild sensitivity to the choice of base-state environments. The observed Lamont sounding is used in the final 5 June 2001 EnKF experiment in order to maintain the greatest amount of realism.

Another set of sounding sensitivity experiments is conducted for the 3 June
Figure 3.15. Same as Fig 3.14, but data are valid at 0040 UTC.

Figure 3.16. Skew T-Log $p$ diagrams of the base-states used in three sounding sensitivity EnKF experiments of the 3 June 1999 storm. The environments are composed of data collected from 1) the 0000 UTC RUC analysis at the model grid point nearest the DOW deployment site with a moistening of the layer between 600 and 400 mb, 2) same as sounding 1 but the temperature profile between 800-600 mb and 925-825 mb is made more statically stable, and 3) the Weisman and Klemp (1982) idealized sounding. Hodographs for each environment are included in the top right of each panel. The gray lines in soundings 1 and 2 show the raw RUC $T_d$ profiles.
Figure 3.17. a–c) Vertical velocity (shaded), vertical vorticity (black contours; outermost is 0.003 s⁻¹, incremented by 0.002 s⁻¹), radar reflectivity computed from model precipitation fields (green contours; the outermost is 40 dBZe, incremented by 5 dBZe), and storm-relative horizontal velocity (vectored) at \( z = 750 \) m AGL at 0032 UTC for the 3 June 1999 sounding sensitivity EnKF experiments. d–f) Perturbation density potential temperature (shaded), vertical vorticity (black contours; same contour intervals as in panels a–c), computed radar reflectivity (green contours; same contour intervals as panels a–c), the \( w = 1 \) m s⁻¹ contour (gray contour), and horizontal vorticity (vectors). All plotted fields are ensemble averages of the model variables immediately after data assimilation has occurred. Data plotted in panels a and d are from the sensitivity experiment using sounding 1, panels b and e show data from the experiment using sounding 2, and panels c and f show data from the experiment using sounding 3 in Fig 3.16.

1999 storm. The following sources of data are used to compose the base-state environment in otherwise identical EnKF experiments: 1) \( \bar{T}, \bar{T}_d, \bar{u}, \) and \( \bar{v} \) profiles taken from the 00 UTC initialization of the RUC model at a horizontal grid point within a 40 km distance of the DOW deployment site (Fig 3.16a), 2) the same data from experiment 1, except that the \( T \) profile is made more stable directly above and below the boundary layer inversion in order to test the ability of the EnKF to sustain a storm in convectively-suppressive conditions (Fig 3.16b), and 3) the \( \bar{T} \) and \( \bar{T}_d \) profiles of the Weisman and Klemp (1982) sounding with the horizontal wind profile used in experiment 1 (Fig 3.16c). In addition to slightly increasing the stability of the \( \bar{T} \) profiles of soundings 1 and 2 where absolutely
unstable or neutral layers are found in the raw RUC data (described in section 3.2.2), we also increased the dew point temperature between 600 and 400 mb from those in the raw RUC environment (raw $T_d$ profile is shown with a gray line in Fig 3.16a,b). In idealized WRF simulations, artificially moistening the sounding in this layer produced the most stable supercell storm among a range of other RUC soundings in which the dew point profile was altered. Results from these three EnKF experiments are shown in Fig 3.17. All three experiments produce similar magnitudes of peak vertical vorticity and rear-flank downdraft, and similar rear-flank gust front organization (Fig 3.17a-c). The soundings with RUC-based thermodynamics (experiments 1 and 2) produce forward-flank cold pools with larger temperature deficits from the environment than in experiment 3, which uses an idealized thermodynamic profile (Fig 3.17d–f). Experiment 1 produces the most elongated forward-flank cold pool and precipitation echo in the downstream direction from the primary updraft. The reasons for these differences among the three experiments are not clear. The increased stability of the modified RUC sounding in experiment 2 appears to reduce the downstream spreading of the forward-flank cold pool and the intensity of the forward-flank gust front. It is possible that the reduction of this cold pool is due to the suppression of spurious convection along the forward-flank gust front, which limits the production of evaporatively chilled air far downstream of the main updraft. In the absence of an observed proximity sounding, the environment containing the 3 June 1999 storm is approximated as the RUC sounding from experiment 2 in the final EnKF experiment.

### 3.3.2 Sensitivity to choice of model microphysics

Some numerical simulation studies have noted that the choice of microphysics scheme can alter certain characteristics of modeled storms; in particular, the intensity of the low-level cold pool (e.g., Gilmore et al. 2004). The sensitivity of the EnKF results to the choice of WRF microphysics options are now considered because such sensitivities may be important to evaluating tornado maintenance.

Figure 3.18 shows low-level thermodynamic and kinematic fields produced in two EnKF experiments of the 5 June 2001 storm using two different microphysics schemes: 1) the Kessler (warm rain) scheme, and 2) the Lin et al. (1983; hereafter
LFO) ice microphysics scheme that includes snow, graupel, and cloud ice. The LFO scheme uses a Marshall-Palmer graupel intercept parameter of $4 \times 10^4 \text{ m}^{-4}$, a graupel density of 900 kg m$^{-3}$, and a rain intercept parameter of $8 \times 10^6 \text{ m}^{-4}$, as is used in Gilmore et al. (2004). Aside from some subtle differences in the peak magnitude of the temperature deficit, the strength of the rear-flank downdraft, the width of the simulated radar reflectivity field in the forward-flank precipitation, and the peak strength of the mesocyclone, the two experiments produce remarkably similar storms. The LFO scheme is used in the final 5 June 2001 experiment in order to maintain a greater amount of realism.

The thermodynamic and kinematic fields produced by the EnKF experiments of the 3 June 1999 case are more sensitive to variations in microphysics schemes than in the 5 June 2001 experiments. Figure 3.19 shows results from three 3 June 1999 experiments that use different microphysics schemes: 1) Kessler; 2) LFO with a graupel intercept parameter of $4 \times 10^4 \text{ m}^{-4}$, a graupel density of 900 kg m$^{-3}$, and a rain intercept parameter of $8 \times 10^6 \text{ m}^{-4}$; and 3) LFO with a graupel intercept parameter of $4 \times 10^3 \text{ m}^{-4}$, a graupel density of 900 kg m$^{-3}$, and a rain intercept parameter of $1 \times 10^6 \text{ m}^{-4}$. The graupel and rain intercept parameters are made smaller in experiment 3 to test the sensitivities of the EnKF results to the inclusion of large graupel and rain, because some large hail stones were observed in this storm.

The most obvious difference among these three experiments is the areal coverage and intensity of the forward-flank cold pool and precipitation fields; experiment 1 produces the coldest and widest forward-flank cold pool while experiment 3 produces the warmest forward-flank cold pool. In experiment 1, cold air in the forward-flank region leads the tornado by at least 30 kilometers and spreads the farthest distance southward, completely surrounding the mesocyclone at the end of the experiment (not shown). Experiment 1 also produces an exaggerated extension of warm inflow air directly east of the rear-flank gust front that extends northward into the forward-flank cold pool. This feature is lessened in the experiments with ice microphysics. The reasons for the differences in the areal coverage of the forward-flank cold pool is not entirely clear. However, it appears that the inclusion of ice decreases the dominance of the forward-flank downdraft and cold pool far downstream (east) of the tornado.
Figure 3.18. a–c) Vertical velocity (shaded), vertical vorticity (black contours; outermost is 0.003 s$^{-1}$, incremented by 0.002 s$^{-1}$), radar reflectivity computed from model precipitation fields (green contours; outermost is 40 dBZe, incremented by 5 dBZe), and storm-relative horizontal velocity (vectored) at $z = 750$ m AGL at 0030 UTC in the two 5 June 2001 microphysics sensitivity experiments. d–f) Perturbation density potential temperature (shaded), vertical vorticity (black contours; same contour intervals as in panels a–c), computed radar reflectivity (green contours; same contour intervals as panels a–c), the $w = 1$ m s$^{-1}$ contour (gray contour), and horizontal vorticity (vectors). All plotted fields are ensemble averages of the model fields immediately after data assimilation has occurred. Data plotted in panels a and c show data from the sensitivity experiment using Kessler microphysics, panels b and d show data from the experiment using Lin et al. (1983) ice microphysics with a graupel intercept parameter of $4 \times 10^4$ m$^{-4}$ and a rain intercept parameter of $8 \times 10^6$ m$^{-4}$.

In all three experiments, relatively cold air is located directly south of the vertical vorticity maximum, while warmer air is located directly to its west. The temperature contrast south through west of the mesocyclone is the greatest in experiment 3 and the smallest in experiment 1. The inclusion of ice and rain drops of increasing size seems to enhance features on the rear-flank of the storm. Based on the lack of surface thermodynamic observations, it is not clear which
0036 UTC 3 June 1999, $z = 750$ m AGL

Figure 3.19. a–c) Vertical velocity (shaded), vertical vorticity (black contours; outermost is 0.003 s$^{-1}$, incremented by 0.002 s$^{-1}$), radar reflectivity computed from model precipitation fields (green contours; outermost is 40 dB$Z_e$, incremented by 5 dB$Z_e$), and storm-relative horizontal velocity (vectored) at $z = 750$ m AGL at 0036 UTC for the 3 June 1999 microphysics sensitivity experiments. d–f) Perturbation density potential temperature (shaded), vertical vorticity (black contours; same contour intervals as in panels a–c), computed radar reflectivity (green contours; same contour intervals as panels a–c), the $w = 1$ m s$^{-1}$ contour (gray contour), and horizontal vorticity (vectors). All plotted fields are ensemble averages of the model variables immediately after data assimilation has occurred. Data plotted in panels a and d are from the EnKF experiment using Kessler microphysics, panels b and e show data from the experiment using LFO ice microphysics with Gilmore et al. (2004) intercept parameters, and panels c and f show data from the experiment using LFO ice microphysics with a graupel intercept parameter $= 4 \times 10^3$ m$^{-4}$ and a rain intercept parameter $= 1 \times 10^6$ m$^{-4}$.

of the LFO schemes produces the most realistic temperature structure. While both LFO experiments capture a qualitatively similar evolution of temperature and low-level kinematic structures on the rear-flank of the storm, the differences in the magnitudes of the analyzed model variables between the two LFO experiments highlight the inherent uncertainties in the EnKF solutions based on a reasonable choice of microphysics parameters. The LFO parameters of experiment 2 are used in the final 3 June 1999 EnKF experiment because the upward motion along the rear-flank gust front attached to the decaying tornado after 0044 UTC resembles
the dual-Doppler solution slightly more than that of experiment 3.
Case Background

This chapter details the lifecycles of each of the four tornadoes examined in this study in order to increase the reader's familiarity with them before their maintenance is assessed in chapter 5. The structure and evolution of some of the storms supporting each tornado are detailed in published or upcoming studies (Dowell et al. 2002c, Marquis et al. 2008, Wurman et al. 2010); however, to increase the reader's familiarity with each storm, a brief description of each also is provided in this chapter.

4.1 Single-Doppler observations of the tornadoes

Several past studies have proposed objective criteria for the automatic detection of mesocyclones and tornado vortex signatures in large quantities of radar data (e.g., Wieler 1986, Mitchell et al. 1998, Alexander and Wurman 2006). Many of these detection algorithms include some combination of measurements of: the gate-to-gate radial velocity differential, horizontal shear along a constant radar range (i.e., azimuthal shear), the horizontal extent of the azimuthal shear, and the depth and
duration of the rotation. In this study, tornado vortices are quantitatively defined and distinguished from the storm-scale circulation by 1) a difference between the peak inbound and outbound values of the radar radial velocity couplet in the tornado, $\Delta V_r > 30 \text{ m s}^{-1}$; 2) a distance between the inbound and outbound radial velocity maxima, $D < 2 \text{ km}$; and 3) an estimated vertical vorticity, $\zeta = 2\Delta V_r / D > 0.1 \text{ s}^{-1}$; all present throughout the lowest 1 km AGL for at least 2 minutes or 3 consecutive radar volumes. Figure 4.1 shows the evolution of $\Delta V_r$, $D$, $\zeta$, and circulation ($\Gamma = 0.5\pi D\Delta V_r$) of each tornado estimated with single-Doppler velocity data collected by the radar with the best temporal coverage in each deployment.

