The Genesis of Three Nonsupercell Tornadoes Observed with Dual-Doppler Radar

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ABSTRACT

Dual-Doppler radar analyses of three tornadoes associated with a multicellular line of storms are presented. The F2–F3 intensity tornadoes occurred on 15 June 1988 near Denver, Colorado, during the Terminal Doppler Weather Radar (TDWR) Project. These tornadoes developed from mesocyclones of no larger than 2 km in diameter that formed along the collision of two surface outflows. The mesocyclones were observed to build in height and intensify with time, coincident with rapid storm growth overhead. All three mesocyclones were clearly associated with the maximum storm updrafts. Downdrafts and associated outflows did not play a role in the formation of one of the tornadoes, but may have contributed to the genesis of the other two tornadoes. It is clear that a downdraft is not a necessary condition for the formation of a nonsupercell tornado, but when present, likely plays a role in determining the timing and intensity of the tornado. This is achieved by the downdraft and outflow causing an increase in the magnitude of the low-level convergence and updraft.

Vertical vorticity production terms were examined for each tornado. Given the close proximity in time and space of the tornadoes, there was surprising variability in the magnitudes and locations of the stretching, tilting, and advection terms for each tornado. In general, however, the predominant contribution to positive vertical vorticity and tornadogenesis was from vorticity stretching in the 0.2–2.0-km layer resulting from intensification of low-level convergence and storm updrafts. Above 2.0 km, increased vertical vorticity resulted from a redistribution of low-level vorticity vertically. Small areas of positive vorticity tilting were found within the regions of large streamwise vorticity just prior to tornadogenesis but not during the formative stages of the mesocyclones, amplifying the already strong contributions to tornadogenesis from vertical stretching of the vortices.

The spatial resolution of the data presented here is as high as any documented in tornado literature. However, limitations in what features are actually resolvable become strikingly apparent and are discussed in the paper.

1. Introduction

On 15 June 1988, four tornadoes occurred in and near Denver, Colorado. The tornadoes were unusually intense for the area causing F2 and briefly, F3 damage (Fujita 1981). Fortuitously, the entire evolution of three of these tornadoes occurred well within the dual-Doppler lobes of two radars spaced 20.5 km apart; allowing for unprecedented collection of high, spatially, and temporally resolved, dual-Doppler data. Wilson (1986), Brady and Szoke (1989), Wakimoto and Wilson (1989), and Wakimoto and Martner (1992) have studied tornadoes common to this region using single Doppler radar and have described them as typically short-lived (<17 min) and weak (F0 and F1). They were associated with nonsupercell storms and occurred while the storm was in its rapid growth stage. These authors have presented convincing evidence from single Doppler data alone, that the vertical vorticity originated near the surface along boundary layer convergence lines. This vorticity then intensified and built upward in concert with the rapidly developing storm. The purpose of this study is to show that the more intense tornadoes on 15 June were also nonsupercell tornadoes and to provide high-resolution details on the evolution of the vorticity and vertical velocity associated with the events.

First, it is desirable to define the term nonsupercell tornado. Wakimoto and Wilson (1989) in their paper on nonsupercell tornadoes simply refer to them as those tornadoes not occurring with supercell storms, that is, occurring with storms that do not contain mesocyclones. This has lead to some confusion since a snapshot of a storm defined as containing a nonsupercell tornado may actually contain a mesocyclone near tornado time that is only the result of a near-surface rotation that has built upward in the classic nonsupercell generation process. In addition, there has been confusion regarding the name of tornadoes developing along a gust front or with a feeder cloud that is associated with a supercell but the actual parent cloud of the tornado does not contain a mesocyclone.

We want to discriminate between tornadoes occurring with supercells that contain well-established
midlevel mesocyclones and those tornadoes occurring with a parent cloud in its growth stage and whose vorticity originates in the boundary layer; this is similar to the Burgess and Donaldson (1979) division. Thus, we define a nonsupercell tornado as a tornado occurring with a parent cloud in its growth stage that does not contain a preexisting midlevel mesocyclone. A midlevel mesocyclone is defined as a rotation above cloud base. While we do not impose a specific vorticity value to be classified as a mesocyclone, project scientists during the Joint Doppler Operational Project (JDOP; NSSL 1979) used an azimuthal shear value of $5 \times 10^{-3}$ s$^{-1}$ as measured by Doppler radar. Thus by the above definition tornadoes occurring with feeder clouds associated with supercell storms that themselves do not contain mesocyclones would be classified as nonsupercell tornadoes. This is the situation with the Newcastle tornado studied by Wakimoto et al. (1995). Thus, nonsupercell tornadoes may occur in the same environment that is producing supercell storms. Typically, a supercell tornado will occur with a storm that is in its mature stage and has a well-established midlevel mesocyclone long before the tornado. A nonsupercell tornado typically occurs with a storm in its rapid growth stage where the vertical vorticity clearly originates near the ground.

