A CONCEPTUAL MODEL OF NONTORNADIC SUPERCELL THUNDERSTORMS

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by
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Abstract

This study uses dual-Doppler observations of nontornadic supercells obtained by ground-based mobile Doppler radars and idealized numerical simulations in order to develop a conceptual model of a nontornadic supercell, particularly at low levels and on the submesocyclone scale. There are relatively few dual-Doppler studies of supercells in the history of severe storms research owing to the relative rarity of supercell occurrences within dual-Doppler radar networks, and the majority of the dual-Doppler studies feature tornadic supercells. Moreover, the submesocyclone scale and lowest few hundred meters generally have not been well-observed in prior studies, which usually have analyzed either pseudo-dual-Doppler airborne radar observations or data from fixed radar networks. In addition to the fact that a steady state must be assumed for relatively long time periods when analyzing dual-Doppler observations from airborne radars (typically 5-7 minutes), the resolution is coarser than what is afforded by ground-based mobile radars because aircraft must maintain larger distances from the mesocyclone for safety reasons. In regards to ground-based dual-Doppler networks comprising fixed radars, the baselines are usually long (40-75 km baselines are common); thus, the centers of the dual-Doppler lobes, where the geometry is most favorable for accurate wind retrievals, are at a greater range from the radars, resulting in a relatively coarse resolution and inability to observe the lowest few hundred meters owing to radar horizon limitations. In the first part of this dissertation, five nontornadic supercell thunderstorms are analyzed using high-resolution dual-Doppler radar data.
obtained by a pair of mobile ground-based radars. Three out of five observed supercells had well-developed low-level rotation. The observed low-level kinematic fields of the nontornadic supercells with low-level rotation are compared to the low-level kinematic fields of tornadic supercells that have been previously documented. In previous studies, tornadic and nontornadic supercells have had strikingly similar kinematic characteristics on the mesocyclone scale. Thus, discrimination between tornadic and nontornadic supercells has been very difficult (probably fewer than 25% of supercells are tornadic).

It is determined that the observed low-level kinematic structure of nontornadic supercells is qualitatively very similar to the low-level kinematic structure of tornadic supercells, notably two out of three observed nontornadic storms had a “bent-back” rear-flank gust front just like the tornadic supercells, and one of those also had a dual rear-flank gust front, a feature that previously has been observed only in tornadic supercells. The low-level mesocyclone in the nontornadic supercells extends to the lowest analysis level in the three cases having low-level rotation, but the low-level circulation in nontornadic mesocyclones is much weaker than in tornadic mesocyclones. Also, the divergence associated with rear-flank downdrafts is stronger in nontornadic supercells than in tornadic supercells. Vortex line analyses in the observed nontornadic storms show that the vorticity field structure is consistent with baroclinic generation of horizontal vorticity and subsequent tilting into the vertical by an updraft, as has been shown in recent observational and numerical simulation studies.

One major limitation of the observational part of this study is the lack of thermodynamic observations. Thermodynamic retrievals from the dual-Doppler wind syntheses were found to be insufficiently accurate for any rigorous quantitative analysis. In the second part of this study, a series of idealized, dry three-dimensional numerical simulations are used to gain some understanding of the relationship between the low-level thermodynamics and kinematics of supercells, assuming that the idealized simulations can replicate the evolution of the low-level kinematic fields observed in actual supercells. The idealized simulations emulate the generation of near-surface rotation beneath supercell-like (i.e., helical) updrafts in a way consistent with our present understanding of the importance of a downdraft in environments in which vertical vorticity is initially absent at the surface. In the simulations herein, the initial conditions are specified to be horizontally homogeneous but vertical wind shear and veering winds with height are present. A heat source is introduced, resulting in a cyclonically rotating updraft having maximum rotation at midlevels and no rotation at the surface. Once an approximately steady state is achieved, a heat sink is introduced at low levels in proximity to the helical updraft. The resulting downdraft and outflow generate rings of baroclinic horizontal vorticity that encircle the heat sink, and this vorticity is subsequently
tilted by the vertical velocity gradients associated with the downdraft-updraft couplet such that a couplet of vertical vorticity develops at the surface, straddling the line that joins the downdraft and updraft extrema. The intensity of the vortices that develop at the surface depends on the degree to which the baroclinic vorticity can be tilted and stretched, which depends in large part on the extent to which the negatively buoyant low-level air originating in the heat sink can be lifted by the overlying updraft driven by the heat source.

In all cases, the cyclonic member of the vorticity couplets that develop at the surface become dominant, and this bias is presumably attributable to the overlying updraft possessing updraft-scale cyclonic rotation. The most intense cyclonic vortices develop when the outflow and associated baroclinic horizontal vorticity from the heat sink is strong enough to spread beneath the overlying updraft, but is not so cold that the air parcels within the outflow cannot be drawn upward by the overlying updraft. Three modes of what is referred to as "tornadogenesis failure" also are noted. (Even though the simulations do not explicitly resolve tornadoseale motions, we casually use the terminology "tornadogenesis failure" to refer to the inability of intense sub-mesocyclone-scale vortices to develop in the model.) The first failure mode occurs when the cold pool does not develop or just a weak cold pool (potential temperature deficits near the vortex less than 2 K) develops briefly. In the presence of a stronger cold pool, the baroclinically generated vortex lines are tilted upward and interact with the updraft. Tornadogenesis occurs when the vertical vorticity is stretched by the main updraft. The second failure mode occurs when the cold pool intensity is stronger than in the tornadogenesis case (potential temperature deficits near the vortex about 4–8 K). In this mode of tornadogenesis failure, the gust front bends back in a similar manner to the simulations that resulted in tornadogenesis but tornadogenesis never occurs. In the case of the second tornadogenesis failure mode the vortex lines emanating from the near-ground vorticity maximum form arches that rise only up to about 3 km AGL and then descend in the rear flank to the south of the vorticity maximum. The third failure more occurs when the cold pool potential temperature deficit is 8 K or higher. In that case, the gust front rushes ahead of the main updraft. The vertical vorticity is generated along the gust front but the vortices are very shallow. The vortex lines in this case rise only up to 200 m AGL before tilting into the horizontal.
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grateful to all of my teachers and professors, from elementary school to graduate school. You have inspired me to become a professor myself and encouraged my passion for teaching.
So the whirlwind originates in the failure of an incipient hurricane to escape from its cloud: it is due to the resistance which generates the eddy, and it consists in the spiral which descends to the earth and drags with it the cloud which it cannot shake off. It moves things by its wind in the direction in which it is blowing in a straight line, and whirls round by its circular motion and forcibly snatches up whatever it meets.

- *Aristotle, Meteorology*
  *Translated by E. W. Webster*
  *Written 350 B.C.E.*
Supercell thunderstorms are characterized by a rotating updraft that persists at least for the duration of time needed for the air parcels to travel from the base of the updraft to its top. Supercell thunderstorms have been extensively researched owing to their ability to create severe weather, especially tornadoes. Not surprisingly, as most significant tornadoes (F2-F5 on Fujita scale) are associated with supercells, a lot of research activity has focused on the process of tornadogenesis.

Rotating updraft intensity in supercells can exceed 50 m s$^{-1}$, and the vertical vorticity associated with the midlevel rotation (mesocyclone) is on the order of 0.01 s$^{-1}$. The midlevel rotation is generated by the tilting of the environmental horizontal vorticity into the vertical by the updraft. Therefore, supercells form only in environments characterized by strong vertical wind shear. A conceptual model of a supercell (Fig 1.1) depicts two downdrafts owing to the evaporational cooling caused by precipitation falling into the relatively dry midlevel air or into...
a subsaturated boundary layer (Lemon and Doswell 1979). One downdraft is found on the forward flank, and the other on the rear flank of a supercell. The relatively cool downdraft air is separated from the relatively warm and moist inflow air by the forward-flank gust front and the rear-flank gust front. The gust front structure kinematically resembles the fronts of an extratropical cyclone, although a supercell obviously is dynamically very different from an extratropical cyclone (e.g., quasigeostrophic theory cannot be applied to a supercell). Rotunno and Klemp (1985) proposed that the air parcels that comprise low-level mesocyclones acquire horizontal vorticity in the baroclinic zone associated with the forward-flank downdraft, and subsequently the vertical vorticity at low levels is generated by tilting of the horizontal vorticity in the updraft. Some recent observations (e.g., Shabbott and Markowski 2006; Beck et al. 2006; Frame et al. 2009) have found that the baroclinic vorticity generation in the forward-flank of observed supercells was not as strong as anticipated from numerical simulations.

If vertical vorticity is not initially present at the ground, the development of rotation there requires a downdraft (Davies-Jones 1982; Davies-Jones and Brooks 1993). Straka et al. (2007) proposed that the rotation near the ground could be generated by the tilting of purely baroclinic vorticity in the vicinity of a relatively cool downdraft (Fig. 1.2). In their conceptual model, a downdraft forms when relatively dry midlevel air flows through precipitation in the back of the storm, or the precipitation falls into subsaturated boundary layer air. As the downdraft forms, the vorticity rings surrounding the downdraft are baroclinically generated and advected downward. Upon reaching the ground, the vorticity rings are tilted by an updraft, creating arches and a vertical vorticity couplet near the ground whereby there is a vertical vorticity maximum to the left of the rear-flank flow and a vertical vorticity minimum to the right. Existence of such vortex line arches associated with the vertical vorticity couplets in the rear-flank of both tornadic and nontornadic supercells have been observed in airborne Doppler radar observations (Markowski et al. 2008). However, the aforementioned conceptual models of low-level rotation generation do not reveal what would be the difference between tornadic and nontornadic supercells. The analyses of in–situ surface observations within rear-flank downdrafts (Markowski et al. 2002) revealed that rear-flank downdrafts associated with significant tornadic supercells have small deficits of virtual potential tempera-
Figure 1.1. Conceptual model of a supercell thunderstorm as proposed by Lemon and Doswell (1979). The “UD” indicates the midlevel updraft, the “FFD” indicates forward-flank downdraft, “RFD” indicates rear-flank downdrafts. A tornado, if present, occurs at the letter “T”. Thin black lines denote storm-relative streamlines. The thick black line encompasses radar echo. A localized intensification of the RFD associated with the intensification of low-level rotation is sometimes referred to as the “occlusion downdraft” and is shaded blue. Adapted from Lemon and Doswell (1979).
Figure 1.2. One possible way by which a couplet of vertical vorticity can be produced by a purely baroclinic process in an environment containing no ambient vorticity (neither vertical nor horizontal). (a) Baroclinically generated vortex rings encircle a buoyancy minimum that extends throughout a vertical column (RFD region of a supercell, for example); the presence of negative buoyancy causes the rings to sink toward the ground as they are generated. (b) If the vortex rings are swept forward as they descend toward the ground owing to the additional presence of rear-to-front flow through the buoyancy minimum. (c) If the leading edge of the vortex rings can be lifted by an updraft in close proximity to the buoyancy minimum then the vortex rings can be tilted further and stretched upward, leading to arching vortex lines and a couplet of cyclonic (C) and anticyclonic (A) vertical vorticity. From Markowski et al. (2008), adapted from Straka et al. (2007).

It was concluded that the presence of buoyant rear-flank downdraft air parcels was required for tornadogenesis, but was not a sufficient condition. If the baroclinic generation of vorticity in rear-flank downdrafts is important for tornadogenesis, it has to be reconciled with in–situ observations of rear-flank downdrafts that support the notion that strong cold pools are detrimental to the formation of significant tornadoes. However, it could be that some baroclinity is necessary for tornadogenesis to occur while the cold pools that are too strong may inhibit the vertical vorticity intensification by vorticity stretching.

Since the 1970s, when numerical simulations of supercells became computationally feasible, researchers have investigated supercell thunderstorms by using numerical models. In order to investigate the transition of the Del City supercell into its tornadic phase, Klemp and Rotunno (1983) used a higher-resolution nu-
Figure 1.3. Composite diagram illustrating the general characteristics of RFDs associated with supercells that produce “significant” tornadoes vs RFDs associated with nontornadic supercells or those that produce weak, brief tornadoes. The thick, dashed contour is the outline of the hook echo, and thin, solid arrows represent idealized streamlines. From Markowski et al. (2002).

Numerical model initiated within the domain of the Del City simulation by Klemp and Rotunno (1981). Klemp and Rotunno (1983) found that in the higher resolution simulations the low-level vertical vorticity increased dramatically, and that the gust front occluded around the smaller-scale downdrafts that developed near the low-level rotation center. They concluded that the intensification of the rear-flank downdraft during the occlusion process was dynamically forced, owing to perturbation pressure gradients induced by strong rotation in the low levels exceeding the rotation aloft.

The concept of downdraft intensification in the presence of strong low-level rota-
tion was later confirmed by Doppler observations of tornadic supercells (e.g., Brandes 1988) and higher-resolution numerical simulations (e.g., Adlerman et al. 1999). Klemp and Rotunno (1983) defined rear-flank downdrafts as the downdrafts that are responsible for the storm outflow behind the gust front and recognized that rear-flank downdrafts are not uniquely linked to tornadogenesis. They hypothesized that if a storm progressed into a tornadic phase, the gust front would become occluded and a strong downdraft would form at low-levels just behind the gust front. They referred to this strong smaller scale downdraft (that is embedded in the rear-flank downdraft) as an occlusion downdraft (Fig 1.1).

1.2 Motivation for the study of nontornadic supercells

Observations of supercells primarily have been obtained by radars over the years, from noncoherent radars in the 1950s and 1960s (e.g., Stout and Huff 1953; Browning and Ludlam 1962), to Doppler radars beginning in the 1970s (e.g., Burgess et al. 1975; Brandes 1977), to mobile radar platforms used in recent field experiments like the Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX, Rasmussen et al. 1994) and the Radar Observations of Tornadoes and Thunderstorms Experiment (ROTATE, Wurman et al. 1997). Before the first Doppler radars came online in the 1970s, the estimate was that about half of supercells were tornadic, and it was hoped that the introduction of operational Doppler radars would improve the tornado warning lead times. However, Doppler radar observations show that only about 25% of observed mesocyclones are actually tornadic (Trapp et al. 2005).