It is likely that the tornadoes are not completely resolved in the DOW data because of beam spreading. The azimuthal width of the radar beam ($\mu$) at the location of the tornadoes (also shown in Fig 4.1) ranges from 300 m to 40 m, depending on the distance between the radar and the center of the tornadoes. On average, the radar gate lengths are about 75 m. There are suggestions of multiple concentric scales of rotation with $D < 2 \text{ km}$ in at least two of the cases discussed in this study, 30 April 2000 and 5 June 2001. During certain observation periods (e.g., 2100–2109 UTC in the 30 April 2000 case), the temporal resolution (5 seconds) is sufficiently high to detail the evolution of these scales. However, the evolution of these scales is more ambiguous when observed with the temporal resolution employed in a typical three-dimensional volume scan (60–90 seconds; e.g., 2110–2130 UTC in the 30 April 2000 case and at all times in the 5 June 2001 case). Only the scales of rotation that are prominent in each near-surface radar sweep are shown in Fig 4.1.

Both tornadogenesis and tornado demise are observed only in one case, 5 June 2001; DOW deployments on the other three storms began after tornadogenesis. Therefore, data collected by the National Climatic Data Center, such as spotter reports and damage surveys, are used to supplement information regarding the intensity and duration of the tornadoes. Estimated tornado duration and intensity determined from the NCDC reports and the DOW data collected in each storm, are summarized in Table 4.1. Based on the information in Fig 4.1 and Table 4.1, the following is concluded about each of the four tornadoes:
Figure 4.1. Single-Doppler estimates of the time histories of $\Delta V_r$, $D$, $\zeta$, and $\Gamma$ for the a) 3 June 1999, b) 30 April 2000, c) 5 June 2001, and d) 22 May 2004 tornadoes. The observed azimuthal data spacing ($\mu$) is also provided. Data from only one radar is presented in each panel except for the 3 June 1999 case, where both DOW 3 (black) and DOW 2 (gray) data are shown to illustrate a more complete history of the tornado. The lifetime of each tornado is indicated in each panel with thick black arrows. Uncertainty in the lifetime of the tornadoes due to observation gaps is indicated with a dashed line. Gray data in panel b are from the first tornado and black data are from the second tornado observed in the 30 April 2000 case.
Table 4.1. A summary of the times of formation and dissipation, the duration, and the peak strength (Fujita scale rating or peak $\Delta V_r$) taken from the NCDC storm reports (top) and estimated from the single-Doppler DOW data (bottom) collected in each of the four storms.

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<th>duration</th>
<th>F-scale</th>
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<td>2127 UTC</td>
<td>2134 UTC</td>
<td>7 min</td>
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</tr>
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<td>0030 UTC</td>
<td>2 min</td>
<td>F0</td>
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<tr>
<td></td>
<td>0030 UTC</td>
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<td>&lt; 1 min</td>
<td>F0</td>
</tr>
<tr>
<td>22 May 2004</td>
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<td>2304 UTC</td>
<td>&lt; 1 min</td>
<td>F0</td>
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<th>dissipation</th>
<th>duration</th>
<th>peak $\Delta V_r$ (m s$^{-1}$)</th>
</tr>
</thead>
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<td>2116 UTC</td>
<td>$\geq$ 16 min</td>
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<td>0036 UTC</td>
<td>10 min</td>
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<tr>
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<td>before 2055 UTC</td>
<td>2312-2313 UTC</td>
<td>$\geq$ 16 min</td>
<td>95</td>
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</tbody>
</table>

1. **3 June 1999**: The first radar observations of this tornado are from DOW 3 at approximately 0016 UTC, after tornadogenesis has occurred (Fig 4.1a). At this time, $\Delta V_r \approx 65$ m s$^{-1}$ but increases to $\Delta V_r > 100$ m s$^{-1}$ by 0030 UTC. This intense wind speed is maintained until about 0043 UTC, when it begins to dramatically decrease until the tornado is no longer discernible on radar at about 0047 UTC. Circulation and core diameter slightly decrease between 0030 and 0043 UTC, indicating that the steady peak wind speed is likely due to a continuous contraction of angular momentum. However, after 0043 UTC, both of these metrics decrease rapidly until 0047 UTC. The NCDC report for this storm confirms the high intensity of this tornado, with F3 damage amounting to approximately 1 million USD. A tornadogenesis time of 0000 UTC is indicated in the NCDC report. However, there is no clear tornado present in the DOW3 Doppler velocity data at 0004 UTC, the final time of data collection in the first radar deployment. Therefore, tornadogenesis must have occurred sometime between 0005 and 0015 UTC, but the exact time is unknown (this uncertainty is indicated with the dashed portion of black arrow in Fig 4.1a). The time of tornado demise listed in the NCDC report agrees with DOW observations. Regardless of the precise time of tornadogenesis, DOW observations from 0016 to 0047 UTC indicate
a tornado duration of at least 30 minutes, with a possible lifetime of up to 40 minutes. This is the most intense, and presumably the longest-lived tornado analyzed in this study.

2. **30 April 2000:** Thoroughly discussed by Marquis et al. (2008), the tornadic activity observed in this case is rather complex. Two concentric circulations (∼ 0.5-km-wide and ∼ 2-km-wide) with similar peak wind speeds ($\Delta V_r = 55 \text{ m s}^{-1}$) are present at the beginning of the DOW deployment. Only the inner (0.5-km-wide) circulation is documented in Fig 4.1b (plotted in gray) because the outer (2-km-wide) circulation does not meet our $\zeta$ or $D$ criteria to be considered a tornado during the short period in which it is observed. The inner circulation is seen in the first mobile sweeps at 2055 UTC at 200 m AGL, indicating that tornadogenesis took place before that time. By 2103 UTC, one intermediate scale of rotation is seen near the ground with an initially stronger peak wind speed than either of the preceding vortices ($\Delta V_r \approx 70 \text{ m s}^{-1}$). After 2103 UTC, the tornado widens to a diameter of about 2-km. Its circulation greatly increases as the core diameter increases and the $\Delta V_r$ remains somewhat steady. At about 2108 UTC, a multiple-vortex structure is present with discrete vorticity maxima on the western half of the circulation (Marquis et al. 2008). One of these vorticity maxima develops into another 0.5-km-wide tornado at approximately 2110 UTC, with $\Delta V_r \approx 40 \text{ m s}^{-1}$ (black lines in Fig 4.1b). By 2116 UTC, shortly after the development of this new tornado, the expanding original tornado is no longer clearly detected in the radar data, indicating a total lifetime of the first tornado of at least 16 minutes. The circulation of the new tornado remains fairly steady; however, the peak wind speed decreases until the end of dual-Doppler analysis (2117 UTC), when the tornadic velocity couplet is barely discernible in the single-Doppler data. There is a short gap in the single-Doppler data between 2119 and 2121 UTC. After the data gap, a similarly-sized tornado with slightly stronger winds ($\Delta V_r \approx 52 \text{ m s}^{-1}$) is present until it dissipates at around 2130 UTC. Because of the data gap, it is not clear if the final tornado is the same as that observed between 2111 and 2116 UTC. However, it seems likely that the radar velocity couplets observed at 2119 and 2121 UTC are from the same tornado, given the similar values of $D$ and $\Gamma$. Under this
assumption, the second tornado observed in this case lasts for approximately 20 minutes. NCDC tornado reports on this day were provided by the DOW team at the time of the storm intercept; therefore, no further information is available beyond the DOW observations and a damage survey, which revealed no significant damage (Table 4.1).

3. 5 June 2001: A tornadic Doppler velocity couplet is observed to form between approximately 0026 and 0027 UTC, and is no longer evident in low-level radar data by approximately 0036 UTC (Fig 4.1c). Two relative maxima of $\Delta V_r$ and $\Gamma$ are observed, one at approximately 0029 UTC and a second at about 0032 UTC. The temporal resolution of the data collected in this storm intercept is too coarse to determine if these two maxima belong to two separate tornadoes that are widening during their 3- to 5-minute lifetimes, or if only one tornado is present and undergoing a pulsed strengthening and narrowing during the observation period. Two reports of tornadoes are indicated in the NCDC storm report archive, one at 0028 and another at 0030 UTC (Table 4.1). These reports include 1- and 2-minute-long tornadoes, both with F0 intensity. However, it is not clear how well the DOW-observed tornado(es) correspond to these reports, because the tornadic area of the storm becomes visibly obscured by rain at the time of tornadogenesis and also because the tornado(es) occurred over open country, yielding no discernible damage swath.

4. 22 May 2004: Thoroughly detailed in an upcoming paper by Wurman et al. (2010), the DOW-resolved structure of this tornado appears to be less complex than that of the 30 April 2000 or 5 June 2001 cases. Tornadogenesis is not observed in this case. During most of the observed portion of the lifetime of the tornado, the circulation remains steady, while the core diameter shrinks and $\Delta V_r$ increases (Fig 4.1d), indicative of a continuous contraction of angular momentum. The complete dissipation of the tornado is not observed by the DOWs; however, a rapid decrease in circulation and $\Delta V_r$ is seen in the final few data volumes from both radars (2310-2311 UTC), suggesting that tornado dissipation is imminent at 2311 UTC. A visible condensation funnel was visible from the location of DOW2 from 2301–2307
UTC, while the tornado is strengthening (in terms of $\Delta V_r$ and $\zeta$). A rope condensation funnel was surrounded by precipitation at approximately 2311 UTC, consistent with the imminent dissipation of the tornado. The NCDC report indicates very short-lived tornado (duration < 1 min), which conflicts with the duration of at least 16 minutes observed by the DOW radars (Table 4.1).

4.2 Overall storm structure

Figure 4.2a, b, e, f shows radar reflectivity and radial velocity data collected at $z \leq 1.5$ km AGL by the closest WSR-88D radar during the DOW intercept of each storm. The mesocyclones are indicated in each panel with a black ring, and the deployment positions of each DOW radar are indicated with dots. Figure 4.2c,d,g,h shows DOW reflectivity and single-Doppler velocity observations collected in the sector scans shown in Fig 4.2a, b, e, f at $z \leq 600$ m AGL. The tornadoes and rear-flank gust fronts are shown in Fig 4.2c, d, g, h with black rings and black lines. From the animations of these radar data, the following can be summarized about the overall structure of each parent supercell.

1. **3 June 1999:** The WSR-88D reflectivity field exhibits a classic supercell structure, with a clear hook echo and a Doppler velocity couplet with $\Delta V_r = 50$ m s$^{-1}$, discernible from a range of almost 120 km (Fig 4.2a). There are no neighboring convective cells present within at least one hour of the DOW intercept period. This storm had a history of cyclic mesocyclogenesis; two low-level mesocyclones, two gust fronts, and two hook echoes that are attached to the same precipitation core are observed during the first DOW deployment a few minutes prior to tornadogenesis (not shown). Only one mesocyclone, gust front, and hook echo are observed during the tornadic period of the storm (Fig 4.2b), but mesocyclogenesis is occurring during the demise of the observed tornado after 0045 UTC.

2. **30 April 2000:** This storm exhibits a high-precipitation supercell structure (Doswell and Burgess 1993), the mesocyclone is completely surrounded by WSR-88D radar reflectivity $> 30$ dBZe (Fig 4.2b). A hook echo is difficult
Figure 4.2. Radar reflectivity and single-Doppler velocity data collected by a nearby WSR-88D (a, b, e, f) and by one of the DOW radars (c, d, g, h) during the intercept of the 3 June 1999 storm (upper-left), the 30 April 2000 storm (upper-right), the 5 June 2001 storm (lower-left), and the 22 May 2004 storm (lower-right). The height (AGL) and time (UTC) that each sweep intersects the mesocyclone is indicated in each panel. The positions of the DOW2 and DOW3 radars during the dual-Doppler deployments are indicated in panels a, b, c, e, f with red and green dots. The positions of the mesocyclones (in panels a, b, e, f) and the tornadoes (in panels c, d, g, h) are indicated with black rings in each panel. Blue contours in panels a, b, e, and f indicate the sector scans of the DOW data displayed in panels c, d, g, h.
to discern in both WSR-88D and DOW reflectivity fields because of the high-precipitation structure and convection that is developing along the rear-flank gust front south-southwest of the tornado. The observed supercell is embedded within a cluster of less-organized convective cells, with its closest neighbors to the north and east. The targeted storm does not appear to collide or merge with any of these convective elements during the observation period. Marquis et al. (2008) show that the DOW-observed primary rear-flank gust front structure is highly occluded during the observation period (Fig 4.2d).