Photographs of the three tornadoes examined in this study are shown in Fig. 1. The tornado in Fig. 1a traversed the open country of the Rocky Mountain Arsenal (see Fig. 2) and was estimated to be F1 intensity in the Storm Data report. The Denver National Weather Service (DEN) classified the tornado in Fig. 1b as causing F2 damage. This tornado uprooted hundreds of city-owned trees, with some damage to buildings and cars. The tornado in Fig. 1c was briefly F3 on the scale but mostly was of F2 intensity. Eighty-five buildings were damaged; several cars and buildings were severely damaged, with a total loss estimated at $5-10$ million. Seven people suffered minor injuries. The damage paths are shown in Fig. 2.

Herein, the evolution of the boundary layer wind fields that play a key role in storm initiation, surface vortex development, and tornadogenesis are documented. Characteristics of the atmospheric environment and storm structure are discussed. Nonsupercell tornadogenesis is addressed by examination of the vertical motions, horizontal divergence, vorticity, and vertical vorticity production terms in proximity of the surface mesocyclones.

2. Review of recent literature on tornadogenesis
   a. Supercell tornadoes

Strong, devastating tornadoes producing greater than F1 damage are frequently the product of severe, long-lived thunderstorms that develop intense updraft circulations, that is, mesocyclones. These thunderstorms, often called supercell storms, have been found to form in environments characterized by winds that increase strongly with height and exhibit either strong, low-level veering of the horizontal wind with height (Browning 1964; Klemp and Wilhelmson 1978) or strong, unidirectional wind shear (Rotunno and Klemp 1985). Some of the first single and dual-Doppler radar observations of tornado-producing storms showed that the nascent tornado appeared following the descent and intensification of the mesocyclone to the surface (Brandes 1977; Burgess and Donaldson 1979). Cloud model simulations by Klemp and Rotunno (1983) indicated that the role of the mesocyclone was not in creating the tornado, but in creating a favorable environment for tornadogenesis in terms of locations of rain-cooled downdrafts and pressure gradients within the storm. They found that the surface rotation arose from solenoidal generation of horizontal vorticity along the forward flank, rain-cooled outflow that was tilted vertically as it came in contact with horizontal gradients in upward vertical velocity associated with the supercell storm (Klemp and Rotunno 1983; Rotunno and Klemp 1985). These results were in general agreement with multiple Doppler radar observations by Ray et al. (1981) and Brandes (1984) that showed tornadogenesis occurred during the stage when an intense downdraft along the northern side of the updraft was observed to descend and wrap around the western side of the main updraft. The low-level circulation was observed to intensify as the leading edge of the outflow wrapped around the updraft, creating an occlusion of cold outflow with warm inflow air. A schematic figure in Lemon and Doswell (1979) highlights several of these features nicely. However, Brandes (1984) found that while tilting of horizontal vorticity was important during the pretornado stage, the mechanism for tornadogenesis was more closely tied to vorticity amplification by increased convergence that resulted from rainy downdraft–updraft interaction.

By increasing the horizontal resolution of their 3D cloud model using a nested mesh model, Klemp and Rotunno (1983) were able to examine in more detail the occlusion process leading up to tornadogenesis. They found that as the low-level rotation intensified, the gust front began to occlude and a strong, smaller-scale downdraft formed directly behind the gust front. They proposed that this (occlusion) downdraft and the occlusion process were dynamically induced by low pressure associated with the intense low-level rotation. The existence of an occlusion downdraft was confirmed in the study of Carbone (1983). Using an adaptive grid

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1 Fujita (1981) defined a rotation with a diameter between 4 and 400 km as a mesocyclone and one with a diameter between 0.4 and 4 km as a mesocyclone. We do not make a distinction in size; thus, our definition also applies to misocyclones.
cloud model, Wicker and Williamson (1993) performed numerical simulations of tornadogenesis with twice the resolution of the Klemp and Rotunno's nested mesh. They found that mesocyclone intensification caused a low pressure region and strong upward pressure gradients to form below cloud base. The vertical pressure gradients accelerated the flow upward and led to a rapid increase in the low-level convergence between the surface and the cloud base, thereby stretching the vertical vorticity and contracting the subcloud circulation into a tornado. Ward (1972), through use of a laboratory model to explore various tornado features, found that when the updraft diameter was large compared with the depth of the inflow, a surface pressure profile characteristic of tornadoes was produced. It would appear that magnitude of the updraft and pressure gradient dictates how much the surface circulation is stretched.