Dual-Doppler observations enable us to retrieve three-dimensional wind fields from radar data (e.g., Barnes 1964). However, relatively few dual-Doppler analyses exist in the literature because it is difficult to deploy mobile radars in such a way that the observed supercell is within dual-Doppler lobes. There are on the order of only 20 cases in all, of which 10 cases (Table 1.1) are of nontornadic supercells.

One of the principal findings of VORTEX was the striking similarity of the kinematic structure of tornadic and nontornadic supercells on the mesocyclone
### Table 1.1. Table of papers discussing dual-Doppler observations of nontornadic supercells.

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<td>Graham, Texas</td>
<td>Ziegler et al. (2001)</td>
</tr>
<tr>
<td>29 April 1995</td>
<td>Sherman, Texas</td>
<td>Trapp (1999); Markowski et al. (2008)</td>
</tr>
<tr>
<td>23 May 2002</td>
<td>Lipscomb county, Texas</td>
<td>Frame et al. (2009)</td>
</tr>
<tr>
<td>15 May 2003</td>
<td>Shamrock, Texas</td>
<td>French et al. 2008</td>
</tr>
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</table>

scale (e.g., Trapp 1999; Wakimoto and Cai 2000; Markowski et al. 2008). The only dual-Doppler observations of supercells in VORTEX were the observations obtained by airborne radars. Trapp (1999) compared pseudo-dual-Doppler analyses of three tornadic and three nontornadic supercells observed by X-band airborne Doppler radar mounted on NOAA P-3 aircraft. Trapp found that both tornadic and nontornadic supercells had low-level mesocyclones in the lowest several hundred meters above the ground (AGL). They concluded that the existence of a low-level mesocyclone was not sufficient for tornadogenesis.

Wakimoto and Cai (2000) analyzed a nontornadic supercell that was observed by the ELDORA Doppler radar (Hildebrand et al. 1994, 1996; Wakimoto et al. 1996) on 12 May 1995 near Hays, Kansas. This supercell was compared to a tornadic supercell observed by ELDORA Doppler radar on 16 May 1995 near Garden City, Kansas (Wakimoto 1998; Wakimoto and Liu 1998). Wakimoto and Cai (2000) concluded that these two supercells had remarkably similar kinematic structure (see their Fig. 15). Based on the features that could be resolved, the updraft and the downdraft structures were very similar, and the vertical vorticity associated with low-level mesocyclones was approximately equal in strength and found in similar location relative to the respective updrafts. The only differences, they noted, were the more intense precipitation echoes behind the rear-flank gust...
front, the more intense storm-relative inflow and the more intense updrafts and vertical vorticity along the rear-flank gust front in the nontornadic (Hays, Kansas) case.

Airborne Doppler radars are not best suited for observations of low-level structure of supercell thunderstorms. The spatial resolution of airborne Doppler radar data is on the order of 500 m. The radar volumes are typically completed in about 5–7 minutes, and owing to the ground clutter the airborne Doppler radars cannot retrieve observations from the approximately lowest 500 m AGL. The ground-based dual-Doppler networks comprising fixed radars, are characterized by the baselines that are usually long (40-75 km baselines are common). Therefore, the centers of the dual-Doppler lobes are at a greater range from the radars, resulting in a relatively coarse resolution and inability to observe the lowest few hundred meters owing to radar horizon limitations. Since VORTEX, the Doppler on Wheels radars (DOW; Wurman et al. 1997) have routinely collected higher-resolution observations of both tornadic and nontornadic supercells. DOW radar data spatial resolution is on the order of 100 m, and the radar volumes are typically completed in about 60 seconds. DOW radars are Doppler radars mounted on trucks and can be deployed close to supercells. The observations of low-level kinematic fields can be retrieved as low as 50 m AGL. Ground-based (truck-borne) mobile Doppler radars like the DOWs are better suited for studies of low-level kinematic structure of supercell thunderstorms.

In the recent years, several high-resolution dual-Doppler analyses have described the low-level kinematic structure of tornadic supercells (e.g., Wurman et al. 2007a,b; Marquis et al. 2008). The low-level kinematic structure of the observed storms is characterized by the “bent-back” rear-flank gust front (Fig. 1.4) and tornadoes are observed in the outflow, at or just behind the tip of the “bent-back” gust front.

The high-resolution dual-Doppler observations also reveal the presence of a dual gust front structure at low-levels (Fig. 1.4). It was proposed that the double rear-flank gust front structure was caused by multiple surges of storm outflow in the rear flank (Marquis et al. 2008). It is not known whether this double rear-flank gust front existed in aforementioned storms prior to tornadogenesis. Such double rear-flank gust front structures have not been observed in nontornadic supercells.
In summary, analyses derived from airborne dual-Doppler data and fixed ground-based dual-Doppler networks have not revealed any recurring kinematic differences (e.g., peak vertical vorticity, magnitude of stretching and tilting, peak updraft and downdraft velocities) between tornadic and nontornadic supercells, perhaps because only the mesocyclone scale was well-resolved and perhaps because of the inability to resolved the lowest few hundred meters. High-resolution numerical studies of supercells (e.g., Finley et al. 2001; Gaudet and Cotton 2006) either produce tornadoes too often, or researchers do not document the nontornadic re-
The purpose of this study is to explore the low-level kinematic structure of non-tornadic supercells on sub-mesocyclone scale using high-resolution dual-Doppler radar observations in order to identify recurring features within the observed non-tornadic supercells. The focus of the analyses is the region where the tornadoes are observed in the high-resolution dual-Doppler observations: the secluded outflow air behind the “bent-back” gust front. The available observations of the five non-tornadic supercells are limited to the radar data obtained by a pair of Doppler-on-Wheels radars (DOW; Wurman et al. 1997) and do not include thermodynamic observations. Idealized three-dimensional numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts are performed to explore whether the kinematic aspects of the observed non-tornadic supercells could be replicated, and how the attendant thermodynamic conditions influence the dynamics of near-ground vorticity generation.

Chapter 2 describes the observations of low-level mesocyclones in non-tornadic supercells. The idealized three-dimensional numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts are presented in Chapter 3. Chapter 4 discusses a conceptual model of non-tornadic supercells. A summary and conclusions are given in Chapter 5. Finally, in the Appendix, I present the methodology and advantages of the multi-pass objective analyses used in the radar analyses herein.
2.1 Datasets

The DOW radars are pulsed, pencil-beam, Doppler radars mounted on trucks (Wurman et al. 1997). They have been used to study a wide range of atmospheric phenomena, including supercell thunderstorms, tornadoes, boundary layer processes, and hurricanes. The wavelength and the stationary, half-power beamwidth of DOW radars are 3 cm and 0.93°, respectively. The mobility of DOWs allows the radar to be deployed close to observed meteorological phenomena enabling us to observe the region of the meteorological phenomenon close to the ground. The narrow beam permits higher data resolution measurements (e.g., at 6 km range, 100 m \times 100 m \times 75 m) than are usually obtained by stationary or airborne radars. Also, the wavelength of 3 cm makes them suitable for collecting observations in rainy environments (such as supercell thunderstorms) because the 3-cm wavelength radiation has been shown to give the “optimal balance between resolution and rain..."
penetration” (Wurman et al. 1997). Such radar data from the ground-based radars can be used to explore whether there are any recurring differences in kinematic structure between tornadic and nontornadic supercells on the scale smaller than the mesocyclone scale and at the heights below 500 m AGL. The data available for this study are five dual-Doppler radar datasets of nontornadic supercells obtained by a pair of DOW radars. In all five datasets, the radial velocity differences associated with low-level mesocyclones had amplitudes of less than 30 m s$^{-1}$ at the lowest radar elevation angles (at about 100-200 m AGL). Also, the DOW radar operators did not observe tornadoes during these deployments nor there are any tornado reports in the Storm Prediction Center’s archives associated with these supercells. Thus, I am confident that all five supercells were indeed nontornadic.

Locations of DOW radars, and dual-Doppler lobes for the cases where WSR-88D radar data are available are given in Fig 2.1.

The first dataset is from a nontornadic supercell observed on 4 May 2001 near Brownfield, Texas. The pair of DOW radars intercepted a supercell that formed in a low-CAPE environment ($\approx 800$ J kg$^{-1}$) along a dryline. The DOW radars deployed in a west–east line, with a baseline between the radars of 17.0 km. The dual-Doppler data were available from 0103 UTC until 0159 UTC, around the time of peak storm intensity.

The second dataset documented a nontornadic supercell that occurred on 12 June 2004 near Sprague, Nebraska (Fig 2.1a). The supercell developed along a dryline in a high-CAPE ($\approx 4100$ J kg$^{-1}$) environment in the warm sector of a developing mid-latitude cyclone about 100 miles south of the warm front. Nearby stations reported 60 miles per hour wind gusts in the outflow. The DOW radars intercepted the storm and deployed approximately along a north–south line with a baseline of 8.4 km. The dual-Doppler data were available from 2203 UTC until 2215 UTC and from 2219 UTC until 2222 UTC, around the time of peak storm intensity.

The third dataset comes from a nontornadic supercell observed on 15 June 2004 near Watertown, South Dakota (Fig 2.1b). This supercell formed in moderate-CAPE environment ($\approx 2000$ J kg$^{-1}$) in the warm sector ahead of a cold front associated with a synoptic scale cyclone with a center of low pressure in southern Saskatchewan. A nearby station reported 60–70 miles per hour wind gusts in the
outflow. The DOW radars deployed along the north–south line with the baseline of 9.0 km. The dual-Doppler data were available from 2210 UTC until 2245 UTC, after the time of peak intensity.

The fourth dataset consists of dual-Doppler radar observations of a nontornadic supercell intercepted by a pair of DOW radars on 17 June 2004 near Las Animas, Colorado. On that day several supercells developed in moderate-CAPE

Figure 2.1. DOW radar locations and dual-Doppler lobes for cases where WSR-88D radar data are available. a) Sprague supercell, at 2220 UTC, 12 June 2004. b) Watertown supercell, at 2233 UTC, 15 June 2004. c) Amarillo supercell, at 0111 UTC, 19 June 2004.
environment ($\approx 1500 \text{ J kg}^{-1}$) along a dryline in southeastern Colorado. No high wind reports are available for this storm. However, nearby there were several 1–1.75 inch hail reports. The DOW radars deployed along a north–south line with a baseline distance of 10.6 km. From 2351 UTC (17 June 2004) until 0023 UTC (18 June 2004) the DOW radars were observing a supercell to their west. This supercell did not develop low-level rotation. From 0024 UTC (18 June 2004) until 0035 UTC the DOW radars were observing a supercell to their east. That supercell had strong low-level rotation and is analyzed in this study. The storm dissipated only 5 minutes after achieving its peak low-level rotation at 0030 UTC.

The fifth dataset consists of the dual-Doppler radar observations of a non-tornadic supercell intercepted by a pair of DOW radars on 19 June 2004 near Amarillo, Texas (Fig 2.1c). Around 0100 UTC, several supercells have developed in the high-CAPE environment ($\approx 3300 \text{ J kg}^{-1}$) in the Texas panhandle. These storms initiated along an outflow boundary created by convection that developed earlier in western Oklahoma. No high wind reports are available for this storm but a nearby station reported 1 inch hail. A pair of DOW radars intercepted the supercell and deployed along a southwest–northeast line with a 13.3 km baseline. The dual-Doppler data were available from 0102 UTC until 0144 UTC, during and after the time of the supercell peak intensity.

### 2.2 Analysis Techniques

In the five aforementioned deployments, the DOW radars were scanning at elevation angles ranging from $0.3^\circ$ to $21.9^\circ$. The elevation angle differences between subsequent elevation angles range from about $0.5^\circ$ for radar sweeps in the lowest $3^\circ$, to $1.5^\circ$ at higher elevation angles. Each radar volume typically takes about 60 s to complete, and the typical top of the radar data that can be synthesized into three-dimensional wind synthesis is 1.5–2.5 km AGL at the intercept distances. The DOW data were rotated from a truck-relative reference frame to the earth-relative reference frame by aligning ground clutter targets with known locations of cell towers and the local road network. Ground clutter and other erroneous data were removed from the data sets.

Radar data are interpolated to a Cartesian grid using an objective analysis
technique to facilitate operations such as three-dimensional isosurface viewing, dual-Doppler wind synthesis, or simple two-dimensional contouring. Through judiciously chosen tuning parameters (which typically are based on the data spacing, $\Delta$), advanced objective analysis methods allow one to filter scales that are poorly resolved in radar observations (Trapp and Doswell 2000). Furthermore, it has been shown that multiple passes (i.e., successive corrections) of an objective analysis steepen the response of the filter, i.e., such techniques are less damping at well-resolved scales (e.g., $8-20\Delta$) while still removing scales that are poorly resolved (e.g., $<4\Delta$) (Koch et al. 1983). Thus, a multi-pass objective analysis can provide a better fit to the observations than a single-pass objective analysis, yet still suppress small-scale noise (Majcen et al. 2008). A detailed description of the benefits of using a 2-pass versus 1-pass Barnes objective analysis is outlined in the Appendix.

Data from two DOW radars are interpolated to a grid using a two-pass Barnes analysis (Barnes 1964; Koch et al. 1983; Majcen et al. 2008). Because DOW radar volumes take about 60 seconds to complete, the data point locations are determined by applying an advection correction to remove artificial tilting with height of observed features due to their motion between consecutive radar sweeps (Matejka 2002). The extrapolation of data to the grid points is not permitted, thus the radar data is only interpolated to the grid points that are within the data domain. In order for the analyzed kinematic fields in the observed storms to be comparable for quantitative analysis, all datasets are objectively analyzed using a two-pass Barnes filter employing the same smoothing and convergence parameters. The coarsest data separation in the region of interest in the aforementioned datasets is estimated to be 300 m. For this data separation distance, the recommended smoothing parameter (Pauley and Wu 1990), $\kappa_0$, is 0.16 km$^2$, and the recommended convergence parameter (Majcen et al. 2008), $\gamma$, is 0.3. For this choice of objective analysis parameters, the theoretical Barnes filter response function is given in Fig. 2.2. The theoretical response function shows that the wavelengths of about 750 m and longer have a theoretical response greater than 0.50. Therefore, I am confident that the features larger than 750 m are well resolved in our analyses.