3. **5 June 2001**: This storm developed within a northwest-to-southeast oriented line of cells moving toward the east-northeast (Fig 4.2e). The targeted storm has a close neighbor located directly to its southwest at the beginning of the DOW deployment (2350 UTC). Mesocyclogenesis and tornadogenesis occurs along a gust front at approximately 0015 UTC, shortly after the two storm cells merge (Dowell et al. 2002c); therefore, it is possible that storm interactions may have resulted in the formation of the tornado. After approximately 0020 UTC, the DOW-observed portion of the storm develops a low-level mesocyclone and gust front structure typical of a rapidly evolving supercell storm (Fig 4.2g). During the tornadic period, the storm exhibits a high-precipitation supercell structure, visibly obscuring the tornado with rain.

4. **22 May 2004**: The targeted supercell is surrounded by several disorganized convective cells immediately to its northwest and east (Fig 4.2f). During most of the DOW observation period, and for several minutes prior to it, the supercell remains isolated from its neighbors. However, near the end of the DOW deployment (2311 UTC), the tornadic storm collides with a cell directly to its east. It is possible that the collision of these storms leads to tornado demise, but it cannot be confirmed with the available DOW data. Similar to the 30 April 2000 and 5 June 2001 storms, this storm exhibits a high-precipitation structure. A hook echo is discernible in the both the WSR-88D and DOW reflectivity fields, but is visually translucent. A highly occluded gust front structure, similar to that observed in the 30 April 2000
storm, is observed with DOW data (Fig 4.2h).

In summary, the data examined in this study span tornadoes of varying peak intensity and duration. Therefore, a comparison of the storms producing each of these tornadoes may yield results that are valuable to understanding why some tornadoes last for longer periods of time than others.
In this chapter, the maintenance of the 3 June 1999, 30 April 2000, 5 June 2001, and 22 May 2004 tornadoes is evaluated by comparing the evolutions of the dual-Doppler-retrieved and EnKF-retrieved kinematic structures and the EnKF-retrieved thermodynamic structures of the storms to the trends of tornado intensity measured with the single-Doppler data. In particular, in this chapter, we discuss the possible role that the evolution of each of the following storm properties plays in tornado maintenance:

1. the contraction of the mesocyclone-scale circulation by the radial inflow,

2. the position of the tornadoes relative to the rear-flank downdraft and gust fronts,

3. the motion of the tornadoes relative to the midlevel updraft,

4. the sources of vertical vorticity to the tornado, and,

5. the temperature of the air surrounding the tornadoes and the structure of the baroclinically-generated horizontal vortex lines produced in the outflow
air mass.

The results of these analyses are compared among the four cases and to past observations of tornadic supercells to determine the commonality of the processes involved with tornado maintenance across several storms. Additionally, processes promoting tornado maintenance are compared to those presumably promoting tornadogenesis to determine if the sources of convergence and vertical vorticity to the tornadoes change throughout their lifetimes.

5.1 Evolution of kinematic structures of the storms

The first goal in our study of tornado maintenance is to relate trends of tornado intensity to changes in the overall storm structure, such as the mesocyclone-scale circulation, the gust fronts, and the rear-flank downdraft.

5.1.1 Mesocyclone-scale circulation

Figure 5.1a–d shows the relative circulation (in a horizontal circuit),

\[ \Gamma = \int_x \int_y \zeta \, dx \, dy, \]

and azimuthally-averaged radial velocity (with respect to the center of the tornado) as functions of time and radius from the center of the vertical vorticity maximum during the dual-Doppler observation period of each of the four storms at \( z = 300 \) m AGL. The trend of \( \Delta V_r \) observed by low-level single-Doppler radar sweeps are shown in Fig 5.1e-h. In the 3 June 1999 storm, the tornado maintains a nearly steady peak wind speed up to approximately 0043 UTC despite a slowly decreasing circulation at all radii (Fig 5.1a,e). Tornado strength decreases after 0043 UTC, when the azimuthally-averaged radial velocity switches from inbound to outbound, indicating that angular momentum is no longer being advected inward. The EnKF data show a similar result (Fig 5.2a,c), although the magnitudes of azimuthally-averaged radial velocities are smaller than in the dual-Doppler data, presumably because the model resolution is more than three times coarser than that of the dual-Doppler wind syntheses. In the 30 April 2000 storm (Fig 5.1b,f),
Figure 5.1. a-d) Circulation (shaded) and azimuthally-averaged radial velocity (contours; m s$^{-1}$) as a function of time and radius from the axis of rotation, both measured at $z = 300$ m AGL in the dual-Doppler analyses for each of the four cases. Dashed (negative) contours in panels a-d indicate inbound radial velocities (relative to the center of rotation) and solid (positive) contours indicate outbound velocities. e-h) $\Delta V_r$ measured using unsmoothed single-Doppler data valid between $z = 100-200$ m AGL for each of the tornadoes (from Fig 4.1.)

The tornado maintains a steady $\Delta V_r$ from approximately 2105 to 2110 UTC, while the storm-scale circulation is steady or even slightly increasing at most radii. During this time, strong inbound radial velocities are observed at radii > 3 km, but much weaker radial velocities are observed closer to the axis of rotation. The tornado(es) weaken after 2110 UTC, when storm-scale circulation decreases and the azimuthally-averaged radial velocity is directed outward at most radii. In the 5 June 2001 storm (Fig 5.1c,g and Fig 5.2b,d; dual-Doppler and EnKF data, respectively), the circulation increases at most radii up to 0030 UTC, after which it begins to decrease. The azimuthally-averaged radial velocity is becoming stronger in the outbound direction during the entire observation period. The diverging winds that develops immediately after the formation of this tornado possibly explains its short lifetime. In the 22 May 2004 storm (Fig 5.1d,h), the wind speed in the tornado increases while the circulation increases slightly at radii < 3 km and the azimuthally-averaged radial velocity is directed inward at radii < 2 km. After 2307 UTC, the peak $\Delta V_r$ associated with the tornado decreases, but the cir-
Figure 5.2. Same as Fig 5.1, except that circulation and azimuthally-averaged radial velocities are measured using the ensemble-mean kinematic fields from the 3 June 1999 (a) and 5 June 2001 (b) EnKF analyses at $z = 250$ m AGL.

circulation remains constant or even slightly increases. However, the inward-directed radial velocities at radii $< 1.5$ km weaken and become directed outward in the last dual-Doppler volume, indicating a reversal of the inward advection of angular momentum as the tornado weakens.

In summary, the peak tangential wind speed in each of the four tornadoes is maintained while mean inward motion is present, particularly within about a 2 km range from the tornado, inward-advected mesocyclone-scale or tornado-cyclone-scale angular momentum. Tornadoes weaken when radial inflow weakens, such that the inward advection of angular momentum ceases or reverses. Interestingly, the two tornadoes with the highest observed wind speeds (3 June 1999, 22 May 2004) occur in the storms with the weakest mesocyclones, whereas the weakest and shortest-lived tornado (5 June 2001) occurs within the storm with the strongest mesocyclone. Because the maintenance of the tornadoes appears to be closely tied to the ability of the mean radial flow to advect angular momentum inward toward
the axis of rotation, it seems that simply the presence of a strong mesocyclone is not sufficient for the production or sustenance of strong tornadoes.

### 5.1.2 Primary rear-flank downdraft and gust fronts

A more complete analysis of the storms requires the use of the full fields of motion, such that asymmetric components of the flow can be assessed. Using these data, the evolution of the strength and motion of the gust fronts and the outflow air on the rear flank of the storms are examined in order to determine if tornadoes weaken by the wrapping of horizontally divergent and potentially cold outflow air around the low-level circulation, as proposed in past studies (e.g., Lemon and Doswell 1979). Figure 5.3 shows the ground-relative locations of the gust fronts and the tornadoes at several times in each of the four cases. The gust fronts are located by following the maximum in the horizontal velocity gradient tensor (Stonitsch and Markowski 2007) at $z = 500$ m using the 3 June 1999 EnKF data (Fig 5.3a) and at $z = 300$ m AGL using the dual-Doppler data for the other three storms (the dual-Doppler data are preferred in this analysis because of their high resolution; however, the EnKF data are used in the 3 June 1999 analysis because they provide a greater spatial coverage than do the dual-Doppler wind syntheses in that case). The values of $\Delta V_r$, as measured by single-Doppler data, near the ground for each of the tornadoes also are shown in Fig 5.3. The trends of the average horizontal convergence along the gust front that makes contact with each tornado (measured by following the maximum of horizontal convergence along the the wind shift) and the maximum value of horizontal divergence near the ground surrounding the four tornadoes are provided in Fig 5.4. These calculations are performed between a 1–3 km horizontal distance from the center of each tornado in order to consider the magnitudes the of horizontal convergence or divergence near the tornado that are not locally enhanced by its presence. In other words, we assume that the value of horizontal convergence along a gust front that intersects each tornado is due to processes associated with the parent storm rather than the tornado.

In the 3 June 1999 storm (Fig 5.3a), the large $\Delta V_r$ within the tornado is sustained while the rear-flank gust front maintains a constant shape, spiraling outward from the tornado. From the brief DOW3 observations that are available
prior to the beginning of the second radar deployment, it appears that this gust front configuration is maintained for several minutes as the tornado is strengthening (between 0015–0030 UTC in Fig 4.1a). Between approximately 0036 and 0041 UTC, the average convergence along the gust front and the peak winds in the tornado are mostly unchanged (Fig 5.4a). After 0040 UTC, the configuration of the rear-flank gust front deforms from a spiral shape to a nearly north-south
Figure 5.4. Time trends of the unsmoothed single-Doppler $\Delta V_r$ of the tornadoes between $z = 100$-200 m AGL (black lines with dots), the maximum value of divergence at $z = 300$ m AGL between a 1-3 km range of the axis of rotation (black lines), and the average value of convergence along the gust front that directly contacts the resolved circulation between 1–3 km range of the tornado (gray lines) for each of the four cases. The solid black and gray lines indicate measurements taken from the dual-Doppler data in each case, the dashed black and gray lines indicate measurements taken from the EnKF data (panels a and c).

oriented line with the weakening tornado located at its northernmost tip. During this transformation, the strength of the dual-Doppler-retrieved convergence along the gust front decreases (Fig 5.4a). The EnKF-retrieved convergence along the gust front remains steady between 0040 and 0044 UTC, a possible consequence of the inferior horizontal resolution of the model (which is about three times coarser than that of the dual-Doppler syntheses). However, the convergence along the gust front intersecting the tornado weakens substantially by 0044 UTC in the EnKF analysis, shortly before the tornado dissipates. The strength of the low-level divergence surrounding the tornado is weakening throughout the dual-Doppler and EnKF analysis periods, although, in the dual-Doppler analyses, the magnitude of the peak divergence decreases more rapidly than in the EnKF analyses when the tornado is rapidly weakening (Fig 5.4a). This decreasing magnitude of divergence
suggests a weakening of the rear-flank downdraft near the tornado, such that a wrapping of divergent outflow air around the tornado does not appear to be the cause of dissipation. At 0046 UTC, the tornado has lost its connection to the convergence along the gust front and dissipates within the next 2 minutes.