Fig. 1. Photographs of the three 15 June 1988 tornadoes studied. (a) Tornado associated with the T1 rotation (see text) located over the Rocky Mountain Arsenal (photo courtesy of Ed Szeke). (b) Tornado associated with the T2 rotation located 2 miles southwest of Stapleton International Airport (see Fig. 2; photo courtesy of Richard Fillhart). (c) Tornado associated with the T3 rotation located in southwest Denver (photo courtesy of Ginger Hein).
b. Nonsupercell tornadoes

Statistical studies on Midwest storms during spring 1977 and spring 1978 showed that tornadoes occurred from 45% of all mesocyclones; and a few tornadoes were not preceded by identifiable mesocyclones (Burgess and Donaldson 1979). Bates (1963) was one of the first to document observations of a tornado that did not form in a rotating thunderstorm. This tornado formed underneath a line of cumulus congestus clouds located upwind of a severe thunderstorm and at the rear edge of an active squall line. No precipitation was observed in the vicinity of the tornado, nor was the cloud base observed to rotate. Burgess and Donaldson (1979) made similar observations of nonrotating, tornado parent clouds using Doppler radar. They found that the tornadoes were typically short in duration, occurred during the developmental stages of storm systems, and on one occasion, a dust whirl was apparent prior to a condensation funnel. Wakimoto and Wilson (1989) examined 27 Colorado, nonsupercell tornadoes and found that the tornadoes formed while the multicellular storms were in their rapid growth stage and typically produced F0–F1 damage. The environments of these storms exhibited weaker vertical shear of the horizontal wind compared to supercell storm environments. No mesocyclones were evident in the tornado-producing storms.

The presence of the surface convergent boundary, either topographically induced or a result of gravity current propagations, appears to be a necessary ingredient for nonsupercell tornadoogenesis. Over mainland Florida, the collision of two outflow boundaries resulted in rapid development of a cloud line that produced a tornado during the cloud updraft stage (Holle and Maier 1980). Golden (1974) found that pulsating cumulus-scale updrafts coupled with adjacent, subcloud shower outflow from a cumulus cloud line are perhaps most crucial to waterspout formation. The updrafts and shower outflow interact together with environmental winds to concentrate vorticity, which in turn can lead to the formation of both the spiral-scale flow surrounding the waterspout and the waterspout itself. In other cases, Golden (1974) discusses that the shear across shower-induced, wind-shift lines may be the vorticity source for waterspouts. The intersection of two large-scale wind shear lines was believed to be of primary importance in the formation of five Colorado tornadoes (Wilson 1986). In a triple Doppler study of a severe, cold frontal rainband, Carbone (1983) showed that strong horizontal vorticity was tilted upward through the interaction of a cold frontal gravity current with a low-level jet ahead of it. By tornado time, this positive vertical vorticity was redistributed, with maximum vorticity resulting due to stretching and convergence along the frontal boundary.

Equally important are the shear instabilities that have been observed to set up along these surface convergence boundaries (Carbone 1983; Wilson 1986; Brady and Szoke 1989; Wakimoto and Wilson 1989; Wilson et al. 1992). In the severe cold frontal rainband, Carbone (1983) observed inflectional instabilities of the Helmholtz type that evolved into two tornadoes in the absence of any mesoscale circulation feature aloft. In the single-Doppler radar studies by Brady and Szoke (1989) and Wakimoto and Wilson (1989), their data showed that nonsupercell tornadoes were initiated when preexisting mesocyclones in the boundary layer became collocated with the maximum reflectivity core (and inferred updraft) of a rapidly developing storm. Wakimoto and Wilson (1989) further proposed that vorticity stretching of the surface mesocyclone by the storm updraft was the primary mechanism for generating the tornado. This is in contrast to dual-Doppler observations by Wilczak et al. (1992), of an F1–F2 intensity Colorado tornado. They found that although stretching of preexisting vorticity occurred, the strengthening of the mesocyclone through the tornado formation stage was dependent on the tilting process. While the role of surface shearing instabilities in tornadogenesis has mostly been discussed in studies of storms not classified as supercells, this does not preclude their potential importance in supercell storms. Indeed, an examination by Brandes (1977) of an Oklahoma City tornado led him to hypothesize that shearing instabilities and perturbation growth along a surface convergent zone located below a strong parent mesocyclone could have been a likely source for tornadogenesis.