The theoretical Barnes filter response function can only be derived for continuous and unbound data. The actual amplitude response will differ somewhat from
the theoretical response and there also may be phase shifts because the data are discrete and bounded (Pauley and Wu 1990; Askelson and Straka 2005; Askelson et al. 2005). Equivalent objective analyses of unbound and continuous data can be produced by using an infinite number of smoothing parameter combinations (Spencer et al. 2007). For discrete data within bounded domains (such as radar data) it is possible that the choice of parameter combinations in the two-pass may affect the response. However, this is not likely to be a significant problem for fairly regularly spaced discrete data (Spencer et al. 2007) such as DOW radar data.

The three-dimensional field is obtained from the radial velocities, $v$, observed by a pair of DOW radars ($v$)

$$v_i = u a_i + v b_i + W c_i,$$  \hspace{1cm} (2.1)
where $i$ denotes radar number ($i = 1, 2$), $W = w + w_t$, where $w_t$ is the fall speed of the precipitation targets, and $a_i = \sin \alpha \cos \beta$; $b_i = \cos \alpha \cos \beta$; and $c_i = \sin \beta$ are the coefficients describing the component of the radial velocity to the Cartesian velocity vector based on the azimuthal angle relative to north, $\alpha$, and the elevation angle from the ground, $\beta$.

The precipitation target fall speed, $w_t$, is traditionally calculated by using an empirical expression that relates the precipitation fall speed and the radar reflectivity factor. In this study it is assumed that $w_t = 0$, for the following reasons: the three-dimensional wind syntheses analyzed in this study typically include only the data observed at elevation angles of less than 10° at which the precipitation fall speeds contribution to the vertical velocity are small; the DOW radar reflectivity factor is not calibrated to the operational radar reflectivity data from which the empirical relationship between precipitation drop speed and reflectivity factor is determined; the attenuation of radar power along the radar beam is not negligible in supercells owing to heavy precipitation and could cause an incorrect horizontal gradient of $w_t$ to be applied.

The observational error equation, $Q = \sum_i E_i^2 = \sum_i [(u a_i + v b_i + W c_i - v_i)^2]$, is minimized in the least-square sense. By taking partial derivatives of $Q$ with respect to $u$, $v$, and $W$ we get a system of three equations:

\begin{align*}
    u \sum_i a_i a_i + v \sum_i a_i b_i + W \sum_i a_i c_i - \sum_i a_i v_i &= 0, \\
    u \sum_i a_i b_i + v \sum_i b_i b_i + W \sum_i b_i c_i - \sum_i b_i v_i &= 0, \quad (2.2) \\
    u \sum_i a_i c_i + v \sum_i b_i c_i + W \sum_i c_i c_i - \sum_i c_i v_i &= 0,
\end{align*}

or, in a simpler notation,

\begin{align*}
    u A_1 + v B_1 + W C_1 - D_1 &= 0, \\
    u A_2 + v B_2 + W C_2 - D_2 &= 0, \quad (2.3) \\
    u A_3 + v B_3 + W C_3 - D_3 &= 0,
\end{align*}
This system of three equations is reduced to a two equation system by neglecting the third equation because it is assumed that only the horizontal wind vector components are resolvable by two radars. Solving for \( u \) and \( v \) yields

\[
\begin{align*}
u &= \frac{D_1 B_2 - D_2 B_1}{A_1 B_2 - A_2 B_1} + W \frac{B_1 C_2 - B_2 C_1}{A_1 B_2 - A_2 B_1} = u^* + W \epsilon_u, \\
v &= \frac{D_2 A_1 - D_1 A_2}{A_1 B_2 - A_2 B_1} + W \frac{A_2 C_1 - A_1 C_2}{A_1 B_2 - A_2 B_1} = v^* + W \epsilon_v.
\end{align*}
\]

(2.4)

The vertical velocity was assumed to be zero at all analysis levels below the lowest data level. After \( u \) and \( v \) are calculated, \( w \) is determined by integrating the anelastic mass continuity equation,

\[
\frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y} + \frac{\partial \rho w}{\partial z} = 0,
\]

upward from the ground where \( w = 0 \), and assuming that

\[
\rho = \rho_0 \exp \left( -\frac{z}{L} \right)
\]

(2.6)

where \( \rho_0 = 1.0 \text{ kg m}^{-3} \), and \( L = 7.0 \text{ km} \).

The newly calculated \( w \) is then used to calculate the new values of \( u \) and \( v \) using (2.4) and the system of equations is solved iteratively. Because extrapolation was forbidden in the objective analysis stage, wind data are not retrieved at the lowest grid level \((z = 0 \text{ m})\), and even at \( z = \Delta z \) \((z = 100 \text{ m})\) in some parts of the dual-Doppler domain. This in turn may result in unrealistic lower boundary condition in some parts of the domain. However, when the extrapolation of wind data is allowed (resulting in wind data being available at \( z = 0 \text{ m} \)), the resultant second-order kinematic fields (such as divergence and vertical vorticity) showed unrealistic features at the ground level, especially near the edges of data domain. Therefore, the extrapolation of the wind data was not allowed. Storm-relative wind vectors are calculated by subtracting the storm motion from the retrieved horizontal wind fields. All wind vectors presented herein are storm-relative wind vectors. The thermodynamic properties could not be accurately retrieved from kinematic fields. A comparison of the kinematic structures analyzed with the analysis methods typical for the DOW radar data and the airborne radar data
Figure 2.3. Low-level horizontal winds, vertical velocity (shaded), and vertical vorticity (contoured) in the Brownfield supercell at 0159 UTC as analyzed with different analysis methods. Vertical vorticity is contoured every 0.01 s$^{-1}$; units of vertical velocity are m s$^{-1}$. a) data analysis method typical for airborne radars (250 m AGL) b) DOW Barnes 1-pass objective analysis (100 m AGL) c) DOW Barnes 2-pass objective analysis (100 m AGL)

(Fig. 2.3) shows the improvement in resolving submesocyclone scale features in the DOW data analyzed with a 2-pass Barnes objective analysis.

The grid spacing in all analyzed cases was 100 m, in both horizontal and vertical. The domain sizes are 17.0 × 17.0 × 3 km$^3$ in the Brownfield and the Sprague case, 18.0 × 18.0 × 3 km$^3$ in the Watertown case, 21.2 × 21.2 × 3 km$^3$ in the Las
Animas case, and $28.6 \times 28.6 \times 3 \text{ km}^3$ in the Amarillo case.

### 2.3 Findings

The following three out of five observed supercells had rotation in the low levels: the Brownfield supercell, the Sprague supercell, and the Las Animas supercell. The analyses of two cases in which low-level rotation was not observed [the Watertown (Fig. 2.4) and Amarillo supercells (Fig. 2.5)] reveal no coherent downdrafts of amplitudes greater than 4 m s$^{-1}$ (not shown). Although this mode of tornado-generation failure may be apparent in visual or DOW radar observations, in most cases supercells are observed by fixed radars (such as the National Weather Service WSR-88D radars) at large distances (100 km or greater). At such distances a WSR-88D radar would only be able to observe the midlevels of supercells and could not distinguish between supercells with low-level rotation and those without low-level rotation. This presents a considerable challenge for tornadogenesis forecasting owing to the fact that only about 25% of observed mesocyclones are tornadic (Trapp et al. 2005).

The focus of this study is the low-level kinematic fields in nontornadic supercells that have a well-developed low-level circulation. The Brownfield supercell depicted at the time when the maximum low-level rotation was observed (Fig. 2.6) had a well-developed rear-flank downdraft and a bent-back gust front structure similar to the low-level kinematic structure observed in tornadic supercells (Fig. 1.4). The maximum vertical vorticity ($x = 8.3 \text{ km, } y = 8.3 \text{ km}$; Fig. 2.6) associated with the low-level mesocyclone was observed just behind the bent-back gust front.

The Sprague supercell also had a well developed bent-back gust front (Fig. 2.7). The strongest rotation at low levels was also found just behind the tip of the bent-back gust front ($x = 9.3 \text{ km, } y = 7.1 \text{ km}$; Fig. 2.7). This storm had a pronounced hook echo just to the southeast of the region of strongest vertical vorticity. The region of strong convergence in the back of the storm ($x = 6.0 \text{ km, } y = 8.0 \text{ km}$; Fig. 2.7) seems to be caused by the outflow from the supercell located to the west of the Sprague supercell (Fig. 2.1a).

The Las Animas supercell had similar kinematic structure, however not all of the rear-flank gust front was observed (Fig. 2.8). The maximum vertical vorticity
Figure 2.4. The Watertown supercell at 2216 UTC: a) Low-level horizontal wind vectors, and uncalibrated reflectivity (shaded) at 500 m AGL, b) radial wind velocity. The radial velocity recorded by DOW2 radar shows a radial velocity couplet characteristic for midlevel mesocyclone rotation in supercells at about 3.5 km AGL.

\((x = 7.5 \text{ km}, y = 9.9 \text{ km}; \text{Fig. 2.8})\) in this case was found behind the gust front, but unlike in the Brownfield and the Sprague supercell the low-level mesocyclone was separated from the gust front by a strong downdraft.
Figure 2.5. The Amarillo supercell at 0121 UTC: a) Low-level horizontal wind vectors, and uncalibrated reflectivity (shaded) at 500 m AGL, b) radial wind velocity. The radial velocity recorded by DOW2 radar shows a radial velocity couplet characteristic for midlevel mesocyclone rotation in supercells at about 4.8 km AGL.
Figure 2.6. Horizontal wind vectors, vertical vorticity, and convergence in the Brownfield supercell at 0159 UTC at 200 m AGL. Solid black line denotes 15 dBz reflectivity contour (uncalibrated). Rear-flank gust front location is denoted by a solid green line. Units of convergence and vertical vorticity are 0.01 s$^{-1}$. a) Convergence (shaded) b) Vertical vorticity (shaded).
Figure 2.7. Horizontal wind vectors, vertical vorticity, and convergence in the Sprague supercell at 2214 UTC at 100 m AGL. Solid black line denotes 15 dBz reflectivity contour (uncalibrated). Rear-flank gust front location is denoted by a solid green line. Units of convergence and vertical vorticity are 0.01 s$^{-1}$. a) Convergence (shaded) b) Vertical vorticity (shaded).
Figure 2.8. Horizontal wind vectors, vertical vorticity, and convergence in the Las Animas supercell at 0030 UTC at 100 m AGL. Solid black line denotes 40 dBz reflectivity contour (uncalibrated). Rear-flank gust front location is denoted by a solid green line. Units of convergence and vertical vorticity are $0.01 \text{s}^{-1}$. a) Convergence (shaded) b) Vertical vorticity (shaded).
2.3.1 Observations of the Brownfield supercell low-level mesocyclone evolution

To elucidate the intensification of the low-level rotation in the Brownfield supercell we investigate the kinematic fields in the vicinity of the low-level mesocyclone a few minutes before and during the time of the maximum low-level rotation (0159 UTC) at the lowest analysis level (200 m AGL).

At 0154 UTC (Fig. 2.9) a low-level mesocyclone is observed behind the “bent-back” rear-flank gust front. The low-level mesocyclone is located in the region of strong divergence. The center of rotation is colocated with the divergence maximum associated with the rear-flank downdraft \((x = 8.6 \text{ km}, y = 7.0 \text{ km})\). The direction of horizontal vorticity vectors along the rear-flank gust front suggest strong baroclinic vorticity generation along the gust front. In the low-level mesocyclone near-inflow region \((x = 8.0 \text{ km}, y = 8.0 \text{ km})\) the horizontal vorticity vectors have a significant crosswise component of horizontal vorticity in contrast with previous modeling studies (e.g., Wicker and Wilhelmson 1995) that suggest that the mesocyclone near-inflow region is characterized with almost purely streamwise vorticity. This may indicate that the crosswise horizontal vorticity in the near-inflow could be a result of not only baroclinic the generation of vorticity in the forward flank baroclinic region (as proposed by Wicker and Wilhelmson) but also a result of the baroclinic vorticity generation in the rear-flank downdraft region (as proposed by Straka et al. 2007) or that the crosswise vorticity may be generated from the streamwise vorticity through the crosswise exchange process (Adlerman et al. 1999).

At 0156 UTC (Fig. 2.10) the center of rotation associated with the low-level mesocyclone \((x = 8.4 \text{ km}, y = 7.8 \text{ km})\) is no longer colocated with the low-level divergence maximum. The convergence at the bent-back part of the gust front intensifies.

The intensification of the low-level convergence associated with the bent-back part of the gust front continued at 0158 UTC (Fig. 2.11). The mesocyclone axis of rotation \((x = 8.2 \text{ km}, y = 8.2 \text{ km})\) is now almost exactly colocated with the convergence maximum at the tip of the bent-back part of the gust front.

The Brownfield supercell reached its peak low-level vertical vorticity at 0159
UTC (Fig. 2.12). The center of rotation is found at the tip of the back-bent rear-flank gust front. The convergence associated with bent-back part of the gust front weakened. At this time the radar observations end, so we are not able to observe whether the rotation intensified further.

A west to east vertical cross section through the vorticity maximum in Brownfield supercell (Fig. 2.13a) reveals a relatively shallow but wide (in west to east direction) downdraft above the low-level center of rotation \((x = 8.3 \text{ km})\). A south to north cross section through the low-level vorticity maximum in Brownfield supercell reveals the northward tilt of the vorticity maxima with height (Fig. 2.13b). Three vorticity maxima can be identified at \(200 \text{ m AGL, at } y = 5.6 \text{ km; } y = 8.3 \text{ km; and } y = 9.8 \text{ km.}\) All of them are tilted northward extending from \(200 \text{ m AGL}\) to the top of the three dimensional wind analysis (\(2.4 \text{ km AGL}\)). The vorticity maximum associated with the low-level mesocyclone axis of rotation \((x = 8.0 \text{ km, Fig. 2.13a; and } y = 8.3, \text{ Fig. 2.13b})\) is shallow, extending only up to about \(500 \text{ m AGL}\). This indicates that the vortex does not interact with the midlevel mesocyclone, i.e. the near-ground mesocyclone is detached from the midlevel mesocyclone.