In the 30 April 2000 storm (Fig 5.3b), the primary rear-flank gust front encircles the tornado, leading it by about 4 km in the north and east directions in all dual-Doppler analyses. The tornado is not directly connected to the primary rear-flank gust front. Throughout the dual-Doppler observation period, the $\Delta V_r$ measured in the tornado decreases. However, complicated changes in vortex structure also occur during the observation period, and a new tornado forms at approximately 2110 UTC that increases in strength after dual-Doppler data collection. Therefore, factors other than a connection to the primary rear-flank gust front appear to be important to tornado maintenance in this case.

In the 5 June 2001 storm (Fig 5.3c), the outflow air wraps around the low-level circulation quite rapidly, causing the tornado to become separated from the occlusion point between the rear-flank and forward-flank gust fronts by a distance of about 10 km (distance along the gust front) within about 6 minutes. The peak low-level divergence in the rear-flank downdraft increases throughout most of the lifetime of the tornado, consistent with a strengthening of the downdraft that is wrapping around the low-level circulation (Fig 5.4c). The EnKF-measured peak magnitude of the low-level divergence in the rear-flank downdraft is similar to that measured in the dual-Doppler fields, and decreases slightly during the final stage of tornado dissipation after the period of dual-Doppler coverage. In the final dual-Doppler volume (0034 UTC), the occluded gust front has completely encircled the tornado, indicating that outflow air has wrapped all the way around the low-level circulation. The tornado dissipates at about 0036 UTC owing to a disconnection of the tornado from the convergence along the primary rear-flank gust front. This observation is consistent with the decreased contraction of angular momentum at the time of tornado dissipation (Figs 5.1c and 5.2b).

The tornado-relative position of the rear-flank gust front in the 22 May 2004 storm (Fig 5.3d) is quite similar to that in the 30 April 2000 storm in many respects: 1) the tornado is surrounded by outflow air on all sides throughout the entire observed portion of its lifetime, 2) the primary gust front maintains a nearly
constant distance from the tornado in time (except after 2307 UTC, when the gust front travels northward relative to the tornado about 2 km), and 3) the convergence along the primary gust front does not clearly contact the tornado throughout the dual-Doppler observation period. Despite these observations, the tornado strengthens between 2259 and 2307 UTC rather than weakening. Therefore, as in the 30 April 2000 case, sources of vertical vorticity that maintain the tornado must originate from somewhere other than along the primary rear-flank gust front.

5.1.3 Secondary rear-flank downdrafts and gust fronts

Figure 5.5 shows an example of the gust front structures in the 30 April 2000 and 22 May 2004 cases using the dual-Doppler derived horizontal convergence fields (Fig 5.5b,d), and in the 5 June 2001 case using the EnKF ensemble-mean vertical motion fields (Fig 5.5c). In these cases, bands of horizontal convergence and updraft are present within the outflow air, each leading an area of stronger horizontal winds and low-level divergence or downdraft than are present elsewhere in the outflow air. There is a suggestion of a similar band of low-level horizontal convergence spiraling southward from the tornado within the outflow air of the 3 June 1999 case in the single-Doppler data (Fig 5.5a); however, convergence and upward motion along this feature cannot be unambiguously retrieved because of a lack of dual-Doppler or reliable EnKF analyses at the time in which it is observed. Marquis et al. (2008) and an upcoming article by Wurman et al. (2010) identify these bands of horizontal convergence within the outflow air of the 30 April 2000 and 22 May 2004 storms as possible secondary rear-flank gust fronts that lead newly developed surges of rear-flank downdraft.

The motion of these secondary rear-flank gust fronts relative to the tornadoes in the 30 April 2000, 5 June 2001, and 22 May 2004 cases is illustrated with dual-Doppler data in Figure 5.6a-c. The development of the secondary gust front in the 5 June 2001 case also is illustrated with a sequence of low-level EnKF kinematic data in Fig 5.6d. The evolution of the secondary gust fronts observed in the 30 April 2000 and 22 May 2004 cases (Fig 5.6a,c) are quite similar; the bands of convergence pivot around the tornadoes, contacting them at low-levels, consistent with the tornadogenesis stage in the Lemon and Doswell (1979) conceptual model.
Figure 5.5. a) Radar radial velocity data in the 0.5° elevation sweep at approximately 0020 UTC on 3 June 1999. The position of the radar is indicated with an ‘R’ and the center of the tornado is indicated with a ‘T’. b) Dual-Doppler horizontal convergence (shaded), storm-relative horizontal winds (vectors), and vertical vorticity (contours; outermost contour is 0.02 s⁻¹, incremented by 0.025 s⁻¹) valid at 2108 UTC on 30 April 2000. c) EnKF ensemble-mean vertical velocity (shaded), storm-relative horizontal winds (vectors), and vertical vorticity (contours; outermost contour is 0.02 s⁻¹, incremented by 0.01 s⁻¹) valid at z = 500 m AGL at 0034 UTC on 5 June 2001. d) Same as panel b, but at 2304 UTC on 22 May 2004. The primary and possible secondary rear-flank gust fronts are indicated in each panel with solid and dashed black lines, respectively.

of a supercell storm containing only one rear-flank gust front. The convergence along each of these secondary gust fronts is observed to increase when they have the greatest forward acceleration (between 2108 and 2115 UTC in the 30 April 2000 case, Fig 5.4b; and after 2307 UTC in the 22 May 2004 case, Fig 5.4d); therefore, the convergence that is in contact with the tornadoes is enhanced as the
secondary gust fronts wrap around the low-level circulation. Accompanying the quickly advancing secondary gust fronts is an increase in the magnitude of low-level divergence directly behind them (Figs 5.4b,d). Because of a lack of continuous dual-Doppler data during their dissipation, it cannot be stated definitively that the presence of a strongly divergent secondary surge of outflow that completely surrounds these two tornadoes ultimately causes their demise. However, in both cases, the rapid evolution of the secondary gust front just before the tornado rapidly weakens, and in the case of the 30 April 2000 case, when the tornado is undergoing a drastic change in vortex structure (Marquis et al. 2008), suggests

Figure 5.6. a-c) Same as Fig 5.3b-d except that the positions of the possible secondary rear-flank gust fronts are emphasized with bold lines. d) Same as Fig 5.5c, but at 0030, 0032, 0034, 0036, 0038, and 0040 UTC on 5 June 2001. The primary and possible secondary gust fronts are marked with solid and dashed black lines, respectively. The location of the remnant mesocylone at times after tornado dissipation is marked with an ‘X’ at 0038 and 0040 UTC in panel d.
that the secondary rear-flank downdraft surges likely play a role in the dissipation of the tornado.

The secondary band of upward motion observed in the 5 June 2001 case evolves in a slightly different manner than in the 30 April 2000 and 22 May 2004 cases. The band of upward motion in the 5 June 2001 case initially forms southeast of the tornado with subsequent northward development, rather than pivoting around the low-level circulation (0028-0032 UTC in Figs 5.6b,d). The updraft band continues to develop northward toward the occluded portion of the preceding rear-flank gust front about 3 km east of the tornado (0032-0040 UTC in Figs 5.6b,d), never contacting the peak of vertical vorticity. During this evolution, the new downdraft surrounds the tornado, and the low-level circulation dissipates while no longer connected to the convergence along the primary gust front, which has a nearly constant magnitude farther away from the tornado during this time (Fig 5.4c).

The evolution of the rear-flank outflow in some of these storms suggests that the tornadoes can be maintained by possible secondary rear-flank gust fronts while separated from the primary rear-flank gust front. Tornado dissipation may occur once the secondary surge of outflow air wraps around the low-level circulation, severing its connection to the low-level convergence. The evolution of the gust front surrounding the strong and long-lived 3 June 1999 tornado suggests that it does not dissipate owing to the influence of surging outflow air that is wrapping around it. Alternative explanations for the demise of this tornado are hypothesized next. In addition to controlling the supply of horizontal convergence to the tornado, thereby affecting the magnitudes of stretching of existing vorticity, the gust fronts and rear-flank downdraft also must supply the mature tornado with vertical vorticity in order for it to be maintained; otherwise, the existing vertical vorticity at low-levels would be exhausted as it is advected upward. The role that these features play in the creation of vertical vorticity are discussed later in this chapter using trajectory analysis and an examination of vortex lines.
Figure 5.7. Peak vertical vorticity (black contours), and the positive maxima of the advection of vertical vorticity by the ground-relative horizontal wind (blue contours) and stretching of vertical vorticity (red contours) within a 2 km horizontal distance of the tornado at several times using the dual-Doppler analyses in the a) 3 June 1999, b) 30 April 2000, c) 5 June 2001, and d) 22 May 2004 cases. While the same contour intervals are not used across all panels, the same contour intervals are used for each of the terms in each panel. Values of tilting and vertical advection are too small to show up in the panels based on the contour intervals chosen. The time associated with each tornado position is labeled in UTC (hhmm:ss) and the height AGL of each analysis is indicated in the bottom right corner of each panel. The tornado track is traced with a gray line in each panel.

5.2 The Dowell and Bluestein maintenance mechanism

As mentioned in the literature review, Dowell and Bluestein (2002b; DB02b) is the only study known to the author to formulate a detailed hypothesis predicting the conditions necessary for tornado maintenance based on the evolution of a super-cell storm. In this section, the applicability of the DB02b tornado maintenance mechanism is tested on our four tornadic supercells.
A first step in assessing the applicability of the DB02b maintenance mechanism on our storms is to determine what factors influence the motion of our tornadoes. Figure 5.7 shows the ground-relative motion of the low-level dual-Doppler vertical vorticity maximum containing each of the four tornadoes. Also contoured in Fig 5.7 are the relative maxima of the horizontal advection (by the ground-relative wind), vertical advection, stretching, and tilting terms from equation 2.1 within a 2 km horizontal distance of the peak vertical vorticity. Near the ground, tilting and vertical advection of vertical vorticity are much weaker than the horizontal advection and stretching terms (although, they may still be important in developing near-ground vertical vorticity, particularly in the absence of pre-existing vertical vorticity). At almost all analysis times, the magnitude of horizontal advection is larger than that of stretching. Figure 5.8 shows the relative strengths of the maxima of the terms in equation 2.1 within 2 km horizontal distance of the dual-Doppler vertical vorticity maxima between \( z = 250 \) m and 1.8 km AGL. While the magnitude of the tilting, stretching, and vertical advection terms become larger with height, the horizontal advection term is the largest of all four terms. A comparison of the magnitudes of these terms using the EnKF data in the 3 June 1999 and 5 June 2001 cases yields similar results, although near-surface (\( z = 250 \) m) stretching is at least as strong as horizontal advection (Fig. 5.9). The magnitudes of these terms in our storms are similar to those shown by DB02b, who found that while all four of these terms were of equal magnitude at \( z > 3 \) km AGL, in the lowest 1 km, the horizontal advection and stretching terms were the largest.

From Fig 5.7, it can be seen that the resolved vertical vorticity maxima most closely follow the maxima in horizontal advection. Similar observations are made at \( z \leq 1.8 \) km using dual-Doppler data and at \( z < 3 \) km using the EnKF results (not shown). The observation that the resolved vertical vorticity maxima most closely follow the nearby maxima of the horizontal advection term indicates that the winds in which the tornadoes are embedded are essentially steering them, consistent with the observations of DB02b. Changes in the ground-relative wind field containing the low-level circulation appear to be responsible for some interesting changes in the motion of the tornadoes. For example, in the 30 April 2000 and 5 June 2001 storms (Fig 5.7b,c), sudden changes in the tornado motion at 2108 and 0029 UTC (respectively) coincide with secondary surges of outflow air nearby. However,
Figure 5.8. Time tendency of the maximum of the: advection of vertical vorticity by the ground-relative horizontal wind (solid blue contours), advection of vertical vorticity by the storm-relative horizontal wind (dashed blue contours), stretching of vertical vorticity (red contours), tilting of horizontal vorticity into the vertical (yellow contours), and vertical advection of vertical vorticity (green contours) within a 2 km horizontal distance of the tornado at $z = 150$ m, 450 m, 1050 m, and 1800 m AGL using the 30 April 2000, 5 June 2001, and 22 May 2004 dual-Doppler analyses. The heights of each panel are labeled on the right.