3. Data collection and wind synthesis

The radar data utilized in this study were collected during the 1988 Terminal Doppler Weather Radar (TDWR) experiment (Turnbull et al. 1989) near Denver, Colorado, using the Lincoln Laboratory’s FL-2 (10-cm wavelength) and the University of North Dakota’s UND (5-cm wavelength) radars. The radar locations are shown in Fig. 2 along with the tracks of radar-detected rotations (T1, T2, T3) associated with the developing tornadoes. These radars continuously scanned fixed 120° azimuth sectors centered over Denver’s Stapleton International Airport (STP). The 20.5-km baseline between the radars and the close proximity of the radars to the airport was designed to ensure high-resolution sampling of low-altitude divergence associated with microbursts. Both radars transmitted with 1° half-power beamwidths. Spacing of the data was approximately 0.15–0.5 km in the horizontal and 0.8–1.2 km in the vertical. Volume updates were 2.5 min for each radar and synchronized. The slower scanning rates of the UND radar limited coverage of storm winds and reflectivity to a depth of about 8–9 km in the region of interest. Hence, the radars were able to collect high-resolution dual-Doppler data on the tornadoes from approximately 9 km AGL to within 0.2 km of the ground.
DENVER TORNADOES
June 15, 1988

Fig. 2. Map of facility locations and tracks of the near-surface vortices. Radar and mesonet locations are indicated. The solid black lines represent the radar-derived tracks of vortices T1, T2, and T3. Magnified plots of the lightly shaded boxed regions illustrate the near-surface T2 and T3 vortex locations (dots) at consecutive volume times. Start and stop times refer to the initial and final radar volume times when near-surface vortices were detected. Tornado damage is indicated in the stippled regions (darker stippling represents regions of heavier damage). Damage paths were determined from insurance claims filed due to the storms. Start time of damage was determined from the radar volume time and rotation that coincided with start of the damage path.
Over 50 min of dual-Doppler data were collected on the complete evolution of the tornadoes. Fourteen volumes of dual-Doppler radar data have been analyzed, providing a means of addressing tornadogenesis in more detail than has previously been possible.

Editing of Doppler data was done in a radar-centered coordinate system using a color image display and interactive software (Oye and Carbone 1981) and discussed in detail by Wilson et al. (1984). The radar data were interpolated to a rectangular Cartesian grid of 0.4 km in the horizontal and at two different spacings in the vertical (0.5 and 1.0 km) using a Cressman weighting technique and an elliptical sphere of influence. The wind field and the horizontal and vertical components of vorticity were obtained using the CEDRIC (custom editing and display of reduced information in Cartesian space) software package developed at NCAR (Mohr et al. 1986). The vertical velocity field w was iteratively computed by upward integration of the anelastic form of the continuity equation. Accuracy of the dual-Doppler-derived wind field is addressed in the appendix.

Additional surface wind, temperature, and moisture information were provided by the NOAA Forecast Systems Laboratory (FSL) fixed surface mesonet stations and Lincoln Laboratory’s FAA—Lincoln Laboratory Operational Weather Studies (FLOWS) mesonet (Woelfson 1987) deployed for the TDWR project. Station locations are shown in Fig. 2.

4. Evolution of the storm environment

a. Synoptic scale

A composite map (Fig. 3) of the synoptic environment on 15 June 1988 at 0600 MDT shows most of the western United States was under the influence of an upper-level ridge. A weak, 50-kPa short-wave disturbance was located over central Colorado, while at the surface, a high pressure system moved into Colorado from the northeast following the passage of a surface cold front on 13 June. East-southeasterly surface winds associated with the anticyclonic flow brought higher than normal dewpoint values (>50°F) into northeastern Colorado. Dewpoint temperatures and 5–12 m s⁻¹ southeasterly winds remained high throughout the afternoon. Doswell (1980), in his examination of severe weather outbreaks over the High Plains, has shown that the conjunction of these synoptic-scale features (discussed above) and the return of dewpoints 45°F or larger in eastern Colorado following the passage of the polar cold front, often marked the beginning of severe weather potential east of the Rocky Mountains.

Figure 4 shows the DEN sounding from 0600 MDT 15 June. The atmosphere was conditionally unstable with a lifted index of −5.5 and convective available potential energy (CAPE) of 1154 m² s⁻². This CAPE is at the high end for environments associated with nonsupercell tornado-producing storms (Wakimoto and Wilson 1989), but lower than that observed in many supercell storm environments. A convective temperature of 25.6°C was needed to overcome the stable layer between 64 and 68 kPa. The winds veered in the lowest few kilometers of the atmosphere at 0600 MDT but lacked the strong increase in wind speed with height characteristic of severe Midwest storm environments. The bulk Richardson number (BRN) calculated from the 0600 MDT sounding over the lowest 6 km resulted in a value of 384. Weisman and Klemp (1982) have shown that storms that develop in environments having values of BRN > 100 may exhibit characteristics associated with multicellular storms. Clearly, there was sufficient potentially buoyant energy available to form vigorous thunderstorms even in the absence of strong vertical shear of the horizontal wind. All that was needed was sufficient heating of the boundary layer air and/or some trigger to unleash the instability of the atmosphere.
occurred that contained mesocyclones, produced damaging hail, were long-lived, and exhibited other supercell-type features. For a more detailed discussion on these latter storms, the reader is referred to the Szoke and Rotunno (1993) paper.