The observed tilt with height of vorticity maxima in the vertical cross section is not necessarily representative of the low-level mesocyclone axis of rotation tilt with height. A comparison between the low-level mesocyclone axis of rotation locations at \(200 \text{ m AGL (Fig. 2.14a) and 1 km AGL (Fig. 2.14b) reveals that the axis of rotation tilts with height to the north-west at an angle of about 53° from the vertical. Also, at 1 km AGL, the strongest vorticity maxima are not associated with the mesocyclone axis of rotation \((x = 7.5 \text{ km, } y = 9.4; \text{ Fig. 2.14b})\) but are found along the rear-flank gust front. Thus, the examination of individual components of vorticity vector field in horizontal and vertical cross sections may not reveal the full three-dimensional structure of vorticity field in supercells.

Vortex line analysis can provide a better insight into the three-dimensional structure of vorticity field in supercells. Vortex lines emanating from the low-level vorticity maxima associated with the low-level mesocyclone were found to go straight up from the respective vorticity maxima until they reached the top of data domain (not shown). In the Brownfield case the top of dual-Doppler data coverage was at \(2.4 \text{ km AGL}\). In order to avoid the data limitations, the vortex lines were instead calculated in the vicinity of the vorticity maximum (Fig. 2.15a)
and compared to the vortex lines emanating from the vorticity maximum located along the gust front (Fig. 2.15b). Both sets of vortex lines form arches that descend in the same general region of negative vertical vorticity.
Figure 2.9. Vertical vorticity (contoured), convergence (shaded) in the Brownfield supercell at 0154 UTC at 200 m AGL, and a) Horizontal wind vectors at 200 m AGL, b) Horizontal vorticity vectors at 300 m AGL. Vertical vorticity is contoured every 0.01 s$^{-1}$. Zero contours are omitted for clarity. Solid green line denotes the rear-flank gust front.
Figure 2.10. Same as Fig. 2.9 but at 0156 UTC.
Figure 2.11. Same as Fig. 2.9 but at 0158 UTC.
Figure 2.12. Same as Fig. 2.9 but at 0159 UTC.
Figure 2.13. West to east, and south to north vertical cross sections through the low-level vorticity maximum ($x = 8.3$ km, $y = 8.3$ km) in Brownfield supercell. Vertical vorticity contoured every $0.01$ s$^{-1}$, zero contour line omitted for clarity. Vertical velocity shaded. Units of vertical velocity are m s$^{-1}$. a) west to east at $y = 8.3$ km. b) south to north at $x = 8.3$ km.
Figure 2.14. Vertical vorticity (contoured), convergence (shaded), and horizontal wind vectors in the Brownfield supercell at 0159 UTC at a) 200 m AGL, b) 1 km AGL. Vertical vorticity is contoured every 0.01 s$^{-1}$. Zero contours are omitted for clarity. Purple dot denotes the location of the axis of rotation at 200 m AGL.
Figure 2.15. Vortex lines emanating from a location about 0.5 km to the south east of the rear-flank vorticity maximum in the Brownfield supercell. Vertical vorticity contoured every 0.01 s$^{-1}$, zero contour line omitted for clarity. Vertical velocity shaded. Units of vertical velocity are m s$^{-1}$. Grey shaded areas denote reflectivity 100 m AGL for orientation purposes.
2.3.2 Observations of the Sprague supercell low-level mesocyclone evolution

This section describes the intensification and the decay of the low-level rotation in the Sprague supercell. We investigate the kinematic fields in the vicinity of the low-level mesocyclone a few minutes before, during and after the time of the maximum low-level rotation (2214 UTC) at the lowest analysis level (100 m AGL).

At 2213 UTC (Fig. 2.16) low-level rotation is centered at the interface between the region of strong divergence \((x = 9.6 \text{ km}, y = 6.2 \text{ km})\) associated with rear-flank downdraft and the rear-flank gust front that has started to wrap around the rear-flank downdraft, at the location that Lemon and Doswell (1979) denoted as the region where tornadoes typically form. The vertical vorticity maximum is located at its northern edge, in the region of strong convergence. This nontornadic supercell has a dual gust front structure similar to the dual gust front structure observed in tornadic storms (e.g., Wurman et al. 2007a).

At 2214 UTC, the low-level mesocyclone reached its maximum intensity of about 0.10 s\(^{-1}\) (Fig. 2.17). The mesocyclone axis of rotation is found just behind the tip of the back-bent gust front \((x = 9.2 \text{ km}, y = 7.0 \text{ km})\). The mesocyclone diameter is only about 2 km. The trajectories of air parcels that comprised the low-level vorticity maximum (at 200 m AGL) at 2214 UTC were calculated back and forward in time (from 2213 UTC to 2221 UTC) and show that the air parcels that comprised the vorticity maximum originated in the low levels of the rear-flank outflow to the northwest of the low-level mesocyclone (Fig. 2.18). After passing through the vorticity maximum the air parcels ascended only up to about 500 m AGL (Fig. 2.19). This indicates that the air parcels were not buoyant enough to be lifted, and consequently the vertical vorticity was not stretched into a tornado. The origins of the air parcels that comprised the vorticity maximum associated with the mesocyclone did not originate in the forward-flank baroclinic zone as proposed by Rotunno and Klemp (1985). The horizontal vorticity vectors (Fig. 2.17b) indicate two regions of baroclinic vorticity generation, one just east of the rear-flank downdraft \((x = 11.2 \text{ km}, y = 6.0 \text{ km})\) and the other one along the rear-flank gust front (e.g., \(x = 12.5 \text{ km}, y = 7.5 \text{ km})\).

Only a minute later, at 2215 UTC (Fig. 2.20), the vertical vorticity decreased,
and at the location where the vorticity maximum was found at 2214 UTC, a maximum in divergence is found indicating a strong downdraft \((x = 9.6 \text{ km}, y = 6.8 \text{ km}; \text{Fig. 2.20})\). It is plausible that the tornadogenesis failure in this case occurred owing to the occlusion downdraft developing in the axis of mesocyclone rotation. However, at the temporal resolution of 1 minute it not clear whether this local minimum of vertical velocity was a result of an occlusion downdraft descending from above or a result of horizontal advection of vertical velocity and the wrapping of the rear-flank downdraft around the low-level mesocyclone center. If the tornadogenesis failure indeed occurred owing to the occlusion downdraft appearance, it would be in contrast with the proposed trigger for tornadogenesis in the Garden City supercell (Wakimoto et al. 1998a and 1998b). It was proposed that the appearance of the occlusion downdraft coincides with the major transition of the vortex from a one-cell vortex to a two-cell vortex just before the tornadogenesis occurs (Wakimoto et al. 1998b; their Fig. 6).

After 2215 UTC, the vorticity maximum that got separated from the main updraft by the downdraft (Fig. 2.20a, \(x = 9.6 \text{ km}, y = 6.6 \text{ km}\)) weakens and is advected by the strong rear-flank outflow in the southeast direction toward the southern part of the rear-flank gust front.

The vorticity maxima in Sprague supercell show a tilt with height, in this case, the tilt is to the southwest (compare Figs. 2.21a and 2.21b). As in the Brownfield case the vertical vorticity is strongest in the lowest analysis level (100 m AGL) and weakens with height.

The vortex lines emanating from the vorticity maximum at 2214 UTC point straight up until they reach the top of the data domain (at about 1.5 km AGL). At all analysis times the vortex lines emanating from the vorticity maxima located along the gust front form arches that descend in the same general region of negative vertical vorticity (e.g., at 2219 UTC Fig. 2.22). None of the vortex lines that pass through the vorticity maxima that are located below 500 m AGL were found to originate in environmental air. The vortex lines that pass through the vorticity maxima that are located at or above 1 km AGL along the gust front originate mostly in the environmental or forward-flank air (e.g., Fig. 2.23), suggesting that the rotation in the lowest 500 m AGL is a result of the tilting of baroclinically generated horizontal vorticity in the cold pool, while the tilting of the environ-
mental horizontal vorticity contributes to the overall mesocyclone rotation only at the heights greater than about 500 m AGL.
Figure 2.16. Vertical vorticity (contoured), convergence (shaded) in the Sprague supercell at 2213 UTC at 200 m AGL, and a) Horizontal wind vectors at 200 m AGL, b) Horizontal vorticity vectors at 300 m AGL. Vertical vorticity is contoured every 0.01 s$^{-1}$. Zero contours are omitted for clarity. Solid (dashed) green line denotes the primary (secondary) rear-flank gust front.
Figure 2.17. Same as in Fig. 2.16 but for 2214 UTC.
Figure 2.18. Trajectories of the air parcels that comprised the low-level vorticity maximum in the Sprague supercell at 2214 UTC. Reflectivity at 100 m AGL is shaded, and horizontal wind vectors plotted for orientation. The view is from south-south-east and above.
Figure 2.19. Trajectories of the air parcels that comprised the low-level vorticity maximum in the Sprague supercell at 2214 UTC projected onto $x$-$z$ plane. The horizontal red line denotes the height of 500 m AGL.
Figure 2.20. Same as in Fig. 2.16 but for 2215 UTC.
Figure 2.21. West to east, and south to north vertical cross sections through the low-level vorticity maximum \((x = 9.3 \text{ km}, y = 7.1 \text{ km})\) in Sprague supercell. Vertical vorticity contoured every \(0.01 \text{s}^{-1}\), zero contour line omitted for clarity. Vertical velocity shaded. Units of vertical velocity are \(\text{m s}^{-1}\). a) west to east at \(y = 7.1 \text{ km}\). b) south to north at \(x = 9.3 \text{ km}\).
Figure 2.22. Vortex lines emanating from a vorticity maximum located on the gust front in the Sprague supercell at 2219 UTC. Vertical vorticity contoured every 0.01 s$^{-1}$, zero contour line omitted for clarity. Vertical velocity shaded. Units of vertical velocity are m s$^{-1}$. Grey shaded areas denote reflectivity 100 m AGL for orientation purposes.
Figure 2.23. Vortex lines passing through a vorticity maximum located at 1000 m AGL in the Sprague supercell at 2214 UTC. View from the southeast, and the top view. Grey shaded areas denote reflectivity 100 m AGL for orientation purposes.
2.3.3 Observations of the Las Animas supercell low-level mesocyclone evolution

The evolution of the low-level rotation in the Las Animas supercell is different from the low-level rotation evolution in the Brownfield and the Sprague supercells. The bent-back rear-flank gust front structure is not observed in this case (Fig. 2.8), and it seems that the rear-flank gust front has rushed ahead of the main updraft although this cannot be confirmed with the available data. The downdraft is found to be elongated along the rear-flank gust front, separating the low-level mesocyclone from the rear-flank gust front.

At 0028 UTC (Fig. 2.24) a weak and narrow mesocyclone is observed behind the occlusion downdraft. The rear-flank downdraft downdraft is elongated along the gust front and the relatively strong convergence zone associated with the low-level mesocyclone is located to the south of the occlusion downdraft.

At 0030 UTC (Fig. 2.25) the rotation intensifies as the low-level convergence in the center of the mesocyclone increases and the the divergence associated with the nearby rear-flank downdraft intensity increases. After 0030 UTC the low level rotation decreases and the storm rapidly dissipates in the next 5 minutes. In this case it seems that the tornadogenesis failure occurred owing to the storm dissipation.

Backward trajectories of the air parcels found in the low-level vorticity maximum at 0030 UTC (Fig. 2.26) show that the air descended from above in the occlusion downdraft and and then started to ascend in the narrow convergence zone south of the occlusion downdraft and ended in the low-level vorticity maximum.

In the Las Animas supercell the vorticity maximum extends all the way to the top of the data domain and shows almost on tilt with height (Fig. 2.27). The south to north vertical cross section shows that the downdraft that separates the vorticity maximum from the updraft associated with the rear-flank gust front (to its north) is very narrow (Fig. 2.27) and extends up to the top of the data domain. The air parcels that comprised the low-level vorticity maximum originate in this downdraft (Fig. 2.26).
Figure 2.24. Vertical vorticity (contoured), convergence (shaded) in the Sprague supercell at 2213 UTC at 200 m AGL, and a) Horizontal wind vectors at 200 m AGL, b) Horizontal vorticity vectors at 300 m AGL. Vertical vorticity is contoured every $0.01 \text{s}^{-1}$. Zero contours are omitted for clarity. Solid green line denotes the rear-flank gust front.
Figure 2.25. Same as in Fig. 2.24 but for 0030 UTC.
Figure 2.26. Backward trajectories of the air parcels that comprised the low-level vorticity maximum in the Las Animas supercell at 0030 UTC. a) view from the south b) view from the west.
Figure 2.27. West to east, and south to north vertical cross sections through the low-level vorticity maximum \((x = 7.6 \text{ km}, y = 9.8 \text{ km})\) in Sprague supercell. Vertical vorticity contoured every 0.01 \(\text{s}^{-1}\), zero contour line omitted for clarity. Vertical velocity shaded. Units of vertical velocity are \(\text{m s}^{-1}\). a) west to east at \(y = 9.8 \text{ km}\). b) south to north at \(x = 7.6 \text{ km}\).
2.3.4 Summary of the low-level mesocyclone characteristics in the observed nontornadic supercells

This subsection summarizes the characteristics of low-level kinematic fields in the observed nontornadic supercells and compares them to the low-level kinematic fields of tornadic and nontornadic supercells found in literature. Azimuthally averaged tangential velocity (Fig. 2.28a) at the times of maximum rotation at the lowest analysis levels (Figs. 2.12, 2.17, and 2.25) in the three observed low-level mesocyclones shows that the azimuthally averaged tangential velocity peaks at the radii of about 0.5 km with the maxima ranging from about 8 m s\(^{-1}\) to 12 m s\(^{-1}\). Because the features of a scale smaller than about 1 km are not well resolved in the analyses, the values of azimuthally averaged variables at radii smaller than about 500 m the smoothing could be influencing results. The observed values of azimuthally averaged tangential velocity (Fig. 2.28a) are generally smaller than the azimuthally averaged tangential velocity in the Garden City and Hays supercells (Fig. 2.28b). Note that because the low-level mesocyclones in the Brownfield, Sprague and Las Animas supercells had diameters of about 4 km or less, the azimuthally averaged values of tangential velocity (and other tangentially averaged variables discussed in this subsection) were calculated only for the radii associated with the mesocyclones (up to 2 km).

Circulation around the circles centered at the axis of the low-level mesocyclones (Fig. 2.29a) in the Brownfield supercells and the Sprague supercells increases almost linearly with increasing radius and reaches about 0.1 km\(^2\) s\(^{-1}\) at the radius of 1.9 km. This is only 20% less than the observed circulation at 500 m AGL at the same radius in the McLean, Texas supercell half an hour before tornadogenesis, but much smaller than the observed circulation at the time of tornadogenesis (Fig. 2.29b).