Figure 5.9. Same as Fig 5.8 except at $z = 250$ m, 1000 m, 2000 m, 3500 m, and 5000 m AGL using the ensemble-mean EnKF analyses of the 5 June 2001 and 3 June 1999 cases.
these tornado motions deviate from the mean only temporarily, returning to their original directions after the secondary outflow surge begins to wrap around the center of rotation. The 22 May 2004 tornado begins to dissipate in the last few dual-Doppler analyses as strong northerly outflow air in the rear-flank downdraft draws closer to the center of rotation, causing the tornado to take on a stronger southward component of motion (Fig 5.7d). The 3 June 1999 tornado maintains a steady strength and northeastward direction of motion from 0030 to 0043 UTC, but rapidly weakens after 0043 UTC, when its direction of motion turns toward the northwest (Fig 5.7a).

A key component of the DB02b maintenance mechanism pertains to the position of the tornado relative to the primary updraft; a tornado may be displaced from the strongest low-level convergence, the potentially buoyant inflow, and the most steady source of vertical vorticity if it is separated from the primary updraft. The vertical depth of the dual-Doppler data in all four cases is insufficient to determine the ground-relative location of the midlevel updraft. Instead, the location of the midlevel updraft is estimated in the 3 June 1999 and 5 June 2001 data assimilation experiments. Figure 5.10 shows a sequence of the ground-relative locations of the near-surface mesocyclone containing the tornado and the midlevel updraft in the 3 June 1999 and 5 June 2001 cases. During the observed steady-intensity portion of its lifecycle (before approximately 0042 UTC; Fig 5.10a-b), the 3 June 1999 tornado is located beneath the northwestern corner of the primary midlevel updraft. At 0040 UTC (Fig 5.10b), just before the tornado begins to rapidly weaken, the portion of the updraft above the tornado weakens while the remaining midlevel updraft southeast of the tornado continues to move toward the east-northeast. At 0044 UTC (Fig 5.10c), the rapidly weakening tornado is located at least 3 km northwest of the $w = 5 \text{ m s}^{-1}$ contour, indicating an increasing separation distance (in a horizontal plane) from the midlevel updraft. This separation distance increases to about 8 km at the time when the tornado is barely discernible on radar (0048 UTC; Fig 5.10d). The tilt of the mesocyclone at 0044 UTC (illustrated with rings in Fig 5.10c) is consistent with a tornado that is vertically sheared by the strong southeasterly inflow at low-levels (Fig 5.7a). The 5 June 2001 tornado moves relative to the center of the midlevel updraft as the entire storm moves to the east (Fig 5.10e-h). However, contrary to the 3 June 1999 case,
Figure 5.10. Ensemble-mean vertical velocity at $z = 5$ km AGL (gray dashed contours of $w = 5, 10, 15$ m s$^{-1}$), vertical vorticity at $z = 500$ m AGL (thin black contours; outermost contour is 0.01 s$^{-1}$, incremented by 0.005 s$^{-1}$), the position of the surface gust fronts (traced with bold gray lines), and the surface track of the tornado (thick black line) at four times in the 3 June 1999 (left) and 5 June 2001 (right) EnKF analyses. The locations of the mesocyclones at $z = 1.5$ km, 3 km, and 5 km AGL are shown in panels c and g with black rings. A description of the tendency of the intensity of the tornadoes in each panel is provided on the right. The ‘M’ in panels b-d indicates the location of a newly developing mesocyclone. The black dashed line in panels e-h indicates the path of the remnant mesocyclone after tornado dissipation.
the tornado in the 5 June 2001 storm remains underneath the midlevel updraft throughout its entire lifecycle. Similar to the 3 June 1999 case, the mesocyclone in this storm also tilts significantly from a vertical axis as it is weakening (Fig 5.10g). However, in this case, the near-surface mesocyclone is located southeast of the midlevel mesocyclone, consistent with the presence of a tornado that is vertically sheared by the strong northwesterly outflow observed at low-levels.

The DB02b maintenance mechanism appears to apply to the 3 June 1999 case because the tornado weakens as it moves away from its parent updraft. At the time of tornado dissipation, the inflow to the midlevel updraft rises from directly beneath it, along a newly-formed rear-flank gust front upon which a new mesocyclone develops (Fig 5.10b-d). Therefore, it appears that the area directly beneath the updraft is a good location for the development and intensification of vertical vorticity at this time. It is possible that the 3 June 1999 tornado would not have dissipated if the balance between the low-level inflow and outflow had been maintained beyond 0042 UTC. The 5 June 2001 tornado dissipates underneath the midlevel updraft. At the time of tornado dissipation, the midlevel updraft is fueled by inflow that rises along the slanted surface of the primary rear-flank gust front that intersects the ground at least 5 km east of the mesocyclone. The tilt of the updraft is illustrated in Fig 5.18. Based on these observations, it appears that having a tornado located beneath the midlevel updraft may not be a sufficient condition for maintenance in all storms, particularly in those with strongly surging outflow winds.

5.3 Trajectory analysis

A valuable tool for determining the sources of vertical vorticity to each of the tornadoes is trajectory analysis, in which air parcels entering the tornadoes can be traced backward in time to determine their origins and detail their history of vertical vorticity. Trajectories are computed using a fourth-order Runge-Kutta scheme with 20-second time steps. Temporal interpolation includes a term that translates the velocity field between analysis times, minimizing errors due to storm motion, particularly in areas containing large velocity gradients. The dual-Doppler data collected in the present supercells are not sufficient to definitively determine the
sources of vertical vorticity to the tornadoes throughout all stages of their lifecycle owing to the short collection periods (< 15 minutes) and shallow depths ($z_{\text{top}} < 2$ km). Marquis et al. (2008) showed the extent of trajectory analysis that is possible in the 30 April 2000 case. Despite the brief amount of three-dimensional data available, certain valuable information can be deduced from the trajectory analysis: 1) inflow air from the ambient environment containing streamwise horizontal vorticity does not enter directly into the tornado at low-levels, but rather, it rises to the top of the data domain along the rear-flank gust front north and east of the tornado; 2) air flow with streamwise vorticity within the observed portion of the forward-flank of the storm does not flow directly into the tornado at low-levels, but rather, it rises to the top of the data domain along the occluded portion of the primary rear-flank gust front north of the tornado; 3) air parcels within the tornado descended from the top of the available data within about 1 km of the axis of rotation; and 4) many parcels within the secondary outflow surge south of the tornado ascend along the secondary gust front as they rotate around the mesocyclone. These and similar results from limited trajectories calculated in the 22 May 2004 storm are consistent with the idea that the highly occluded pool of outflow air is preventing the tornado from directly ingesting potentially buoyant environmental air. Instead, near the ground, the tornado is composed of outflow located behind two gust fronts, but is maintained for several minutes regardless.

Trajectories in the 5 June 2001 and 3 June 1999 storms are calculated using the EnKF analyses, which provide a greater coverage of three-dimensional kinematic fields than are available in our dual-Doppler data sets. In the 5 June 2001 case, parcels in a grid centered on the mesocyclone at $z = 750$ m AGL are traced backward at two times: 1) when the tornado has reached its peak strength (0032 UTC), and 2) the time marking tornado dissipation (0036 UTC). In both sets of trajectory calculations, the final backward integration time is 0020 UTC, which is the first trustworthy analysis in the 5 June 2001 EnKF experiment after the model spin-up period. Figure 5.11 compares the origins of parcels entering the circulation at these two times. Parcels entering the low-level circulation when the tornado is at its peak strength (red dots) and when it is dissipating (green dots) consist of a mix of cool outflow air and warmer environmental air; therefore, there was not a significant change in the source regions of air feeding into the tornado during
its development and its decay. Similar source regions are identified by trajectory analysis during the tornadogenesis stage using dual-Doppler data from this case (Dowell et al. 2002c). At both times, parcels entering the low-level circulation are experiencing a weak tilting of horizontal vorticity into the vertical followed by a period of stretching as they reside in upward motion along the rear-flank gust front (Fig 5.12b,c,e,f). Once acquiring significant vertical vorticity (Fig 5.12a,d), many
Figure 5.12. Vertical vorticity (top), stretching (middle), and tilting of horizontal vorticity into the vertical (bottom) observed along the red parcel trajectories (a-c) and the green parcel trajectories (d-f) shown in Fig 5.11. The gray line in each panel follows the observed vorticity, stretching, and tilting along the parcels indicated with stars in Fig 5.11.

Parcels experience oscillatory signals of stretching and tilting as they encounter upward and downward motion while orbiting the axis of rotation (e.g., the parcel that resides at the center of low-level circulation at 0032 and 0036 denoted with gray lines in Fig 5.12 and stars in Fig 5.11). The parcel that is located at the center of the circulation at $z = 750$ m AGL at 0032 UTC, when the tornado is strongest...
(gray lines in Fig 5.12a-c), experiences negative stretching starting at about 0030 UTC, suggesting that a downdraft near the axis of rotation will soon reduce the peak vertical vorticity. Interestingly, an unusually large value of tilting along this trajectory path offsets the loss of vertical vorticity due to compression, consistent with a steady history of vertical vorticity for this parcel. The parcel that is located at the center of the circulation at $z = 750$ m AGL at 0036 UTC, when the tornado is weakening (gray lines in Fig 5.12d-f), experiences strong negative stretching (and negligible tilting) starting at 0032 UTC that remains negative through 0036 UTC. A few parcels entering the tornado from the ambient environment near the gust front southeast of the tornado originate from heights much greater than all of the other parcels entering the tornado from the inflow (e.g., the green parcel located at $z = 2.55$ km in Fig 5.11a). These unusual trajectories may be a result of a discontinuity in the EnKF kinematic fields at the edge of the radar data coverage, which some of the parcels encounter en route to the tornado.

Figure 5.13a shows the origins of parcels entering the low-level mesocyclone of the 3 June 1999 storm while the tornado is maintaining a steady strength (0030 UTC). These parcels are tracked by calculating streamlines using the EnKF data valid at 0030 UTC because dual-Doppler data are not available at this time and because the EnKF results that are available before this time are produced using synthetic DOW observations (described in chapter 3.2). The streamlines are presumed similar to trajectories because the structure of the storm appears to be mostly unchanged between 0016 and 0030 UTC (described in chapters 3.2 and 4). Figure 5.13b,c shows the origins of parcels entering the tornado after it has begun to dissipate (at 0044 UTC) by calculating trajectories using the EnKF data between 0026 and 0044 UTC. Contrary to the parcel trajectories calculated in the 5 June 2001 storm, there is a notable difference in the origin of the air surrounding the tornado at these two times. While the tornado is maintaining a steady strength, air entering the low-level circulation travels through the forward-flank cold pool very close to the ground and either, 1) ascends into the mesocyclone as it encounters the rear-flank gust front immediately northeast of the tornado, or 2) wraps around the low-level circulation within the rear-flank downdraft region (Fig 5.13a). Certain parcels descend from the rear-flank region south and southwest of the tornado and take sharp turns into the low-level circulation. It is believed
Figure 5.13. Ensemble-mean $\theta'_\rho$ (shaded), vertical velocity (thin black contours; solid contours are 2, 4, 6, 8 m s$^{-1}$, dashed contours are -2, -4, -6, -8 m s$^{-1}$), and vertical vorticity (thick black contours; outermost is 0.015 s$^{-1}$, incremented by 0.01 s$^{-1}$) at $z = 750$ m AGL at 0030 UTC (a), 0044 UTC (b), and 0026 UTC (c) using the 3 June 1999 EnKF analyses. The surface gust fronts are traced with thick black lines. The location of the tornado is highlighted in each panel with a yellow ‘T’. The paths traveled by parcels located in a grid surrounding the steady tornado at $z = 750$ m AGL at 0030 UTC are shown in a storm-relative frame with gray lines in panel a. The locations of these parcels at 0012 UTC are indicated with gray dots and their heights (km AGL) are shown. The paths traveled by parcels located in a grid surrounding the dissipating tornado at $z = 750$ m AGL at 0044 UTC (gray dots in panel b) are shown in a ground-relative frame with gray lines in panel c. The locations of these parcels at 0026 UTC are indicated with gray dots and their heights (km AGL) are shown in panel c.
that these unusual parcel paths may be a result of a discontinuity in the EnKF kinematic fields at the edge of the radar data coverage, similar to the 5 June 2001 case. Parcels that orbit the circulation close to its axis of rotation acquire vertical vorticity through tilting and stretching along the rear-flank gust front a short distance northeast of the tornado (Fig 5.14a-c). While a few parcels entering the dissipating tornado follow similar trajectories as those entering the steady tornado, a new stream of air with origins from $z > 300$ m AGL in the ambient environment travels into the low-level circulation relatively uninfluenced by the forward-flank
outflow (Fig 5.13b-c). These trajectories are consistent with a shift of the direction of
the winds surrounding the low-level circulation toward the direction of the environmental inflow during tornado dissipation (Fig 5.7a). Despite the presence of the environmental air, the tornado rapidly dissipates as the low-level convergence along the rear-flank gust front weakens, causing the inflow parcels to experience less tilting and stretching (Fig 5.14d-f).