\[ b. \text{Mesoscale} \]

The evolution of events on this day was rather complex and more detailed discussions are provided by Purdom and Weaver (1990) using satellite imagery and by Szoke and Rotunno (1993), who compiled information from several data sources. A brief overview of the sequence of events is presented here. Figure 5 shows a 3-h evolution of the near-surface winds around Denver as observed with the PROFS mesonet. The southeasterly flow, dominant throughout eastern Colorado, is reflected by the DEN wind in Fig. 5a. North-northwesterly flow between Denver and the foothills to the west, persisted throughout the afternoon. This wind regime and associated convergence line is a frequently occurring phenomena when the synoptic-scale flow is southeasterly. It is topographically induced and called the Denver convergence and vorticity zone (DCVZ) (Szoke et al. 1984; Wilczak and Glendening 1988; Crook et al., 1989). The DCVZ set up during the morning on 15 June and persisted into the afternoon, as indicated in Fig. 5a by the convergence of wind in the PROFS mesonet and the northeast—southwest line of radially convergent flow (solid contour, labeled DCVZ, marks this boundary) detected by the radar.

A line of cumulus clouds developed above the DCVZ, as observed on satellite imagery (Purdom and Weaver 1990), at about 1300 MDT,\(^3\) but the clouds dissipated following the passage of a cold outflow from a storm off the foothills to the west at 1330. It is conceivable, as postulated by Scorcer and Ludlum (1953) among others, that a local area of more moist and unstable air may have remained following cloud decay. By 1500, Fig. 5b, the DCVZ was dissipating, as indicated by the dashed line. Two outflow boundaries, B1 and B2, emerged from storms to the northwest and southeast of Denver. The storm that produced B1 had a mesocyclone as early as 1345; the storm that was southeast of Denver was also long-lived, contained a midlevel mesocyclone, and produced golfball-size hail (Szoke and Rotunno 1993). A time series of temperatures recorded by the PROFS mesonet stations showed 2°C temperature drops over a 20-min period following the passage of both the B1 and B2 outflows, with the warmest air located between B1 and B2 (see Figs. 5bc). A mesonet moisture analysis by Szoke and Rotunno (1993) using the PROFS and FLOWS data indicated that the relative humidity behind B2 was at least

\[ \text{\footnotesize Note: FIG. 4. Skew } T-\text{log} \text{ plot of DEN rawinsonde data at 0600 and 1800 MDT 15 June 1988. Black solid lines are the temperature and dewpoint curves for 0600 MDT; gray, dashed lines are for 1800 MDT. The convective condensation level (CCL) for the 0600 sounding is marked. Below the CCL, the dry adiabat is dashed and the mixing ratio is dotted. The moist-adiabatic parcel ascent is denoted by the light, solid curve above the CCL. The curve encloses a positive area or CAPE of 1154 m}^2 \text{s}^{-2} \text{ assuming a convective temperature of } 25.5°C \text{ is reached. The PROFS mesonet station at DEN reached convective temperature by 1450. Insert figure shows the hodograph for the 0600 wind profile located to the right of the figure. Speed rings are in 5 m s}^{-1} \text{ increments and the heavy bold vector is the mean storm motion for the line of storms. Height levels are marked in km and correspond to the height levels shown on the left-hand side of the skew-}T \text{ plot.} \]
20% higher than the humidity associated with B1. The boundaries collided at 1550 (see Fig. 5c), 10 km northwest of DEN and STP and 25 km northwest of the FL2 radar. This collision occurred 15–30 min prior to the onset of the three tornadoes, in the vicinity of the weakened DCVZ. Convective temperature had already been reached prior to collision, but calculations indicate a surface air parcel would need to ascend to a height of 2.2 km to reach its level of free convection (LFC). With the increased convergence at the surface arising from the impending collision, air was forced upward over these boundaries. Computations using the 0600 sounding indicated that air at a height of approximately 1.5 km (the height of the gust fronts) would need to ascend an additional 0.7 km in height to reach the LFC; this extra lift would be provided by the actual collision of the two outflows. We believe it was because of this collision of two outflows and resultant strong updrafts in the vicinity of the DCVZ, coupled with the relatively large instability of the atmosphere, that the 15 June tornadoes were stronger in intensity (F2–F3) than previously documented nonsupercell tornadoes.