The intensity of the convergence around the mesocyclone axis of rotation is estimated by calculating the azimuthally averaged radial velocities (Fig. 2.30). Everything else being equal, the net convergence of air around the mesocyclone axis of rotation would indicate the vertical vorticity increase through vortex stretching, while the net divergence would lead to the vertical vorticity decrease. In the Brownfield supercell, the azimuthally averaged radial velocities are slightly nega-
Figure 2.28. Azimuthally averaged tangential velocity calculated around the centers of observed mesocyclones as a function of radius. a) The Brownfield, Sprague and Las Animas mesocyclones at the lowest analysis level, b) the Garden City (600 m AGL) and Hays (800 m AGL) supercells. The dashed line represents the profile of a combined Rankine vortex with the same maximum tangential wind speed. Adapted from Wakimoto and Cai 2000.
Figure 2.29. Circulation calculated around the centers of observed mesocyclones as a function of radius. a) The Brownfield, Sprague and Las Animas mesocyclones at the lowest analysis level, b) The McLean tornadic supercell at 500 m AGL. From Dowell and Bluestein 2002.

tive in the closest 0.9 km, indicating a weak convergence of air toward the axis of rotation; at radii larger than 0.9 km the azimuthally averaged radial velocities are positive indicating the net divergence of air away from the axis of rotation. A similar pattern is observed in the Las Animas supercell, the azimuthally averaged radial velocities are slightly negative at the radii less than 0.5 km, indicating a weak convergence of air toward the axis of rotation; and at radii larger than 0.5 km the azimuthally averaged radial velocities are positive indicating the net divergence of air away from the axis of rotation. In the Sprague case, the azimuthally averaged radial velocities are always negative and decreasing to -5 m s$^{-1}$ at the radius of 1.7 km indicating a relatively strong convergence of air toward the axis of rotation.
Figure 2.30. Azimuthally averaged radial velocity around the center of mesocyclone scale rotation as a function of radius at the lowest analysis level.

Note that the low-level mesocyclone in the Sprague supercell was the strongest of the three: 0.10 s$^{-1}$ peak vertical vorticity in the Sprague supercell, compared to 0.05 s$^{-1}$ peak low-level vertical vorticity in the Las Animas supercell and 0.06 s$^{-1}$ peak low-level vertical vorticity in the Las Animas supercell (Figs. 2.12a, 2.17a and 2.25a).

Centrifugal acceleration experienced by the air parcels in the low-level mesocyclone shows that although very different in their structure, all three observed
low-level mesocyclones have the peak in the centrifugal acceleration at the radius of about 0.3 km (Fig. 2.31). For the flow that is in cyclostrophic balance, the centrifugal acceleration is proportional to the force that the radial wind component would need to overcome to in order to stretch the vertical vorticity. A similar profile of the centrifugal forcing contributions to the radial velocities of the air parcels entering a tornado-like vortex has been found in the idealized axisymmetric numerical simulations of tornadogenesis (Markowski et al. 2003). In their numerical simulations the centrifugal forcing contribution to the radial velocity was the dominant positive (outflow) forcing for the radial velocity in the vicinity of the axis of rotation.

To summarize, the two observed supercells, the Watertown (Fig. 2.4) supercell and the Amarillo supercell (Fig. 2.5) had no downdrafts and no rotation developed in low levels. In the Brownfield, Sprague and Las Animas supercells, the low-level mesocyclone has been found to extend all the way down to the lowest analysis level.
The low-level mesocyclone in the Brownfield and Sprague supercells was found at the tip of the bent-back gust front. The Brownfield and the Sprague supercell had the back-bent gust front structure similar to the gust front structure observed in the tornadic supercells (e.g., Wurman et al. 2007; Marquis et al. 2008). The dual gust front structure observed in the Sprague supercell (e.g., Fig. 2.20) is very similar to the dual gust front structure observed in the Kiefer tornadic supercell (Fig. 1.4). In the Las Animas supercell, a short-lived low-level mesocyclone was observed. The recurring feature found at different analysis times in the Brownfield and Sprague supercells is the shallowness of the vorticity maxima associated with the low-level mesocyclone axis of rotation (e.g., Figs. 2.13 and 2.21).

The low-level kinematic structure of the Las Animas supercell is distinctly different than the low-level kinematic structure of the Brownfield and Sprague supercells. In the Las Animas supercell the gust front is not bent back as in aforementioned cases. The low-level mesocyclone in this case is completely separated from the rear-flank gust front by the rear-flank downdraft. Although the whole rear-flank gust front was not observed, it seems that the rear-flank gust front was located well ahead of the main updraft which may have contributed to the rapid dissipation of the storm just a few minutes after the low-level mesocyclone developed.

Strong horizontal vorticity is observed along the rear-flank gust fronts in all three cases (Figs. 2.12b, 2.17b, and 2.25b). The horizontal vorticity vectors are mostly parallel to the gust front. This is indicative of the strong baroclinic generation of horizontal vorticity in the baroclinic zone associated with the rear-flank gust front. In the Sprague supercell, the horizontal vorticity vectors (Fig. 2.17b) suggest that the horizontal vorticity is generated baroclinically both along the primary gust front, and along the interface between the rear-flank downdraft and the secondary gust front.

The vortex line analysis elucidates the three-dimensional structure of vorticity fields in the observed supercells. In the Brownfield and Sprague supercells the vortex line structure is consistent with the tilting of baroclinically generated vortex lines into the vertical by an updraft. The vortex lines emanating from the vorticity maxima that are located near the centers of low-level mesocyclones form arches (as low as 100 m AGL).
that rise up to about 2 km AGL and then descend into a region of positive vertical vorticity located to the south of the mesocyclones (e.g., Fig. 2.15). The vortex lines that emanate from the vorticity maxima located on the gust front below 500 m AGL also form arches that descend down to the regions of negative vertical vorticity. In contrast, the vortex lines that emanate from the vorticity maxima located on the gust front at or above 1000 m AGL typically originate in the environmental or forward flank air and are tilted upward by the updraft (e.g., Fig. 2.23). This indicates that the horizontal vorticity close to the ground (below about 1 km AGL) is generated baroclinically at the interface between the rear-flank downdraft and the updraft associated with the rear-flank gust front. At heights greater than 1 km AGL, the tilting of horizontal vortex lines that originate in the environmental air also contribute to the vertical vorticity generation.
CHAPTER 3

Idealized three-dimensional numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts

3.1 Rationale

The lack of thermodynamic observations is a major limitation of the observational part of this study. Thermodynamic retrievals from the dual-Doppler wind syntheses are found to be insufficiently accurate for a rigorous quantitative analysis (for details, see the Appendix). In the second part of this study, a series of idealized, dry three-dimensional numerical simulations are used to investigate the relationship between the low-level thermodynamics and kinematics of supercells, assuming that the idealized simulations can replicate the evolution of the low-level kinematic fields observed in actual supercells. The idealized simulations emulate the genera-
tion of near-surface rotation beneath supercell-like (i.e., helical) updrafts in a way consistent with our present understanding of the importance of a downdraft in environments in which vertical vorticity is initially absent at the surface.

3.2 Experiment design

Idealized numerical simulations of the rear-flank region of supercells are performed using the Bryan Cloud Model (CM1; Bryan and Fritsch 2002). The model is initialized using a semicircular hodograph (McCaul and Weisman 2001) such that

\[ v(z) = \exp(1)A \frac{n z}{H} \exp \left( -\frac{n z}{H} \right) \]  

(3.1)

\[ u(z) = \text{sgn}(z - z_0) \{ A^2 - [v(z)]^2 \} \]  

(3.2)

where \( u \) is the zonal component of the wind vector, \( z \) is the vertical coordinate, \( A \) is the hodograph radius, \( n \) is the profile compression parameter, \( H \) is the vertical scale, \( v \) is the meridional component of the wind vector and \( z_0 \) is the height where \( v(z) \) has a maximum. In the simulations performed herein the hodograph (Fig. 3.1) is constructed using the following values: \( A = 8 \text{ m s}^{-1}, n = 3, H = 6 \text{ km} \).

The potential temperature, \( \theta \), at the surface is set to 300 K and increased 1 K per 1 km in the lowest 10 km AGL, and 10 K per 1 km above 10 km AGL. The domain size is \( 25 \times 25 \times 18 \text{ km}^3 \). Grid spacing is 100 m in the horizontal. A stretched grid spacing was used in the vertical whereby \( \Delta z = 100 \text{ m} \) in the lowest 1 km AGL, \( \Delta z = 250 \text{ m} \) between 1 km and 8 km AGL, and \( \Delta z = 400 \text{ m} \) above 8 km AGL. Water vapor mixing ratio is set to zero throughout the domain. Open boundary conditions are used at the model lateral boundaries, while the free-slip conditions are used at the lower boundary. The influence of varying surface drag on the low-level rotation generation was not investigated, so the free-slip lower boundary is used in order to isolate the influence of different thermodynamic conditions on the dynamics of near-ground vorticity generation.

In order to simulate the latent heat release in the updraft of a supercell, a cylindrical heat source is located in center of the domain and is activated at the simulation start. The heat source is given by an additional forcing, \( S_w \), in the
potential temperature tendency equation

\[
S_w = \begin{cases} 
A_1 \left( 1 - \frac{(x-x_1)^2 + (y-y_1)^2}{R_1^2} \right) \left( 1 - \frac{(z-z_1)^2}{Z_1^2} \right) & \text{if } (x-x_1)^2 + (y-y_1)^2 \leq R_1^2 \text{ and } (z-z_1)^2 \leq Z_1^2 \\
0 & \text{otherwise}
\end{cases}
\]  

(3.3)

where \( x_1 = 12.5 \text{ km}, \ y_1 = 12.5 \text{ km}, \ z_1 = 4.75 \text{ km}, \ R_1 = 3 \text{ km}, \ Z_1 = 4 \text{ km}, \) and \( A_1 = 0.05 \text{ K s}^{-1}. \)

Given a fixed heat source and the resulting stationary updraft, the hodograph implies purely streamwise environmental vorticity at all levels; thus, a helical, supercell-like updraft results. By \( t = 900 \text{ s} \) the simulated updraft reaches a steady state, and the midlevel mesocyclone is colocated with the updraft (Figs. 3.2, and 3.3). Note that the vertical vorticity is zero near the ground (Fig. 3.3) at that time because the vertical vorticity is generated solely through the tilting of the horizontal vorticity by the updraft (Figs. 3.4). Note the helical structure of the vortex lines owing to baroclinic vorticity generation around the updraft. The updraft tilts eastward with height owing to almost purely westerly winds in mid
to upper levels (Fig. 3.1). At $t = 900$ s, a cylindrical heat sink that is located 1 km to the west of the heat source in the low-levels is activated in order to simulate the rear-flank downdraft. The location of the heat sink is chosen to be in the region where the rear-flank downdraft is commonly observed in supercell thunderstorms. The heat sink is given by adding an additional term, $S_c$, to the potential temperature tendency equation

$$S_c = \begin{cases} 
A_2 \left(1 - \frac{(x-x_2)^2 + (y-y_2)^2}{R_2^2}\right) \left(1 - \frac{(z-z_2)^2}{Z_2^2}\right) & \text{if } (x-x_2)^2 + (y-y_2)^2 \leq R_2^2 \text{ and } (z-z_2)^2 \leq Z_2^2 \\
0 & \text{otherwise}
\end{cases}$$

(3.4)

where $x_2 = 11.5$ km, $y_2 = 12.5$ km, $z_2 = 0.0$ km, $R_2 = 1$ km, $Z_2 = 3$ km, and $A_2$ is the heat sink amplitude.

By varying the amplitude of the heat sink I am able to change the temperature of the rear-flank downdraft, and consequently the temperature of the storm outflow. A similar model setup has been used previously (Walko 1993) to study the generation of near-surface rotation beneath supercell-like updrafts. Walko used a stationary heat source which was inconsistent with his environmental wind profile (a unidirectional hodograph, resulting in a splitting storm rather than a stationary storm consistent with a stationary heat source). The approach used in this study is different in the fact that it uses a semicircular hodograph to initialize the storm. Such a wind profile would produce an almost stationary supercell and is in a better agreement with the choice of a stationary heat source and a stationary heat sink. Owing to the horizontal and vertical grid spacing of 100 m in the low levels one cannot expect to resolve tornadoes but one can expect that circulations on the “tornado-cyclone” scale are well resolved.
3.3 Results

3.3.1 Simulations with weak cold pools

In the simulations where the heat sink amplitude is small (0.02 K s\(^{-1}\) in this example), the cold pool beneath the supercell-like updraft is weak and short-lived (5–20 minutes) and the low-level circulation does not persist (Fig. 3.5). This is very simi-
Figure 3.3. A west to east vertical cross section through the simulated updraft ($y = 14$ km). Vertical vorticity (red) is contoured every $0.005 \text{ s}^{-1}$. Vertical velocity (black) is contoured every $5 \text{ m s}^{-1}$. Zero contours are omitted for clarity.

lar to the observed Watertown and Amarillo supercells where the midlevel rotation is present but there is no storm outflow nor rotation observed in the low levels of the storm (Figs. 2.4, and 2.5). This represents the first tornadogenesis failure mode and is found in the simulations where the heat sink amplitude is small. The observations of the thermodynamic properties of rear-flank downdrafts (Markowski et al. 2002) have shown that supercells that produced significant tornadoes were
Figure 3.4. Vortex lines passing through the vertical vorticity maximum at $z = 4$ km AGL. a) projected onto $x$-$y$ plane, b) projected onto $y$-$z$ plane. Vortex lines that originate in the environment are tilted by the updraft into the vertical.

characterized by small deficits of virtual, and equivalent potential temperatures (Fig. 1.3). The results of the numerical simulations described herein, indicate that if the cold pool is too weak and short lived, the rotation near the ground does not develop.

Although this tornadogenesis failure mode is easy to identify visually (weak downdraft), the operational ground based radars (such as the National Weather
Service WSR-88D radars) that usually observe supercells at large distances may only be able to scan midlevels of the storm, so the nontornadic supercell of this type is still hard to identify with the operational Doppler radars.