5.4 Temperature of the rear-flank outflow

Markowski et al. (2002) found that supercells with rear-flank outflow that is not significantly colder than the environment produced long-lived and strong tornadoes. Motivated by these observations, Markowski et al. (2003) showed in idealized simulations that relatively warm outflow allows for a greater low-level buoyancy and a greater contraction of angular momentum at low-levels than colder outflow, which contains a greater negative buoyancy and centrifugal force. In their study, the temperature of the outflow air is hypothesized to be important to tornadoogenesis. This section explores the effects of the evolution of the outflow temperature on the maintenance of mature tornadoes. While no in situ thermodynamic data were collected during the DOW storm intercepts, the EnKF experiments provide estimates of these unobserved variables in the 3 June 1999 and 5 June 2001 cases.

Figure 5.15 shows the evolution of the ensemble-mean $\theta'_{\rho}$ field averaged within a 3 km radius of the vertical vorticity maximum between $z = 0$ and 3 km AGL for the 3 June 1999 and 5 June 2001 storms. In the 3 June 1999 case, the temperature of the air surrounding the tornado is slowly increasing between 0026 and 0038 UTC (Fig 5.15a) while the tornado is maintaining a fairly steady peak wind speed (Fig 5.15b). The air containing the low-level circulation begins to cool as the tornado rapidly weakens and dissipates. Figure 5.16a-d shows the evolution of the $\theta'_{\rho}$ and vertical motion fields in horizontal planes at $z = 750$ m AGL in the 3 June 1999 storm. From this sequence and from Fig 5.10, we see that the low-level warming shown in Fig 5.15a occurs while the position of the tornado relative to the main updraft and the surface gust front is constant (Fig 5.10a-b). The cooling of the air surrounding the tornado after 0040 UTC occurs while the tornado travels into the forward-flank cold pool away from the midlevel updraft. Because the temperature
Figure 5.15. Ensemble-mean $\theta'_{ρ}$ averaged within a 3 km horizontal radius of the vertical vorticity maximum as a function of time and height between $z = 0$ and 3 km AGL (panels a and c). Panels b and d show the time tendencies of the unsmoothed single-Doppler $\Delta V_r$ for the tornadoes (black lines with dots), the maximum magnitude of the horizontal divergence at $z = 500$ m AGL within a 3 km radius of the vertical vorticity maximum (black lines), and the magnitude of the horizontal updraft-relative wind averaged within a 3 km radius of the vertical vorticity maximum of the 3 June 1999 storm (gray line in panel b).

of the air surrounding the low-level circulation is warming as the tornado maintains a steady strength, and the temperature during the dissipation of the tornado is not colder than at certain times during its mature stage, it is not clear that the buoyancy of the air surrounding the tornado directly affects its maintenance in this case. Instead, the warming of the outflow air may be symptomatic of a weakening rear-flank downdraft, which is consistent with a decrease in the magnitude of the peak divergence and an increase in the updraft relative wind (causing a shift in its direction toward the low-level inflow) within a 3 km range of the tornado at $z = 500$ m (Fig 5.15b).

The average $\theta'_{ρ}$ of the air surrounding the low-level circulation in the 5 June 2001 storm increases throughout the mature phase of the tornado (Fig 5.15c-d). This change in the outflow temperature occurs because the air comprising the secondary surge of outflow that develops shortly after tornadogenesis is warmer than the preceding pool of outflow, shown in Fig 5.16e-h. Owing to the uncertainty in the
Figure 5.16. Ensemble-mean $\theta'_\rho$ (shaded) and vertical velocity (thin black contours; solid contours are 4, 6, 8 m s$^{-1}$, dashed contours are -4, -6, -8 m s$^{-1}$) at $z = 750$ m AGL in the 3 June 1999 storm (left) and in the 5 June 2001 storm (right). Surface gust fronts are traced with black lines. The location and $\Delta V_r$ (unsmoothed single-Doppler) of the tornadoes are shown with green circles in each panel. An ‘X’ in panel h indicates the location of the remnant mesoscale circulation after tornado dissipation. An ‘M’ in panels c-d indicate the location of a new mesocyclone.
Figure 5.17. a-c) Ensemble-mean $\theta'_\rho$ (shaded), horizontal vorticity (vectors) and vertical motion (thin black contours; solid contours are $w = 4, 8, 12, 16 \text{ m s}^{-1}$, dashed contours are $w = -4, -8, -12, -16 \text{ m s}^{-1}$) at $z = 3 \text{ km AGL}$ (panels a,c) and $z = 750 \text{ m AGL}$ (b), and vertical vorticity (thick black contours; outermost is 0.015 s$^{-1}$, incremented by 0.01 s$^{-1}$) at $z = 750 \text{ m AGL}$ in the 5 June 2001 storm. Surface gust fronts are traced with thick black lines in each panel. The horizontal ground-relative positions of parcels that pass through the warm secondary downdraft at $z = 750 \text{ m AGL}$ at 0034 UTC (b), tracked by forward- and backward-integrated trajectory calculations, are shown with green dots in each panel. The gray line in panel a highlights a horizontal vortex line segment surrounding the negatively buoyant downdraft. Panel d shows several skew $T$-log $p$ diagrams of vertical thermodynamic profiles collected at several model grid points in the outflow air southwest of the tornadogenesis site at 0024 UTC before the development of the secondary downdraft (blue lines) and the thermodynamic profiles along several of the parcels descending in the warm secondary downdraft between 0020 and 0034 UTC (red lines).
EnKF-retrieved pressure and the diabatic heating variables, all processes affecting the vertical forcing of parcels in these data cannot be determined; therefore, the exact causes of the warming in the secondary downdraft are unknown. Trajectory calculations of certain parcels that pass through this warm pocket of air at $z = 750$ m AGL at 0034 UTC, shown in Fig 5.17a-c, are observed to descend within the rear-flank downdraft from altitudes of up to $z = 3.5$ km AGL at 0020 UTC. Figure 5.17d shows the temperature and dew point temperature history in the form of skew $T$-log $p$ diagrams for some of these parcels as they descend within the rear-flank downdraft (red lines). Temperature and moisture profiles within the cool outflow at 0024 UTC, prior to the development of the warm secondary downdraft, also are shown in Fig 5.17d (blue lines). Several of these outflow profiles are shown to indicate the spatial variability of the temperature profiles within the ensemble-mean analysis. Figure 5.17d indicates that the descending parcels are negatively buoyant aloft, but become positively buoyant below 700-750 mb. This change in parcel buoyancy suggests the possibility of heatburst dynamics as a cause of the warm secondary downdraft. In such events, negatively buoyant air that descends into an existing pool of cool air overshoots its static equilibrium level because of its large downward momentum. The descending air warms after it overshoots is equilibrium level, dry adiabatically if the air becomes unsaturated, producing a pocket of warm air near the ground (e.g., Johnson 1983). In the 5 June 2001 storm, cold outflow generated from a previous downdraft that occurs well before tornadogenesis provides the temperature inversion upon which the negatively buoyant air from the secondary downdraft impinges. Figure 5.18 shows a vertical cross section of the ensemble-mean $\theta'_p$ and vertical motion fields along the bold gray line in Fig 5.16g. The secondary rear-flank downdraft surge penetrates into the low-level cold pool resulting in a warm pocket of air in the lowest 2 km AGL, consistent with the expected behavior of a heat burst. The forward-integrated trajectories of parcels within the warm outflow at 0034 UTC show that the band of ascent along the possible secondary gust front is composed of the positively buoyant air near the ground. However, as shown in Fig 5.6d and Fig 5.16f-h, the band of upward motion leading the downdraft surge does not directly contact the tornado. Therefore, this analysis shows that even though the air surrounding the tornado is relatively warm, the parcels are not ascending
Figure 5.18. Vertical cross section of ensemble-mean $\theta'_p$ (shaded), storm-relative wind (vectors), and vertical motion (thin black contours; solid contours are $w = 2, 4, 6, 8, 10$ m s$^{-1}$, dashed contours are $w = -2, -4, -6, -8, -10$ m s$^{-1}$) along the thick gray line in Fig 5.16g. Arrows below the horizontal axis indicate where the primary gust front (right arrow) and possible secondary gust front (left arrow) intersect the ground along the cross section.

within the tornado, thereby not amplifying vertical vorticity through stretching.

While this scenario is complex, it is not necessarily unique to the 5 June 2001 storm, offering the possibility that this warm secondary rear-flank downdraft could occur in other cases. Indeed, some past studies detailing in situ observations of surface temperature have indicated secondary outflow surges that are warmer than preceding ones (e.g. Markowski et al. 2002, Finley and Lee 2008). However, three-dimensional kinematic and thermodynamic data are not available in their cases to determine the cause of the warm outflow air. The EnKF thermodynamic fields presented herein contain caveats, the foremost of which is that such data are retrieved by the assimilation of only single-Doppler velocity observations. However, the EnKF thermodynamic fields also contain certain aspects that support their validity. For example, the warm secondary surge of outflow air in the 5 June 2001 case bares consistencies with past in situ observations of supercell storms and heatbursts, and a smooth evolution of the ensemble-mean temperature field in both the prior and posterior analyses.
Figure 5.19. Dual-Doppler vertical vorticity (shaded), horizontal vorticity (vectors), and vertical motion (contours; outermost contour is 1 m s$^{-1}$, incremented by 2 m s$^{-1}$) at $z = 400$ m AGL at about 2100 UTC on 22 May 2004. The primary and secondary rear-flank gust fronts are traced with thick dashed lines, and the location of the tornado is marked with a ‘T’. Vortex lines that pass through the locations marked with dots are traced with black lines. A projection of the vortex lines into the $y$-$z$ plane is shown on the left.

5.5 Vortex line analysis

Straka et al. (2007) proposed that baroclinically-generated vortex rings surrounding the rear-flank downdraft can be tilted in such a way along the rear-flank gust front that they form arch shapes, generating cyclonic vertical vorticity at the ground where a tornado is typically seen (the details of this mechanism are reviewed in chapter 2.3). This mechanism for the generation of vertical vorticity is suggested to be important to tornadogenesis, but it is not known if it plays a significant role in the maintenance of a mature tornado. Therefore, the applicability of this mechanism to the maintenance of our tornadoes is now tested by drawing vortex lines that pass through organized areas of vertical vorticity surrounding the tornado and gust fronts near the ground in the dual-Doppler and the EnKF analyses.
Figure 5.19 shows three-dimensional vortex lines that pass through an area of vertical vorticity at $z = 400$ m AGL along the edge of the primary rear-flank gust front about 4 km north-northeast of the 22 May 2004 tornado. These vortex lines arch upward to $z < 1$ km and southeastward into an area of anticyclonic vertical vorticity along the gust front about 5 km east of the tornado. This pattern is consistent with the vortex line arching mechanism proposed by Straka et al. (2007). In this case, the positive vertical vorticity along the arch is separated from the tornado by several kilometers. This observation suggests that the tilting of horizontal vorticity by the updraft along the primary rear-flank gust front does not directly supply the tornado with vertical vorticity; however, not enough dual-Doppler data are available in this case to determine if vertical vorticity gained along parcel trajectories passing through this area later enters the tornado. Dual-Doppler data are not available during and just after tornadogenesis, precluding a knowledge of the importance of this vortex line arching mechanism when the tornado presumably is attached to the primary rear-flank gust front. Due to the shallow depth of the dual-Doppler volumes, vortex lines that pass through areas of strong vertical vorticity closer to the tornado and the secondary rear-flank gust front are too short to be useful for the mapping of the vorticity fields.