The first radar-detectable echoes appeared at 1540, prior to collision, near a height of 6 km AGL. Their location at 1545 is shown in Fig. 5c. Near-surface vortices, detected by Doppler radar, were located along the
southern boundary, B2, no more than 6 min prior to collision. Onset and dissipation of three of these vortices are listed in Fig. 2 and their tracks are shown in Fig. 5c. It is interesting to note the longevity of the near-surface vortices prior to tornado occurrence (see Fig. 2). Storms were already rapidly growing above the T1 and T2 vortices, but no radar-detectable echo was present above T3 at collision time. By 1615, a multicellular line of 50–60-dBZ, storms existed above the collision zone, and is in evidence in Fig. 5d.

c. Local scale

Several of the vortices (1–2 km in horizontal dimension) located along the convergence zone and along the northwest side of the associated updrafts can be seen in the dual-Doppler-derived wind field at 1600 (Fig. 6), along with regions of cyclonic shear. These vortices had a mean wavelength of 4–5 km and are similar in scale to those observed by Mueller and Carbon (1987). The linear orientation of these vortices is reminiscent of observations of dust devils sited in a literature review by Barcilon and Drazin (1972). In their attempt to ascertain dust devil formation, they examined the effect of Helmholtz and Rayleigh–Taylor instabilities in destabilizing the flow. They proposed that the coupling of these instabilities in an unstable environment can lead to a row of strong vertical vortices that could become stretched by convection and intensify to form dust devils. The 15 June data presented here lacks sufficient resolution to determine if Rayleigh–Taylor instabilities were present. It seems likely that the vortices observed in Fig. 6 initially formed due to Helmholtz-type shearing instabilities.

Profiles of vertical vorticity magnitudes for T2 and T3, shown later in the paper, illustrate that the shearing instabilities located along B2 were essentially confined to the lowest 1 km of the boundary layer and were maintained at a constant magnitude during the 6 min prior to the B1–B2 collision. Vertical development and intensification of these shearing instabilities occurred following the collision of the two gravity currents; the

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**Fig. 6.** Dual-Doppler-derived horizontal, storm-relative winds (vectors), and vertical vorticity (contours are $5 \times 10^{-3}$ and $15 \times 10^{-3}$ s$^{-1}$) at 1600 MDT and a height of 0.2 km. The vertical velocity field at 2.2 km is overlaid. Upward (downward) vertical velocity regions are shaded (hatched) in increments of 4 m s$^{-1}$. The location of circulations T1, T2a, T2b, and T3 are shown. The scale vector is in the lower right-hand corner.
collision causing an increase both in the convergence of flow in the vicinity of these instabilities and in the updrafts feeding the storms, thereby enhancing the intensification of the shearing instabilities due to buoyancy effects. Mesonet data showed cool temperatures remained along the convergence line following collision, while temperatures north of the collision warmed by 1°–2°C and temperatures south of the collision stayed 1°–2°C colder.

In a subsequent exploration of vortex development in dust devils and other columnar vortices, Maxworthy (1973) speculated the source of vertical vorticity in the horizontal vortex lines encountered a horizontal gradient of vertical velocity, that is, rising hot thermals over heated, flat terrain and subsequent local, low-level convergence. A more common manifestation of ambient horizontal vorticity in the boundary layer is in the form of horizontal roll vortices that arise from vertical wind shear and surface heating and tend to be aligned with the mean wind (LeMone 1973). Kropfli and Kohn (1978) and Christian and Wakimoto (1989), among others, have shown that Doppler radar can be used to detect horizontal rolls. In a study by Wilson et al. (1992) on the role of boundary layer convergence zones and horizontal rolls in thunderstorm initiation, surface misocyclones developed preferentially near the intersection points of the DCVZ and the horizontal rolls. Three of the misocyclones evolved into nonsupercell tornadoes. Examination of the CP-2 and FL-2 radars in the vicinity of the DCVZ, during the early afternoon hours failed to show the presence of thermally driven, horizontal rolls. Rolls could have been present and gone undetected but we have no evidence of this.

The first contact of B1 outflow with B2 occurred near the T1 vortex. The magnitude of the vertical vorticity increased dramatically within this vortex to a dual-Doppler-derived value of $20 \times 10^{-3}$ s$^{-1}$. This rotation strengthens to tornado intensity within the next 4 min. Vortices T2a and T2b were observed to merge over a 5-min period, first aloft and finally at the surface by 1612, to become the single, distinct vortex associated with the T2 tornado. Note, the T2 tornado, which began at 1622, had the largest condensation funnel. The transient circulations located between T1 and T2, in Fig. 6, lacked vertical continuity and storm development overhead was not as vigorous. These vortices dissipated soon after 1600. Wilson et al. (1992) have shown that the surface vortices that underly the high reflectivity cores in the multicellular storms are the vortices that preferentially evolve into tornadoes. This is also consistent with the authors daily observations, for three summers, of high-resolution Doppler radar data in eastern Colorado.