### 3.3.2 Simulations with moderate cold pools

In the simulations with moderate cold pools the rear-flank gust front bends back around the circulation similar to the observed tornadic supercells (e.g., Wurman et al. 2007; Marquis et al. 2008). A similar “bent-back” gust front structure is also observed in the Brownfield and the Sprague supercells.

In the first simulation presented in this subsection (Fig. 3.6), the heat sink amplitude, $S_c$, is set to $0.033 \text{ K s}^{-1}$. The circulation in the low levels evolves into a gust front that bends counterclockwise at its northern tip. The development of such a circulation starts with vertical vorticity forming along the gust front (Fig. 3.6a). The vorticity maximum intensifies (Figs. 3.6b) and a closed circulation around the vorticity maximum is formed while the gust front bends around the rear-flank downdraft (Figs. 3.6c and d). The whole process takes only 10–15 minutes. At the time of maximum low-level vertical vorticity, the gust front is bent-back and the vortex is found underneath the main updraft. The maximum potential temperature deficits in the vicinity of the vortex are about -2 K and the maximum vertical vorticity is about 0.9 s$^{-1}$. Vortex lines emanating from the vorticity maximum located at the bent-back part of the rear-flank gust front rise vertically into the updraft and extend all the way to the top of the updraft (Figs. 3.7 and 3.8). Trajectories of air parcels that comprise the near-ground vorticity maximum (Fig. 3.9) originate in the cold pool air and are lifted up in the updraft. This simulation is believed to emulate the processes and structures present in a tornadic supercell.

In the second simulation presented in this subsection, the heat sink amplitude, $S_c$, is increased to $0.067 \text{ K s}^{-1}$, and it leads to the larger potential temperature deficits behind the rear-flank gust front (Fig. 3.10). As in the previous simulation the vertical vorticity first develops along the rear-flank gust front Fig. 3.10a). The gust front bends back but the vortex does not intensify as much as in the previous simulation. The potential temperature deficits in the vicinity of the vortex are about -4 K and the maximum vertical vorticity is about 0.6 s$^{-1}$. In this
simulation the “bent-back” gust front structure does not bend back far into the rear flank as in the previous simulation. The vertical vorticity near the ground is weaker than in tornadic simulations. The vortex lines that emanate from the low-level vorticity maximum form arches that rise up to 3 km AGL and then descend (Figs. 3.11 and 3.12). Trajectories of air parcels that comprise the near-ground vorticity maximum (Fig. 3.13) originate in the cold pool air and are lifted up only up to about 1 km AGL and then descend back into the cold pool. This suggests that the vertical vorticity was not stretched into a tornado by the main updraft owing to the excessively negatively buoyant outflow air. This simulation represents the second mode of tornadogenesis failure.

This type of nontornadic supercell is hard to identify both visually and by the operational Doppler radars. A well-developed rotation is observed in the low levels of the supercell. Because of that, this supercell type is visually very similar to tornadic supercells.

In the model, the strength of the simulated vortices can be increased not only by weakening the cold pool strength by reducing the heat sink amplitude, but also by increasing the intensity of the low-level updraft (not shown), for example, by increasing the vertical wind shear in the initial wind profile, or by increasing the intensity of the heat source. Herein, I only vary the intensity of the heat sink in order to compare the near-ground rotation generation associated with different cold pool characteristics to the low-level rotation observed in the nontornadic supercells.
Figure 3.5. Potential temperature perturbation (shaded) and vertical vorticity (contoured, black) at 50 m AGL in a simulation with no low-level rotation. Vertical velocity is contoured (green) at 450 m AGL for orientation purposes. Vertical vorticity is contoured every 0.01 s$^{-1}$. Vertical velocity is contoured every 2 m s$^{-1}$. Zero contours are omitted for clarity. a) at $t = 1200$ s, b) at $t = 1500$ s, c) at $t = 1800$ s, d) at $t = 2100$ s.
Figure 3.6. Potential temperature perturbation (shaded) and vertical vorticity (contoured, black) at 50 m AGL. Vertical velocity is contoured (green) at 450 m AGL for orientation purposes. Vertical vorticity is contoured every 0.01 s$^{-1}$. Vertical velocity is contoured every 2 m s$^{-1}$. Zero contours are omitted for clarity, negative values are dashed. Time $t = t_0$ denotes the “tornadogenesis” time. a) at $t = t_0 - 15$ min, b) at $t = t_0 - 10$ min, c) at $t = t_0 - 5$ min, d) at $t = t_0$. 
Figure 3.7. Vortex lines passing through a low-level vorticity maximum located on the bent-back part of the rear-flank gust front in the “tornadic” numerical simulation. View is from the south-southeast. The potential temperature perturbations (purple) and the vertical vorticity (black) are contoured for the orientation purposes.
Figure 3.8. Vortex lines passing through a low-level vorticity maximum located on the bent-back part of the rear-flank gust front in the “tornadic” numerical simulation. View is from the east. The red horizontal line denotes 5 km AGL.
Figure 3.9. Air parcel trajectories emanating from the near-ground vorticity maximum in the “tornadic” numerical simulation projected onto $x - z$ plane. The red horizontal line denotes 2 km AGL.
Figure 3.10. Potential temperature perturbation (shaded) and vertical vorticity (contoured, black) at 50 m AGL in the “nontornadic” numerical simulation with moderate cold pool. Vertical velocity is contoured (green) at 450 m AGL for orientation purposes. Vertical vorticity is contoured every $0.01 \, \text{s}^{-1}$. Vertical velocity is contoured every $2 \, \text{m s}^{-1}$. Zero contours are omitted for clarity, negative values are dashed. Time $t = t_0$ denotes the “tornadogenesis failure” time. a) at $t = t_0 - 15 \, \text{min}$, b) at $t = t_0 - 10 \, \text{min}$, c) at $t = t_0 - 5 \, \text{min}$, d) at $t = t_0$. 
Figure 3.11. Vortex lines passing through a low-level vorticity maximum located on the bent-back part of the rear-flank gust front in the numerical simulation with a moderate strength cold pool. View is from the south-southeast. The potential temperature perturbations (purple) and the vertical vorticity (black) are contoured for the orientation purposes.
Figure 3.12. Vortex lines passing through a low-level vorticity maximum located on the bent-back part of the rear-flank gust front in the “nontornadic” numerical simulation with moderate cold pool. View is from the east. The red horizontal line denotes 3 km AGL.
Figure 3.13. Air parcel trajectories emanating from the near-ground vorticity maximum in the “nontornadic” numerical simulation with a moderate cold pool projected onto x - z plane. The red horizontal line denotes 2 km AGL.
3.3.3 Simulations with strong cold pools

In the simulations where the heat sink amplitude is large (greater than 0.10 K s\(^{-1}\)), the simulations develop a strong cold pool (temperature deficits larger than 8 K) beneath the supercell-like updraft (Fig. 3.14). The gust front surges far ahead (1 – 2 km) of the main updraft and very shallow (up to 200 m AGL) vortices develop along the gust front (Figs. 3.15 and 3.16). The vortex lines emanating from the low-level vorticity maxima turn almost immediately horizontal and southeastward. The vortex lines in these case do not interact with the main updraft.

Similarly to the case of tornadogenesis failure mode associated with weak downdrafts, this type of nontornadic supercell is also easy to identify visually because of its strong outflow. However, because the operational ground based radars (such as the National Weather Service WSR-88D radars) typically observe supercells at large distances and are not able to scan the low levels of the storm, they cannot detect the rear-flank gust front surging ahead of the main updraft. Thus, this type of tornadogenesis failure is also hard to identify with the operational Doppler radars.
Figure 3.14. Potential temperature perturbation (shaded) and vertical vorticity (contoured, black) at 50 m AGL in the numerical simulation with the extremely strong cold pool. Vertical velocity is contoured (green) at 450 m AGL for orientation purposes. Vertical vorticity is contoured every 0.01 $s^{-1}$. Vertical velocity is contoured every 2 m $s^{-1}$. Zero contours are omitted for clarity, negative values are dashed. a) at $t = 1200$ s, b) at $t = 1800$ s, c) at $t = 2100$ s, d) at $t = 2400$ s.
Figure 3.15. Vortex lines passing through a low-level vorticity maximum located on the rear-flank gust front in the “nontornadic” numerical simulation with a strong cold pool. View is from the south. The potential temperature perturbations (purple) and the vertical vorticity (black) are contoured for the orientation purposes.
Figure 3.16. Vortex lines passing through a low-level vorticity maximum located on the bent-back part of the rear-flank gust front projected onto y-z plane. The red horizontal line denotes 2 km AGL.
3.4 Summary

The results of numerical simulations reveal three tornadogenesis failure modes depending on whether the rear-flank downdraft develops and on the cold pool intensity. The first tornadogenesis failure mode is observed in the simulations with the weak rear-flank downdraft where the cold pool does not develop or a weak cold pool (potential temperature deficits less than 2 K) develops only briefly. Note that in the absence of the downdraft (i.e., before the heat sink is turned on) there is no vertical vorticity near the ground.

The tornadogenesis is observed in the simulations where the cold pool potential temperature deficit near the bent-back gust front is about 2–4 K. In this case, most of the vortex lines that emanate from the near-ground vorticity maximum enter the main updraft and interact with the midlevel mesocyclone. The second tornadogenesis failure mode is found when the cold pool intensity is about 4–8 K. In that case, the gust front bends back in a similar way like in the tornadogenesis simulations. In this case the vortex lines emanating from the near-ground vorticity maximum form arches that rise up to 3 km AGL and then descend in the rear flank to the south of the vorticity maximum. The third tornadogenesis failure mode is observed when the cold pool potential temperature deficit is 8 K or higher. In that case, the gust from rushes ahead of the main updraft. The vertical vorticity is generated along the gust front but the vortices are very shallow. The vortex lines in this case rise only up to 200 m AGL before tilting into the horizontal.
A conceptual model of a nontornadic supercell

This Chapter compares the vortex line structure of observed nontornadic supercells to the vortex line structure found in the numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts, and presents a conceptual model of a nontornadic supercell. The vortex line structure in the observed supercells with well-developed low-level rotation analyzed herein (e.g., Fig. 2.15) is similar to the vortex line structure found in the idealized numerical simulations (e.g., Figs. 3.11 and 3.12). The vortex line structure of the Brownfield supercell closely resembles the vortex line structure found in the nontornadic simulations with a moderate outflow. The vortex lines in the Sprague supercell and the Las Animas supercell form arches emanating from the low-level vorticity maxima located on the rear-flank gust front. Unfortunately owing to the fact that the dual-Doppler syntheses are possible only up to about 2.5 km AGL we cannot observe whether the vortex lines emanating from the vorticity maxima located in the centers of low-level mesocyclones form arches. These vortex lines go straight up until they...
reach the top of the three-dimensional dual-Doppler wind synthesis.

The results of a series of numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts suggest three possible mechanisms for tornadogenesis failure (Fig. 4.1). I propose a conceptual model of a nontornadic supercell whereby the first tornadogenesis failure mode (Fig. 4.1a) is observed in the supercells with no rear flank downdrafts or weak rear-flank downdrafts where the cold pool does not develop or just a weak cold pool (potential temperature deficits less than 2 K) develops briefly (e.g., Fig. 3.5). In this case, the cold pool does not interact with the updraft and there is no low-level rotation present. Tornadogenesis occurs when there is enough outflow to create a cold pool that interacts with the updraft whereby the baroclinically generated horizontal vorticity is tilted and subsequently stretched into a tornado by the updraft (Fig. 4.1b). Note that in this case, the vortex lines that form arches are still observed.

The second tornadogenesis failure mode (Fig. 4.1c) is observed when the cold pool intensity is about 4–8 K (e.g., Fig. 3.10). In this case, the gust front bends back in a similar way like in the simulations that resulted in tornadogenesis (Fig. 3.6). The vortex lines emanating from the near-ground vorticity maximum form arches that rise up and then descend in the rear flank to the south of the vorticity maximum but the vortex never reaches the tornado intensity. The third tornadogenesis failure mode (Fig. 4.1d) is observed when the cold pool potential temperature deficit is 8 K or higher (e.g., Fig. 3.14). For such large cold pool deficits, the gust front rushes ahead of the main updraft. The vertical vorticity is generated along the gust front but the vortices are very shallow. The vortex lines in this case rise only up to about 200 m AGL before tilting into the horizontal and become parallel to the rear-flank gust front.

The Brownfield (Fig. 2.6), and the Sprague supercells (Fig. 2.7) could be classified as the nontornadic supercells that are associated with moderate outflows (Fig. 4.1c), while the Watertown supercell (Fig. 2.4), and the Amarillo supercell (Fig. 2.5), are the examples of a supercell without (or with a negligible) outflow (Fig. 4.1a). The classification of the Las Animas supercell (Fig. 2.8) is harder to do, owing to the fact that the whole rear-flank gust front structure has not been observed suggesting that the gust front may have rushed well ahead of the rear-flank downdraft. If this was really the case, the Las Animas supercell could be
Figure 4.1. Vortex line and gust front structure of tornadic and nontornadic supercells. Red cylinder represents a rotating updraft. White/blue cylinders represent rear-flank downdrafts of varying thermodynamic characteristics (darker colors being colder). Light green lines denote environmental vortex lines. Dark green lines denote vortex lines created baroclinically in by the rear-flank downdraft. Solid black line represents the position of the rear-flank gust front.
classified as a supercell with a strong outflow (Fig. 4.1d).
This study examined the low-level kinematic fields in nontornadic supercells observed by a pair of high-resolution mobile ground-based radars (DOW). Five nontornadic supercells were analyzed, and three out of five had low-level rotation (the Brownfield supercell, the Sprague supercell, and the Las Animas supercell). The other two supercells (the Watertown supercell and the Amarillo supercell), had a circulation associated with the midlevel mesocyclone but no rear-flank downdrafts nor low-level circulation.

All three supercells that had well-developed low-level rotation had the low-level mesocyclones extending all the way down to the lowest analysis levels (100–200 m AGL). The Brownfield and the Sprague supercells had bent-back gust fronts with the centers of the low-level rotation located at the tips of the bent-back gust fronts. A dual gust front structure similar to the dual gust front structure of Kiefer tornadic supercell (Wurman et al. 2007b) was observed in the Sprague supercell. In the Las Animas supercell the low-level rotation was observed further back in the rear flank. In the Las Animas supercell the tornadogenesis failure seem to have occurred owing to the supercell dissipating right after the mesocyclone developed,
while in the Brownfield supercell, the tornadogenesis failure has not been observed, and the reason for that is the fact that the low-level mesocyclone was intensifying up until the end of the dual-Doppler radar deployment.