Figure 5.20 shows three-dimensional vortex lines near the primary and secondary gust fronts in the 30 April 2000 case. Similar to the 22 May 2004 case, certain vortex lines, when projected into a horizontal plane, parallel the primary rear-flank gust front (green and yellow vortex lines in Fig 5.20). The cyclonic vertical vorticity generated by the tilting of the vortex lines along the occluded portion of the primary rear-flank gust front west-northwest of the tornado (yellow lines in Fig 5.20) appears to be separated from the tornado by a horizontal distance of at least 3 km. This observation suggests that, similar to the 22 May 2004 storm, the vortex line arching mechanism occurring at the updraft along the primary rear-flank gust front may not directly aid in the maintenance of the mature tornado. The trajectory calculations in this case show that parcels located at radii greater than about 3 km from the tornado either orbit the low-level circulation at a constant radius or spiral outward (Marquis et al. 2008). Therefore, these trajectories, although limited owing to the shallow and brief nature of the dual-Doppler data, further suggest that vertical vorticity produced along the primary
Figure 5.20. Same as Fig 5.19, but using the dual-Doppler data valid at 2113:43 UTC on 30 April 2000. The gust fronts are traced using thick black lines. A projection of the vortex lines into the $y$-$z$ and $x$-$z$ planes are shown on the left and the bottom.

gust front may not pass into the tornado. Vortex lines passing through the area of vertical vorticity within a 2 km distance NW-SW of the tornado are not useful for mapping the three-dimensional vorticity fields because of the shallowness of the data domain. Therefore, it cannot be stated if vortex lines that pass through this area form arches in the outflow air between the primary and secondary rear-flank gust fronts, possibly by the tilting and vertical advection of horizontal vortex lines just ahead of the surging secondary gust front.

Vortex line arches also are found along the southwestern side of the secondary rear-flank gust front (red and blue lines in Fig 5.20). These arches pass through the area of positive vertical vorticity east-northeast of the tornado at the $z = 400$ m AGL level, some within a 2 km horizontal distance from the tornado. The limited trajectories calculated in this area show many parcels orbiting the low-level circulation at a nearly constant radius from its center, suggesting that vertical vorticity
generated along the secondary gust front remains near the tornado. Therefore, while it cannot be determined definitively due to the shallow depth of the dual-Doppler data, it is possible that the secondary gust front may assist with the maintenance of the mature tornado by way of the Straka et al. (2007) vortex line tilting mechanism. The orientation of the vortex lines behind the secondary gust front is consistent with a secondary outflow surge that is colder than the previous one. Markowski et al. (2003) hypothesize that the temperature of the outflow was an important factor in tornadogenesis. However, in the presence of a mature tornado, intense jets of vertical velocity that arise owing to its interaction with the ground (described in chapter 2.2) may be sufficient to lift quite negatively buoyant outflow parcels to their levels of free convection. Therefore, the present data suggests that relatively cold rear-flank outflow surges (relative to previous outflow surges) may assist with tornado maintenance through the injection of additional favorable vortex lines despite the fact that more negatively buoyant air may be entering the tornado.

Figure 5.21 shows the vortex lines that pass through areas of positive vertical vorticity around the tornado and along the primary rear-flank gust front at $z = 500$ m AGL in the EnKF analyses at 0028 and 0036 UTC on 5 June 2001 (these times correspond to immediately after tornadogenesis and when the tornado is dissipating). At both times, vortex lines passing through the vertical vorticity present along the primary rear-flank gust front arch southwestward, reaching heights up to 3 and 4 km AGL (at 0028 and 0036 UTC, respectively), and continue through areas of negative vertical vorticity at $z = 500$ m AGL about 8–10 km south-southwest of the tornado. Vortex lines passing through these areas in the dual-Doppler data suggest similar arch patterns before they end at the top of the relatively shallow domain. At both times, these vortex lines arch over the low-level secondary rear-flank downdraft, indicated with bold black lines in Fig 5.21.

Vortex lines that pass through the low-level mesocyclone (blue lines and some green lines in Fig 5.21a, red lines in Fig 5.21b) also arch through the negative vertical vorticity southwest of the tornado at $z = 500$ m AGL, and when projected into the horizontal plane, roughly parallel the band of vertical motion leading the secondary downdraft. These vortex line arches are much shallower than those associated with the primary rear-flank gust front, and are shallower at the time
Figure 5.21. Same as Fig 5.20, but using the EnKF ensemble-mean analyses valid at 0028 UTC (a) and 0036 UTC (b) on 5 June 2001. A projection of the vortex lines into the $y-z$ and $x-z$ planes are shown on the right of panel b and the bottom of panel a.
of tornado dissipation than they are immediately after it has formed. The latter observation is consistent with an updraft along the gust front that is diminished immediately near the tornado.

Interestingly, the shallow vortex line arches associated with the secondary outflow surge are pointing in a direction opposite to those expected with a relatively warm downdraft. As mentioned in section 5.4, parcels that are rapidly descending south of the tornado are negatively buoyant before warming. Shown in Fig 5.17a-b, negatively buoyant parcels at $z \approx 3$ km AGL that are descending in the secondary downdraft are encircled by horizontal vorticity with an orientation that is consistent with the idealized simulation of Straka et al. (2007). However, the northerly component of vorticity increases along the trajectories of the parcels descending in the secondary downdraft despite the reversed horizontal buoyancy gradient (Fig 5.17b). The easterly component of vorticity along these trajectories increases while the parcels descend mostly on the northern periphery of warm downdraft, but the northerly component increases principally due to the tilting of negative vertical vorticity by positive $\partial v/\partial z$. Therefore, though the reversed buoyancy gradient plays a role in the modification of horizontal vorticity, the tilting of vertical vorticity into the -$\hat{y}$ direction also contributes to the shape of the three-dimensional vortex lines.

Figure 5.22 details the evolution of the vertical vorticity field near the ground along the gust fronts in the 5 June 2001 case. The sheet of vertical vorticity present along the primary rear-flank gust front, presumably formed by the Straka et al. (2007) vortex line arching mechanism and possibly shearing instability, has formed discrete vortices that are advected along the gust front toward the tornado. Between approximately 0026 and 0029 UTC (Fig 5.22a-c), the developing tornado appears to merge with vertical vorticity traveling toward it along the occluded portion of the primary gust front. The strengthening of the tornado at this time is consistent with the stretching of the newly-ingested vertical vorticity maxima by low-level convergence along the gust front. Between 0029 and 0032 UTC, the tornado weakens (Fig 5.22h) while the secondary outflow surge rushes southeastward nearby (illustrated by the motion of the maximum tornado-relative wind speed 3 km southeast of the tornado in Fig 5.22c-f). While weak vertical vorticity is being generated along the leading edge of the secondary surge (traced
Figure 5.22. a-g) A sequence of vertical vorticity (contours; $\zeta = 0.015, 0.025, 0.035, 0.05, 0.07, 0.09, 0.11, 0.13, 0.15, 0.17, 0.19 \text{ s}^{-1}$), tornado-relative horizontal wind (vectors), and the magnitude of the tornado-relative horizontal wind (shaded) from the dual-Doppler analyses in the 5 June 2001 storm. The location of the tornado is indicated with a ‘T’ at each time. The primary gust front is traced with a thick black contour (uncertainty in the location of the gust front is indicated with long dashed lines), and the possible secondary gust front is traced with a short dashed line. Individual vorticity maxima along the primary gust front are traced in time using letters A through H. An A’, A'', and A”’ indicate the possible division of the vorticity maxima labeled A in panel a or the formation of new vorticity maxima in the vicinity of the A maxima. The time tendency of the unsmoothed single-Doppler $\Delta V_r$ and circulation (from Fig 4.1c) is reproduced in panel h.

with a dashed black line in Fig 5.22), the strong outflow winds southwest of the tornado direct the vertical vorticity maxima along the occluded portion of the primary gust front into the outflow air several kilometers south-southeast of the center of rotation, preventing its ingestion into the tornado. Due to the highly distorted nature of the primary gust front near the tornado after 0030 UTC, it is unclear if the brief secondary maximum in tornado strength that occurs at approximately 0032 UTC (Fig 5.22h) is due to a renewed ingestion of vertical vorticity present along the primary rear-flank gust front after the passage of the secondary outflow surge. There is some indication that vertical vorticity generated along the leading edge of the secondary outflow surge may briefly assist with the maintenance of this tornado late in its lifetime (along the dashed line in Fig 5.22g).
Summary and Conclusions

A combination of high-resolution dual-Doppler wind syntheses and ensemble Kalman filter data assimilation experiments performed using Doppler on Wheels data collected in four tornadic supercell intercepts was examined in order to determine how changes in storm structure influence the maintenance of tornadoes.

Each of the four tornadoes was maintained while connected to the low-level convergence along rear-flank gust fronts. The strongest and presumably longest-lived tornado (3 June 1999) maintained contact with the convergence along the primary rear-flank gust front (i.e., the boundary between the cool outflow and the ambient environment) for a considerable time. Dissipation occurs as the convergence weakens at the tornado when strong low-level inflow advects the tornado relative to the motion of the midlevel updraft. After its formation, the weakest and shortest-lived tornado (5 June 2001) maintains a connection to the primary rear-flank gust front for only a few minutes. It dissipates underneath the midlevel updraft when it loses its connection to the primary rear-flank gust front after rapidly surging outflow completely wraps around it. The two tornadoes of intermediate intensity and duration (30 April 2000 and 22 May 2004) persist while
Figure 6.1. Schematic illustration of the evolution of the primary and secondary gust fronts observed in the 30 April 2000 and 22 May 2004 storms (top) and in the 5 June 2001 storm (bottom). Black lines trace the peak in low-level convergence along the gust fronts. Gust fronts are shown at three times in each case ($t_1 < t_2 < t_3$). Dashed lines indicate uncertainty in the placement of the gust fronts due to their weakening or development. The gray circles with ‘$\zeta$’ represent the location of the tornadoes at each time.

... completely surrounded by outflow air at low-levels, a condition that is expected to promote tornado dissipation (Lemon and Doswell 1979). These tornadoes are connected to lines of convergence (possibly secondary rear-flank gust fronts; Wurman et al. 2007, Marquis et al. 2008) after they have been orphaned from the primary rear-flank gust front. These tornadoes presumably dissipate after they lose their connection to the low-level convergence along the secondary gust fronts, which occurs after the secondary outflow surge completely wraps around the low-level circulation. The development of a secondary band of upward motion is observed within the outflow air of the 5 June 2001 storm, but it does not make contact with the tornado; therefore, it does not assist with the maintenance of the tornado.
after it has lost its connection to the primary rear-flank gust front. Figure 6.1 schematically illustrates the development and motion of the multiple gust fronts in the 30 April 2000, 5 June 2001, and 22 May 2004 storms.

A comparison of the relative placement of the gust fronts to the tornadoes in all four cases suggests that a long-lived connection to the primary rear-flank gust front is ideal for the maintenance of strong and long-lived tornadoes. While connections to secondary gust fronts appear to assist with the maintenance of the 30 April 2000 and 22 May 2004 tornadoes, the fact that these boundaries quickly spiral around the low-level circulation suggests that they only briefly assist with tornado maintenance. The observation that the peak winds in these tornadoes are not sustained as long as those in the 3 June 1999 case is consistent with this speculation. Regardless of the maximum intensity achieved by these two tornadoes, the fact that they are connected to the secondary gust fronts indicates that heterogeneity in the outflow air may be an important factor in tornado maintenance.