5. Storm structure and evolution

To illustrate the general relationship between the precipitation, horizontal wind, vertical velocity, and vorticity fields within the ensuing storms, 4 of the 14 analyzed dual-Doppler volumes are presented here. A sequence of 0.2 and 4.2 km AGL plots are shown in Figs. 7 and 8 for the four time periods spaced 10 min apart showing the large-scale structure. Storm relative winds are plotted using a mean storm motion vector, shown relative to the environmental hodograph profile in Fig. 4. Storm motion was from the northwest at 305°, 4.5 m s$^{-1}$. In Figs. 7 and 8, wind vectors at every other grid point are plotted.

Similar to Fig. 6, the convergence line and associated boundary layer vortices are apparent in Figs. 7a and 7c. At the 0.2-km height there is a transition from many weak vortices (Figs. 7a and 7c) to two strong vortices associated with T2 and T3 (Figs. 8a and 8c). Initial maximum vorticity values at 0.2-km height are $10^{-15} \times 10^{-3}$ s$^{-1}$ (Fig. 7a) that increase to $30 \times 10^{-3}$ s$^{-1}$ (Fig. 8c). The vortices do not reach the 4.2-km level until a storm cell is observed to grow above them, similar to what Wakimoto and Wilson (1989) observed for nonsupercell tornadoes.

The first major storm developed in the vicinity of the T1 rotation and the B1 and B2 collision point. Ensuing storm development above the convergence line evolved in a manner similar to that described by Chisholm and Renick (1972) in their discussion of multicellular evolution in hailstorms. As can be seen in Figs. 7 and 8, as the first storm (near T1) reaches maturity, new cell development occurs along its right (south) flank above T2. As the mature (T1) storm enters the dissipation stage, the (T2) daughter cell grows rapidly over the next 10 min toward maturity, and a new storm cell can be seen to develop near T3 (Fig. 8b). (Note that the y-axis grid is shifted to the south in Fig. 8.) Maximum reflectivities within the storms were greater than 55 dBZ, and storm tops extended up to the tropopause height (10.4 km AGL in Fig. 4).

A 2–3-km-wide region of 4 m s$^{-1}$ and greater updrafts at 2.2 km AGL resulting from the collision are located parallel to and immediately southeast of the convergence line (see Fig. 6). Beginning at 1604 this linear region of updraft evolved into more cellular regions associated with each storm. Maximum updraft magnitudes associated with the developing storms ranged from 10 to 20 m s$^{-1}$.

a. 1554–1557 MDT

The maximum reflectivity core within the (T1) storm was located about 4 km southeast of the surface convergence zone, as can be seen in Fig. 7 at 1554. At this time, the T1 vortex was the strongest misocyclone along the convergence line and has built upward from the surface with time (as will be shown later) to a depth
Fig. 7. Dual-Doppler-derived, horizontal storm-relative winds, reflectivity (color filled in dBZ, increments) and vertical vorticity (mauve-colored contours) in the vicinity of the T1, T2a,b, and T3 mesocyclones at 0.2 and 4.2 km AGL. Vertical vorticity contours are in increments of $10 \times 10^{-3} \text{ s}^{-1}$ starting at $5 \times 10^{-3} \text{ s}^{-1}$. The vortices corresponding to the four mesocyclones are annotated. Scale vector is in the lower right-hand corner: (a) 1554–1557 at 0.2 km, (b) 1554–1557 at 4.2 km, (c) 1604–1607 at 0.2 km, (d) 1604–1607 MDT at 4.2 km. The location of the vertical cross section (AB) displayed in Fig. 9 is shown in panel (b).
of 6.2 km. In Fig. 7b, the center of the vortex is located within a weak-echo region (WER), at \((x = -11.6, y = 18.0)\). No mesocyclone, as defined in section 1, was detected within the body of the storm prior to the tornado genesis process. Below 4.2 km, the vortex is aligned vertically. Above 4.2 km, the axis of the vortex tilts slightly to the southeast, toward the main precipitation core within the storm (\(>45 \text{ dBZ}\), in Fig. 7b).

A vertical northwest–southeast cross section taken through T1 and the WER, Fig. 9, shows a pronounced echo weak vault with maximum upward vertical velocities and maximum vertical vorticity within the vault region. This vault persisted over the next eight minutes, almost to tornado time (1604). Its location on the northwest side of the storm within the main updraft core can be viewed as almost an inverted image of updraft and vault locations typically observed along the right forward flank in supercell storms (see Fig. 5 in Browning 1964). Given the southward-pointing, vertical wind shear that was evident below 5 km in the 1800 sounding (Fig. 4), it should perhaps be of no surprise that the location of the updraft and rotation centers is along the northwest flank of the storm. Modeling studies by Weisman and Klemp (1984, 1986)
have shown that for weak to moderately sheared environments, an eastward-pointing, ambient wind shear would tend to produce a cyclonically rotating updraft along the right flank of the storm where low-level outflow is most opposed to the storm relative inflow. Here, with a southward-pointing wind shear, the west-northwest side of the storm located along the region of strongest surface convergence corresponds to the right flank in the Weisman and Klemp modeled storms. Analysis of air particle trajectories (not presented) showed the low-level southerly flow, upon impact with the convergence line, became caught up in the broad updraft region and then curved back toward the main storm development south of the line. Low-level flow from the north-northeast generally fed directly into the circulation around the mesocyclones and rose cyclonically in the updraft region into the storms overhead. Midlevel flow from the north-northwest, evident in Fig. 9, also fed storm development.