The circulation in the low-level mesocyclones in the observed nontornadic supercells that had well-developed low-level rotation have been found to be comparable to the circulation in the low-level mesocyclones of supercells that produced a tornado at the time about 30 minutes before tornadogenesis. The low-level mesocyclone observed in the Sprague supercell (which was the most intense out of three) had the strongest azimuthally averaged inward radial velocities and the smallest diameter ($\approx 2$ km). The peak azimuthally averaged tangential velocities found in the Sprague supercell were $\approx 12 \text{ m s}^{-1}$, which is comparable to the values found during tornadogenesis failure in the Hays, Kansas nontornadic supercell (Trapp 1999). However it is much less than the $22 \text{ m s}^{-1}$ observed during a tornadogenesis failure associated with a low-level mesocyclone that developed later within the same storm (Wakimoto and Cai 2000). The trajectories of the parcels that enter the low-level mesocyclone at the time of peak rotation in the Sprague supercell ascend only up to 500 m AGL indicating that they may not have been buoyant enough to could be stretched into a tornado.

The horizontal vorticity vectors indicate that there is baroclinic generation of horizontal vorticity along the rear-flank gust front in all three observed supercells with low-level rotation. The vortex line analysis shows that the vortex lines associated with the low-level mesocyclone axis of rotation rise up to the top of the radar data domain so their evolution above 1.5–2.5 km AGL cannot be assessed. Vortex lines emanating from the the low-level vertical vorticity maxima found along the rear-flank gust front, and in the vicinity of the low-level mesocyclone axis of rotation, form arches similar to the arches proposed by Straka et al. (2007).

Thermodynamic fields could not be retrieved from the DOW radar data with desired accuracy. Therefore, a series of idealized three-dimensional numerical simulations of the generation of near-surface rotation beneath supercell-like updrafts are performed in order to investigate if the kinematic aspects of the observed nontornadic supercells could be replicated, and if so, how the attendant thermodynamic conditions could influence the dynamics of near-ground vorticity generation. The results of numerical simulations suggest three possible tornadogenesis failure
modes.

The Brownfield supercell and the Sprague supercell are found to have similar low-level kinematic structure to the simulations with moderate outflow, whereas the Watertown, and Amarillo supercells are examples of supercells without (or with a negligible) outflow. The classification of the Las Animas supercell is harder, in part owing to the fact that it was observed only in the last 10 minutes of its lifetime and the whole rear-flank gust front was not observed.

A conceptual model of a nontornadic supercell is proposed whereby the first tornadogenesis failure mode is found in supercells with weak rear-flank downdrafts where the cold pool does not develop or only a weak cold pool develops. The second tornadogenesis failure mode is found in supercells with moderate outflow. In this case, the gust front bends back in a similar way to tornadic supercells and the baroclinically generated horizontal vorticity is tilted into the vertical by the updraft but the vertical vorticity is never stretched into a tornado owing to excessively negatively buoyant outflow air. The supercells that could produce a tornado are associated with the moderate outflows that are not excessively negatively buoyant whereby a tornado forms when the vorticity is stretched by the main updraft. The third tornadogenesis failure mode is associated with supercells with strong outflow. In this case, the gust front rushes ahead of the main updraft. Although vertical vorticity is generated along the gust front by the tilting of baroclinically generated vertical vorticity, the vortices are shallow and do not interact with the main updraft.

One of the directions for future research should be focused on the observations of the surface and the upper air thermodynamic properties of the supercell rear-flank downdraft region. The simulations presented herein suggest that rather small differences in the thermodynamic properties of the rear-flank downdraft can make a difference between a tornadogenesis and a tornadogenesis failure. Also, understanding the cloud microphysics that may be responsible for the observed difference in the thermodynamic properties of the rear-flank downdrafts would be beneficial for our overall understanding of supercell dynamics.

The analysis of back trajectories of the air parcels that comprise low-level mesocyclones show that the air in the centers of low-level rotation often originates below the lowest analysis levels (100–200 m AGL). It would be useful to know whether
the extrapolations of the low-level wind fields to the ground are accurate enough to calculate back trajectories below the lowest trajectory levels and what is the sensitivity of the trajectory calculations to different extrapolation methods.

There are only 15 tornadic supercells observed by two Doppler radars (including the five nontornadic supercells analyzed in this study). Given the sheer number of supercells that occur annually this is a very small sample. Therefore, another research direction should be a continued collection and analysis of the dual-Doppler observations of both tornadic and nontornadic supercells in order to expand the sample.


Multi-pass Barnes objective analysis technique

The objectively analyzed radial velocity data are used to produce dual-Doppler wind syntheses, and comparisons are made between the kinematic fields of the reference simulation (which will be regarded as the truth) and those derived from dual-Doppler wind syntheses utilizing single- and multi-pass objectively analyzed synthetic radial velocity data. The improvement of multi-pass objective analyses on higher-order calculations such as buoyancy retrievals and trajectory calculations also is assessed.

The numerical simulation used to produce the synthetic radar data is performed using version 4.5.2 of the Advanced Regional Prediction System (ARPS; Xue et al. 2000, 2001), and is initialized with the composite sounding from the well-documented 20 May 1977 Del City, Oklahoma, supercell thunderstorm (Ray et al. 1981; Johnson et al. 1987). The simulation domain is 64 km \( \times \) 64 km \( \times \) 18 km, with the vertical and horizontal grid spacings set to 250 m (Fig. A.1). The simplified simulation design utilized warm rain microphysics and neglected surface
Figure A.1. ARPS model horizontal wind vectors and rainwater concentration plotted at $z = 250$ m at $t = 3660$ s. Filled black squares denote radar locations. The black circle denotes dual-Doppler lobe. The bigger black square denotes three-dimensional wind synthesis domain. The smaller black square (inscribed in the circle) is the domain used for perturbation pressure and buoyancy retrievals.

fluxes and radiation. A warm bubble is used to initiate the storm.

Synthetic radar data were generated for two radars (“R1” and “R2”, respectively) positioned at $(x,y) = (20$ km, $14$ km) and $(x,y) = (20$ km, $24$ km), whereby the southwest corner of the domain is the origin, “scanning” in $180^\circ$ sectors centered at $270^\circ$. At $t = 3660$ s, the low-level mesocyclone is near the center of the dual-Doppler lobe (defined as a region where the beam crossing angle is no less than
30° and no greater than 150°), to the west of the radars (Fig. A.1). The azimuthal and range sampling intervals of the synthetic radars are 1° and 100 m, respectively. Radial velocities are calculated at 15 different elevation angles between 0.5° and 22.5°. Thus, the radar positioning, resolution, and scanning strategy emulate that which might typify data collection in an actual dual-Doppler deployment of mobile radars within a field experiment (e.g., Ray et al. 1975; Brandes 1978; Beck et al. 2006; Wurman et. al. 2007a,b).

The radial velocity $V_i$, measured by the $i$th radar, is

$$V_i(r_i, \theta_i, \phi_i) = u \sin \theta_i \cos \phi_i + v \cos \theta_i \cos \phi_i + w \sin \phi_i \quad (A.1)$$

where $u$, $v$, $w$ are the model zonal, meridional, and vertical wind components, respectively, interpolated to the radar gate located at range $r_i$, azimuth angle $\theta_i$, and elevation angle $\phi_i$, using the trilinear interpolation. To simplify the error analysis it is assumed that particle fallspeeds are zero and that synthetic radar volumes were collected instantaneously (i.e., effects related to storm translation are not considered). Furthermore, no attempt was made to emulate power-weighted volumetric radar sampling. Neglecting the aforementioned error sources obviously is unrealistic, but it is necessary in order to isolate the effect of the objective analysis technique on the three-dimensional wind syntheses (Clark et al. 1980).

On the other hand, random errors are added to the synthetic radial velocity fields in order to add to their realism. The standard deviation of the random errors added to the radial velocities was $\sigma = 1 \text{ m s}^{-1}$ (Rabin and Zrnic 1980).

Objective analyses of the synthetic radial velocity data were performed using the distance-dependent weight function described by Barnes (1964). Radial velocities from R1 and R2 were interpolated to a 20 km $\times$ 20 km $\times$ 3 km grid having a horizontal and vertical grid spacing of 250 m. The radial velocity analyzed at a gridpoint is an adjustment of a background field by a weighted average of the difference between the observed radial velocities and the background field interpolated to the radial velocity observations, i.e,

$$V_{j,n} = V_{j,n-1} + \frac{\sum_{k=1}^{N} \omega_{jk,n}(V_k - V_{I_k,n-1})}{\sum_{k=1}^{N} \omega_{jk,n}} \quad (A.2)$$

where $V_{j,n}$ is analyzed radial velocity at the $j$th grid point after the $n$th pass,
$V_{j,n-1}$ is analyzed radial velocity at the $j$th grid point after the $(n-1)$th pass, $V_k$ is the $k$th radial velocity observation (with $N$ total observations within a “cutoff” radius $R_c$, beyond which the weight is insignificant), $V_{Ik,n-1}$ is $V_{j,n-1}$ interpolated to the location of the $k$th observation (using the same weight function as in the interpolation to the grid), and $\omega_{jk,n}$ is the weight assigned to the $k$th observation for the $j$th grid point. When $n = 1$, $V_{j,n-1} = V_{Ik,n-1} = 0$. The weight function $\omega_{jk,n}$ is defined as,

$$\omega_{jk,n} = \exp\left(-\frac{r_{jk}^2}{\kappa_0 \gamma^{n-1}}\right) \quad (A.3)$$

where $r_{jk}$ is the distance between the $j$th grid point and the $k$th radial velocity observation, $\kappa_0$ is the smoothing parameter of the first pass, and $\gamma$ is the convergence parameter which sets the response of the multi-pass scheme (Koch et al. 1983). The smoothing parameter, $\kappa_0$, of 0.34 km$^2$ was chosen following the recommendation of Pauley and Wu (1990) that the optimal smoothing parameter should be $\kappa_0 = (1.33 \Delta_d)^2$, where $\Delta_d$ in this case is the coarsest data separation in the synthetic radar data rather than the analysis grid spacing that Pauley and Wu considered. The extrapolation of data to grid points was not permitted.

Five experiments are presented herein: a 1-pass Barnes analysis, three 2-pass Barnes analyses ($\gamma = 0.1$, 0.3 and 0.9), and a 3-pass Barnes analysis ($\gamma = 0.3$). In all of the experiments, we used an isotropic, Barnes weight function. The theoretical response functions (Fig. A.2) (assuming continuous data) for the various Barnes filters show that response functions of multi-pass Barnes filters are steeper than the 1-pass response function, so the shorter wavelengths can be effectively subdued while almost fully retaining the longer wavelengths.

The theoretical response of the 1-pass analysis ($R_1$) is (Barnes 1964; Koch et al. 1983),

$$R_1 = \exp \left[-\kappa_0 \left(\frac{\pi}{\lambda}\right)^2\right] \quad (A.4)$$

where $\lambda$ is the wavelength of the input field. The response function of the $n$-pass analysis ($R_n$) for $n > 1$ can be obtained from (Koch et al. 1983),
Figure A.2. Theoretical Barnes filter response function ($\kappa_0 = 0.34 \, \text{km}^2$) for 1-pass filter (solid bold), 2-pass filter with $\gamma = 0.1$ (dotted), 2-pass filter with $\gamma = 0.3$ (dot-dashed), 2-pass filter with $\gamma = 0.9$ (dashed), and 3-pass filter with $\gamma = 0.3$ (thin solid).

\begin{equation}
R_n = R_{n-1} + (1 - R_{n-1}) \exp \left[ -\kappa_0 \gamma^{n-1} \left( \frac{\pi}{\lambda} \right)^2 \right] \tag{A.5}
\end{equation}

The actual amplitude response for our data will differ somewhat from the theoretical response and there also may be phase shifts because the data are discrete and bounded (Pauley and Wu 1990; Askelson and Straka 2005; Askelson et al. 2005). The theoretical responses are used as a guide in choosing our parameters, and
we use the root-mean-square error to measure the improvement in the analyzed field. Equivalent objective analyses of unbound and continuous data can be produced by using an infinite number of smoothing parameter combinations (Spencer et al. 2007). For discrete data within bounded domains (such as radar data) it is possible that the choice of parameter combinations in the two-pass or three-pass schemes will affect the response. However, this is not likely to be a significant problem for fairly regularly spaced discrete data (Spencer et al. 2007) such as synthetic radar data used herein.

After the objective analysis to obtain the radial velocities for both radars at all gridpoints, the three-dimensional winds were retrieved by integrating the anelastic mass continuity equation upward from the ground. Because extrapolation was forbidden in the objective analysis stage, wind data were not retrieved at the lowest grid level (z = 0 m) and it was assumed that $w = 0$ at $z = 0$. After the syntheses of the three-dimensional winds, the perturbation pressure and buoyancy were retrieved using the technique described by Hane and Ray (1985). Time derivatives were estimated using centered differences of syntheses obtained 60 s before and after the analysis time. The buoyancy and perturbation pressure retrieval involves solving an elliptic equation that is sensitive to the boundary conditions. These fields were retrieved in a square subdomain of the dual-Doppler lobe (Fig. A.1), in order to simplify the implementation of the (Neumann) boundary conditions.

To investigate how closely the retrieved horizontal and vertical wind components, vertical vorticity ($\zeta$), and horizontal divergence ($\nabla_h \cdot v$) match the simulation results (the truth), the lowest 1 km of analyses are compared to the numerical simulation results via rmse and linear correlation coefficients. The same is done for buoyancy and perturbation pressure, but only at $z = 750$ m. These quantities are only known to within a constant (that varies with height) owing to the Neumann boundary conditions used in the pressure and buoyancy retrieval (Hane and Ray 1985). Prior to comparing the model buoyancy and perturbation pressure fields with the retrieved buoyancy and perturbation pressure fields, a constant is added to the retrieved buoyancy and perturbation pressure field so that the mean buoyancy and perturbation pressure of the retrieval is equal to the mean buoyancy and perturbation pressure of the model output, respectively. All of the aforementioned comparisons are done at $t = 3660$ s, at which time a mature supercell is present in
the dual-Doppler domain (Fig. A.1).