The 3 June 1999 tornado dissipates while separated from the midlevel updraft by several kilometers (in a horizontal plane). During the time of tornado dissipation, inflow is drawn straight up into the midlevel updraft along a new rear-flank gust front containing a developing low-level mesocyclone. This observation suggests that an area beneath the updraft is the best location for tornado maintenance in this storm. However, in the 5 June 2001 case, a tilted updraft draws upon near-surface air that is located several kilometers to the east of the mesocyclone. As a result of this tilted ascent, the midlevel updraft is sustained while the convergence at the tornado directly beneath it erodes. Therefore, the tornado dissipates even though it is located beneath the midlevel updraft. This observation suggests that having a tornado located beneath the primary updraft is not always a sufficient condition for maintenance, particularly in storms with strongly surging outflow.

Three-dimensional vortex line patterns found along the primary rear-flank gust fronts in the 30 April 2000, 5 June 2001, and 22 May 2004 storms are consistent with the tilting of horizontal vorticity generated along the temperature gradient at the leading edge of relatively cool outflow air. The vortex lines resemble the arched vortex line patterns shown in idealized simulations (Straka et al. 2007) and comparatively coarser (relative to DOW resolution) airborne radar observations (Markowski et al. 2008). However, in the 30 April 2000 and 22 May 2004 cases,
the tilting of horizontal vortex lines along the primary rear-flank gust front does not appear to directly supply the mature tornadoes with vertical vorticity because of the highly occluded nature of the outflow. Vortex line arches observed along the secondary rear-flank gust front are more closely connected to the 30 April 2000 tornado, suggesting that a surge of outflow air that is cooler than the previous one supplies it with vertical vorticity (a schematic diagram of this scenario is shown in Fig 6.2). A similar vortex line pattern is observed in the 5 June 2001 storm despite the fact that the data assimilation experiment retrieves a warm secondary rear-flank downdraft. This unexpected result appears to be reconciled partly by the fact that the downdraft is negatively buoyant prior to warming dry adiabatically as it descends through the cold pool, but also because tilting of vertical vorticity significantly contributes to the production of horizontal vorticity.

The evolution of the thermodynamic data retrieved in the 3 June 1999 and 5 June 2001 data assimilation experiments suggests that a warming of the outflow air surrounding the tornadoes does not directly aid their maintenance. In both cases, the warming of the outflow air appears to be symptomatic of changes in the rear-flank downdraft that adversely affect tornado maintenance. In the 3 June 1999 case, the warming is consistent with a weakening rear-flank downdraft that leads to the imbalance between inflow and outflow winds surrounding the tornado, which causes its motion to deviate from the midlevel updraft as in Dowell and Bluestein.
In the 5 June 2001 storm, the warm secondary surge of outflow, whose formation and structure is consistent with past observations of heat bursts, does not directly aid tornado maintenance because the positively buoyant outflow air is not ascending within the tornado.

The concept that a secondary surge of relatively cold outflow air may support tornado maintenance by tilting and stretching baroclinically-generated horizontal vorticity along the secondary gust front implies that the intense vertical jet found within a mature tornado is sufficient to lift the negatively buoyant parcels to their levels of free convection (if they have one). Therefore, though the presence of cold outflow air may not support the contraction of angular momentum at low-levels necessary to form intense or long-lived tornadoes (Markowski et al. 2002, 2003), the generation of new surges of relatively cold outflow air after tornadogenesis actually may assist with tornado maintenance in these other ways.

Though this work has compared certain attributes of four supercell storms, a more complete understanding of tornado maintenance cannot be attained until a larger sample of cases is examined. Future studies should examine high-resolution kinematic and thermodynamic data collected in many other storms (such as those collected in the VORTEX2 field experiment) in order to confirm the hypotheses of tornado maintenance posed in this study. Additionally, future work should attempt to diagnose the cause of multiple rear-flank downdraft surges in supercell storms, which could not be performed in some of our cases owing to the limited amount of radar data available. A knowledge of these topics will likely increase the skill of forecasts and nowcasts of tornado longevity.

Finally, other studies have demonstrated that individual posterior EnKF analyses assimilating single-Doppler data produce a realistic result in convective storms (e.g., Snyder and Zhang 2003, Zhang et al. 2004, Dowell et al. 2004). However, this study has shown that more complex analyses of the EnKF fields, such as the calculation of parcel trajectories, appear to show dynamic consistency with past observations and simulations of tornadic supercells as long as an abundant amount of radar data are available to assimilate. Even though the time history of the thermodynamic fields along the parcel trajectories in the 5 June 2001 case is consistent with past observations of a heat burst, a lack of in situ thermodynamic data precludes the confirmation of the realism of the EnKF result. Future stud-
ies that attempt to deduce realistic dynamic relationships between updated model variables in EnKF data assimilation experiments should test the sensitivities of the results to the assimilation of a complete set of in situ thermodynamic observations at the surface and aloft, if available.


Davies-Jones, R. 2006: Tornadogenesis in supercell storms - what we know and what we don’t know. Preprints Symposium on the Challenges of Severe Convective Storms, Atlanta, Georgia, Amer. Meteor. Soc., P2.2.


Dual-Doppler Wind Synthesis

The following procedure, outlined in the manual of the NCAR Cedric program, describes steps taken to synthesize the 3D wind field \((u, v, w)\) from the collection of radial velocity data \((v_r)\) by two radars,

\[
v_{r,m} = u a_m + v b_m + W c_m,
\]

where \(m\) denotes radar number \((m = 1, 2)\), \(W = w + w_t\), where \(w_t\) is the fall velocity of the precipitation targets, and \(a_m = \sin(\alpha) \cos(e)\); \(b_m = \cos(\alpha) \cos(e)\); and \(c_m = \sin(e)\) are geometric coefficients defining the directly observed components of the Cartesian velocity vector based on the radar beam angle (\(\alpha\) is the azimuthal angle relative to north, and \(e\) is elevation angle from the ground) at each point in space. The combined data from two radars comprise the total observed wind field. Observed components are fit in a least-squares sense minimizing the sum of the square of the observation error, \(Q = \sum_m E_m^2 = \sum_m [u a_m + v b_m + W c_m - v_{r,m}]^2\). Taking partial derivatives of \(Q\) with respect to \(u, v,\) and \(W\) produces a system of three equations (summed over both radars) set equal to zero:
\[ u \sum_{m} a_{m}a_{m} + v \sum_{m} b_{m}b_{m} + \sum_{m} c_{m}c_{m} - \sum_{m} d_{m}d_{m} = 0, \]  
(A.2a)

\[ u \sum_{m} a_{m}a_{m} + v \sum_{m} b_{m}b_{m} + \sum_{m} c_{m}c_{m} - \sum_{m} d_{m}d_{m} = 0, \]  
(A.2b)

\[ u \sum_{m} a_{m}a_{m} + v \sum_{m} b_{m}b_{m} + \sum_{m} c_{m}c_{m} - \sum_{m} d_{m}d_{m} = 0. \]  
(A.2c)

Equation A.2c is neglected for a dual-Doppler solution, assuming that only the horizontal wind field can be retrieved with two radars. The geometry of the radars in each deployment was chosen such that the anticipated tornado location would be as close as possible to the location where radar beams cross at a 90° angle near the ground, guaranteeing the most accurate representation of the horizontal component of the wind. Solving for \( u \) and \( v \) using (A.2a, b),

\[ u = \frac{D_{1}B_{2} - D_{2}B_{1}}{A_{1}B_{2} - A_{2}B_{1}} + W \frac{B_{1}C_{2} - B_{2}C_{1}}{A_{1}B_{2} - A_{2}B_{1}} \equiv u^{*} + W \epsilon_{u}, \]  
(A.3a)

\[ v = \frac{D_{2}A_{1} - D_{1}A_{2}}{A_{1}B_{2} - A_{2}B_{1}} + W \frac{A_{2}C_{1} - A_{1}C_{2}}{A_{1}B_{2} - A_{2}B_{1}} \equiv v^{*} + W \epsilon_{v}. \]  
(A.3b)

After \( u \) and \( v \) are calculated, \( w \) is determined by integrating the anelastic mass continuity equation,

\[ \frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y} + \frac{\partial \rho w}{\partial z} = 0, \]  
(A.4)

upward from the ground, where \( w = 0 \) m s\(^{-1}\). The density of air, \( \rho \), is assumed to be,

\[ \rho(z) = \rho_{o} \cdot e^{-\frac{z}{H}}, \]  
(A.5)

where \( \rho_{o} = 1 \) kg m\(^{-3}\) and \( H = 7 \) km. Inserting equations (A.3a, b) into (A.4) and
integrating between consecutive vertical levels ($z_{\text{top}}$ and $z_{\text{bot}}$),

\[
(\rho w)_{z_{\text{top}}} = (\rho w)_{z_{\text{bot}}} - \Delta z \left[ \frac{\partial u^*}{\partial x} + \frac{\partial v^*}{\partial y} \right]_{z_{\text{top}} - z_{\text{bot}}} \]
\[
- \Delta z \left[ \frac{\partial (\rho \epsilon_u w)}{\partial x} + \frac{\partial (\rho \epsilon_v w)}{\partial y} \right]_{z_{\text{top}} - z_{\text{bot}}}. \tag{A.6}
\]

Because $w$ is on the rhs, the solution to this expression is achieved by iterative integration; $\rho w$ is first calculated at $z = z_{\text{top}}$ using the convergence of the observed component of the horizontal winds ($u^*, v^*$) over $\Delta z = z_{\text{top}} - z_{\text{bot}}$, but is then recalculated when its value is placed back into the rhs. Recalculations of $\rho w$ continue until the solution changes less than 0.1 kg m$^{-2}$s$^{-1}$ between successive iterations. The result is used as $\rho w$ at $z_{\text{bot}}$ in the integration over the next vertical layer.

Prior to the calculation of $w$, $W$ should be broken down into $w + w_t$, and $w_t$ determined using radar parameters (note, the $w_t\epsilon$ terms in equation A.3 can be incorporated into the values of $u^*$ and $v^*$, as was done in equation A.6). This traditionally is done using an expression relating precipitation drop size (and thus terminal velocity) and the radar reflectivity factor, determined empirically in past studies. The dual-Doppler syntheses in this study are performed with $w_t = 0$ because: 1) DOW radar reflectivity factor is not calibrated to operational radar reflectivity data from which the empirical relationship between drop size and reflectivity factor is determined, 2) attenuation of the radar power along beam radials would cause a presumably unrealistic horizontal gradient of $w_t$ to be applied, and 3) the radar elevation angles are fairly shallow ($< 15^\circ$). Sensitivity tests have been performed with the 5 June 2001 and 22 May 2004 cases in which the addition of a $w_t$ calculation with a constant value added to DOW reflectivity, making it roughly similar to the amplitude observed by a WSR-88D in each storm, generally yielded changes in the $w$ solution of less than 25% at $z = 2$ km AGL (although, a 50% difference was noted in some small areas) when compared to a solution with no $w_t$ correction. Actual sensitivities may differ in syntheses using properly calibrated DOW reflectivity data.
Vita
James Marquis

James Marquis was born in Sterling, Virginia, on 31 July 1981. James developed a strong interest in science and weather at an early age and this interest followed him throughout his education. He graduated from Park View High School in Sterling, Virginia, in 1999. He then enrolled at the Pennsylvania State University in the fall of 1999, where he earned a Bachelor of Science in Meteorology in 2003. To further his meteorological education, James enrolled in the Penn State graduate program in meteorology in the fall of 2003 to pursue a Master of Science degree under Dr. Yvette Richardson. He earned his M.S. in 2005, and the title of his thesis was *Kinematic Observations of Miscoyclones Along Boundaries During IHOP*. James continued in the Ph.D. program at Penn State in fall 2005, under Dr. Yvette Richardson. At Penn State, James was awarded the Anne C. Wilson Graduate Student Research Award, and the Arnulf I. Muan graduate fellowship.

James has accepted an appointment to a post-doctoral position at Penn State analyzing radar and in situ data collected in tornadic storms during the VORTEX2 field research campaign. He hopes to continue to advance his knowledge of severe storms meteorology and data assimilation.