Also, trajectory calculations are performed on both the model output and three-dimensional wind syntheses. For that purpose, the wind syntheses are obtained and the model output is sampled at \( t = 3600 \) s and every 60 s (to simulate the sampling rate of dual-Doppler deployments mentioned above) thereafter until \( t = 4500 \) s. The trajectories are initialized at \( t = 3600 \) s at \( z = 600 \) m within a 10 km × 10 km region centered in the dual-Doppler lobe. Initial positions are 1 km apart in both the \( x \) and \( y \) directions, giving a total of 121 trajectories, which are computed for 15 min using a 4th-order Runge-Kutta method and a timestep of 20 s. The average distance between the trajectories in the model output and the dual-Doppler analyses is computed at each timestep for those trajectories staying within the dual-Doppler lobe throughout the entire 15-min integration.

The values of the correlation coefficients (Table A.1) between the model and retrieved horizontal wind components exceed 0.96 for all objective analysis procedures. The rmse of the \( v \) wind component is almost twice that for the \( u \) component, apparently owing to the sampling geometry and/or the characteristics of the true wind field. The \( u \) rmse of 1.34 m s\(^{-1}\) (Table A.2) is very close to the random error added to the synthetic radar radial velocities. In the multi-pass wind syntheses the rmse of \( u \) is reduced by more than 50% in the example of the 3-pass analysis, but that particular analysis probably would be considered unacceptably noisy by most (Fig. A.3). The 2-pass analyses reduce the rmse by more than 30%, and are much less noisy. The improvements are found mostly in better representation of smaller-scale features (e.g., the narrow maximum in \( u \) located parallel to and northeast of the hook echo; \( x = 7 \) km, \( y = 16 \) km), and the location and amplitude of maxima and minima (e.g., the secondary minimum in the southeastern part of the analysis domain; \( x = 9 \) km, \( y = 5 \) km).

The rmse of \( v \) is reduced by more than 40% in multi-pass Barnes analyses compared to a 1-pass analysis. The improvements are again found in the depiction of smaller-scale features and maxima and minima of the \( v \) field (Fig. A.4). However, some analyses seem much noisier (e.g., the 2-pass analysis with \( \gamma = 0.1 \) and the 3-pass analysis with \( \gamma = 0.3 \)) than others (e.g., the 2-pass analysis with \( \gamma = 0.3 \) and 2-pass analysis with \( \gamma = 0.9 \)), although the rmse and correlation coefficients of those analyses are very similar.
Figure A.3. Zonal wind component at $z = 750$ m. Units are m s$^{-1}$. Zero contour is not shown. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$ (f) 3-pass analysis $\gamma = 0.3$. 
Figure A.4. Meridional wind component at $z = 750$ m. Units are m s$^{-1}$. Zero contour is not shown. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$, (f) 3-pass analysis $\gamma = 0.3$. 
Table A.1. Correlation coefficient between ARPS output and analyses in the lowest 1 km of analysis domain. Perturbation pressure ($p'$) and perturbation density potential temperature ($\theta'_\rho$) correlation coefficients are computed at $z = 750$ m.

<table>
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<th>2-pass, $\gamma = 0.3$</th>
<th>2-pass, $\gamma = 0.9$</th>
<th>3-pass, $\gamma = 0.3$</th>
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<td>$w$</td>
<td>0.86</td>
<td>0.86</td>
<td>0.89</td>
<td>0.89</td>
<td>0.84</td>
</tr>
<tr>
<td>$\zeta$</td>
<td>0.78</td>
<td>0.91</td>
<td>0.89</td>
<td>0.83</td>
<td>0.90</td>
</tr>
<tr>
<td>$\nabla_h \cdot v$</td>
<td>0.72</td>
<td>0.85</td>
<td>0.83</td>
<td>0.77</td>
<td>0.84</td>
</tr>
<tr>
<td>$p'$</td>
<td>0.90</td>
<td>0.95</td>
<td>0.95</td>
<td>0.96</td>
<td>0.97</td>
</tr>
<tr>
<td>$\theta'_\rho$</td>
<td>0.73</td>
<td>0.79</td>
<td>0.78</td>
<td>0.76</td>
<td>0.79</td>
</tr>
</tbody>
</table>

Table A.2. Root mean square errors in the lowest 1 km of analysis domain. Smoothing parameter in all three analyses is $\kappa = 0.34$ km$^2$. Units of $u$, $v$, $w$ are 1 m s$^{-1}$. Units of vertical vorticity ($\zeta$) and horizontal divergence ($\nabla_h \cdot v$) are 0.001 s$^{-1}$. Perturbation pressure ($p'$) and density perturbation potential temperature ($\theta'_\rho$) errors are computed at $z = 750$ m. Units of perturbation pressure are mb. Units of density perturbation potential temperature are K.

<table>
<thead>
<tr>
<th>Variable</th>
<th>1-pass</th>
<th>2-pass, $\gamma = 0.1$</th>
<th>2-pass, $\gamma = 0.3$</th>
<th>2-pass, $\gamma = 0.9$</th>
<th>3-pass, $\gamma = 0.3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u$</td>
<td>1.34</td>
<td>0.75</td>
<td>0.85</td>
<td>1.04</td>
<td>0.66</td>
</tr>
<tr>
<td>$v$</td>
<td>2.15</td>
<td>1.44</td>
<td>1.41</td>
<td>1.60</td>
<td>1.22</td>
</tr>
<tr>
<td>$w$</td>
<td>1.13</td>
<td>1.12</td>
<td>0.99</td>
<td>0.98</td>
<td>1.21</td>
</tr>
<tr>
<td>$\zeta$</td>
<td>2.85</td>
<td>1.87</td>
<td>2.02</td>
<td>2.46</td>
<td>1.96</td>
</tr>
<tr>
<td>$\nabla_h \cdot v$</td>
<td>2.44</td>
<td>1.86</td>
<td>1.92</td>
<td>2.21</td>
<td>1.97</td>
</tr>
<tr>
<td>$p'$</td>
<td>0.35</td>
<td>0.22</td>
<td>0.22</td>
<td>0.20</td>
<td>0.18</td>
</tr>
<tr>
<td>$\theta'_\rho$</td>
<td>1.45</td>
<td>1.16</td>
<td>1.44</td>
<td>1.83</td>
<td>0.79</td>
</tr>
</tbody>
</table>

Even more improvement in the depiction of small-scale features is visible in analyses of $w$ (Fig. A.5). The 1-pass analysis is not able to reproduce a narrow updraft ($x = 11$ km, $y = 4$ km) stretching along the rear-flank outflow and a narrow downdraft ($x = 11$ km, $y = 11$ km) found along the hook echo (Fig. A.5b). The analyses with smallest $w$ errors (Table A.2) are the 2-pass analyses with $\gamma = 0.9$ (Fig. A.5c) and $\gamma = 0.3$ (Fig. A.5d).

The correlation coefficient for derivative fields such as vertical vorticity and divergence is smaller than for the horizontal and vertical wind components. The 2-pass Barnes analysis with $\gamma = 0.1$ improves the correlation coefficient the most for vertical vorticity, from 0.78 (in the 1-pass analysis) to 0.91, and the correlation coefficient for divergence from 0.73 (in the 1-pass analysis) to 0.85. The rmse is
Figure A.5. Vertical wind component at $z = 750$ m. Units are m s$^{-1}$. Zero contour is not shown. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$, (f) 3-pass analysis $\gamma = 0.3$. 
reduced by 35% for vertical vorticity and by 25% for divergence.

The analyses of vertical vorticity (Fig. A.6) show that the 1-pass analysis damp
the amplitude of the vorticity maximum just east of the tip of the hook echo. The
multi-pass analyses depict the amplitude and location of maxima and minima of
vertical vorticity much better, although the 2-pass analysis with $\gamma = 0.9$ does not
perform as well as other multi-pass analyses.

From the analyses of the horizontal divergence field (Fig. A.7), it is clear that
the 1-pass analysis only depicts the convergence region of the main updraft, but
none of the small-scale features found along and southwest of the hook echo. The
multi-pass analyses depict small-scale maxima and minima of horizontal divergence
better, especially the 2-pass analyses with $\gamma = 0.1$ and $\gamma = 0.3$.

The perturbation pressure field (Fig. A.8) is characterized by a significant pres-
sure minimum in the northeastern part of the domain, with a fairly flat pressure
field in the rest of the domain. All analyses located the pressure minimum correctly,
with multi-pass analyses having smaller amplitude errors and better representation
of the perturbation pressure gradient southwest of the perturbation pressure min-
imum. The correlation coefficients (Table A.1) exceed 0.96 in all of the analyses,
and the rmse (Table A.2) is relatively low ($< 0.35$ mb in the 1-pass analysis and
$< 0.22$ mb in the multi-pass analyses).

The buoyancy field analyses [Fig. A.9; density potential temperature pertur-
bations are used here to represent the buoyancy field (Emanuel 1994)] are in gen-
eral agreement with the model output but lack the small-scale features found in
the model. The buoyancy retrieval performed using the multi-pass analyses show
slightly improved correlation coefficients (Table A.1) and rmse (Table A.2) in all
but one multi-pass analysis (2-pass, $\gamma = 0.9$). The 3-pass analysis has the smallest
rmse (0.79 K) of all analyses, but the 2-pass analyses (especially the one with with
$\gamma = 0.1$) capture the buoyancy gradients near the storm better. The buoyancy
retrievals may be adequate to characterize the general properties of the outflow,
but probably are not sufficiently accurate for calculations of baroclinic vorticity
generation along trajectories.

The results of trajectory calculations are shown in Fig. A.10. Both 2-pass
analyses significantly reduce the errors, especially in the later stages of integration,
which is expected since the errors of $u$, $v$, and $w$ are smaller in multi-pass analyses.
Figure A.6. Vertical vorticity at $z = 750$ m. Units are $0.01 \text{s}^{-1}$. Zero contour is not shown. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$ (f) 3-pass analysis $\gamma = 0.3$. 
Figure A.7. Horizontal wind divergence at $z = 750$ m. Units are $1 \text{s}^{-1}$. Zero contour is not shown. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$ (f) 3-pass analysis $\gamma = 0.3$. 
Figure A.8. Perturbation pressure at $z = 750$ m. Units are mb. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$ (f) 3-pass analysis $\gamma = 0.3$. 
**Figure A.9.** Perturbation density temperature at $z = 750$ m. Units are K. Area where 250m rainwater concentration is larger than 1 g/kg is shaded for orientation purposes. (a) ARPS, (b) 1-pass analysis, (c) 2-pass analysis $\gamma = 0.9$, (d) 2-pass analysis $\gamma = 0.3$, (e) 2-pass analysis $\gamma = 0.1$ (f) 3-pass analysis $\gamma = 0.3$. 

\( \theta' \) contour interval: 1 K
compared to the 1-pass analysis. The average errors of the trajectories in the 1-pass analysis after 5 min, 10 min, and 15 min, are 853 m, 2425 m, and 4346 m, respectively. The 2-pass analysis with $\gamma = 0.3$ at the same integration times has errors of 651 m, 1566 m, and 2675 m, respectively. It is interesting that the 3-pass analysis has the smallest trajectory errors during the first 7 min of integration (Fig. A.10). After that, until the end of integration, the 2-pass analysis with $\gamma = 0.3$ has the smallest trajectory errors. Though I am uncertain of the robustness of the accuracy of the trajectories computed for the time interval and wind fields unique to these experiments, I am confident that the trajectories computed from wind syntheses derived from multi-pass objective analyses are superior to those computed from wind syntheses derived from single-pass objective analyses.

These results suggest that in many cases the 2-pass Barnes objective analyses have smaller rmse and are better correlated to the model output than the 1-pass Barnes analysis. The reduction of rmse and the higher correlation coefficients are
more pronounced in the first-order derivatives of the wind field, such as vertical vorticity and divergence, than in the individual wind components themselves. Even better improvement can be seen in trajectory calculations with 2-pass analyses, which benefit from more accurate horizontal and vertical wind fields. Correct representation of vertical vorticity and horizontal divergence is especially important in analyzing supercell thunderstorms, which are among the most popular targets of mobile radars in field experiments. Together with a better representation of the three-dimensional wind components and trajectories, these improved analyses can improve qualitative and quantitative analysis of kinematic structure of supercells thunderstorms. Unfortunately, even with this improved accuracy in the representation of the three-dimensional wind components and trajectories, calculations of vorticity budgets along trajectories are still not accurate enough (not shown).

Also, the comparison between 1-pass and 2-pass analyses reveals that the amplitudes of the maxima and minima of the analyzed fields will depend on the choice of the analysis technique and on the choice of parameters, such as smoothing and convergence parameters. This points to the conclusion that while a qualitative analysis of the analyzed fields may be possible in most cases, a direct quantitative analysis between fields analyzed using different analysis techniques or different analysis parameters may be misleading. Thus, we recommend the use of the same smoothing and convergence parameters for all analyzed fields when the quantitative comparison between different storms is performed.
Vita

Mario Majcen

Mario Majcen was born in Zagreb, Croatia. Mario developed a strong interest in science and weather at an early age and this interest followed him throughout his education and continues to this day. He graduated from 15th Gimnasium High School in Zagreb, Croatia in 1994. He then enrolled at the University of Zagreb, Croatia, where he earned a Bachelor of Science in Geophysics (Meteorology and Seismology) in 2002. To further his meteorological education, Mario entered the University of Utah graduate program in meteorology in the fall of 2003 to pursue a Master of Science degree under Dr. Jan Paegle. He earned his M.S. in 2005, and the title of his thesis is *SALLJEX forecast experiments with a global variable resolution Euler model*. Mario decided to enroll in the Ph.D. program at Penn State in fall 2005, under the tutelage of Dr. Paul Markowski. At Penn State, Mario was awarded the Anne C. Wilson Graduate Student Research Award in 2005, and the Centennial Research Award in 2006.

Mario has accepted an appointment to be an Assistant Professor in the Department of Earth Sciences at the California University of Pennsylvania at California, Pennsylvania. There, he hopes to continue his research and share his passion for meteorology with his students both inside and outside of the classroom.