It should be noted, though, that for the moderately sheared environment (their $U_s = 30 \text{ m s}^{-1}$) in Weisman and Klemp (1984), resulting storms were multicellular with a supercell at the southern end of the line. Right flank updrafts were in association with the supercell storm. In our case, the T1 storm is part of multicellular line of storms. And while this storm exhibited many structural features observed with supercell storms, a mesocyclone was not produced above cloud base from the tilting of the ambient horizontal vorticity produced by the vertical wind shear; this in light of the fact that other supercell storms containing mesocyclones developed in the vicinity of these tornadic storms. Thus, while environmental wind shear conditions obviously played a role in storm development and resulting storm structure, other factors also determine storm type.

In Fig. 7a, the misocyclones T2a, T2b, and T3 have distinct closed vortices at 0.2 km. At this time, the misocyclones are shallow features confined to the lowest few kilometers of the boundary layer. No closed circulations are observed at 4.2 km, Fig. 7b. Reflectivities of at least 30 dBZ, associated with cumulus congestus development are now detected in the vicinity of the T2 misocyclones. No storm echo is found above the T3 rotation.

b. 1604–1607 MDT

Ten minutes later, at 1604, tornado T1 is observed over Rocky Mountain Arsenal (see Fig. 2) and is associated with a 55-dBZ storm. As a consequence of the T1 storm propagation to the southeast toward the UND radar, it was not possible to obtain dual-Doppler winds within the T1 rotation and storm above 3.2 km (Fig. 7d). In the intervening time period prior to 1604, the WER observed at 4.2 km evolved into a bounded weak-echo region (WER) that was quite distinct at levels 1.2–3.2 km. Updrafts persisted within the WER.

Vortex strength has increased only slightly within the T2a, T2b, and T3 misocyclones by this time (Fig. 7c). Rapid storm development is transpiring near T2 (Fig. 7d). The T2a vortex now extends up to 4.2 km and is aligned vertically at this time. Above 4.2 km, the updraft is observed to tilt southeast toward the 45-dBZ core. No significant storm development has occurred yet near T3.

c. 1614–1617 MDT

By 1614, precipitation from the T1 storm can still be seen in Fig. 8a centered at $(x = -6.0, y = 12.0)$ but the T1 tornado has already dissipated. Video footage of the T1 tornado indicated that while a dust column still extended upward from the surface, no funnel was obvious at or below cloud base at 1609, even though a funnel was visible earlier.

The double vortex (T2a and T2b) at 1604 has merged and is now labeled T2. It is clear that T2 and T3 have intensified dramatically (Fig. 8a) over the last 10 min and are much stronger than the T1 vortex during its tornado stage. A small hooklike appendage, typically observed with supercell tornadoes, is apparent as precipitation wraps around the T2 vortex. At this time, the T3 rotation is quite visible above 4.2 km. As with the T1 storm, the T2 and T3 rotation centers are located along the NW or right flanks of the storms. Rotations have been observed to build upward from the surface with time. No midlevel mesocyclones generated by tilting of ambient horizontal vorticity were present in the
storms. Storm development above the T3 misocyclone is not as advanced as the 45–50-dBZ, T2 storm nor does it show any supercell-like features during its lifetime, in contrast to the T1 and T2 storms.

d. 1624–1627 MDT

At approximately 1617, visual sightings were made of funnels pendant from the cloud bases of the T2 and T3 storms. Since most of the horizon was obstructed to storm observers by buildings and other city structures, insurance claims were used to determine locations and times of surface damage. The start of lighter surface damage (see Fig. 2) associated with the T2 storm is coincident with the first funnel observations. Damage of this nature was generally confined to broken windows, damaged roofs, and broken tree limbs. Lighter damage caused by T3 began a few minutes later at 1622. The most intense damage to property caused by the T2 and T3 tornadoes began at 1624.

The T2 and T3 storm intensities at this time are at least 55 dBZ (see Figs. 8c and 8d) and the vertical vorticity within the tornadoes was strongest below 2.2 km. The T2 vortex tilted 3 km to the southeast with height toward the maximum reflectivity core. This tilt is obvious in the visual observations of the funnel (Fig. 1b). The hooklike appendage observed earlier with the T2 rotation is still apparent in Fig. 8c, but the wrap-around of precipitation is not quite as dramatic. Aloft, in Fig. 8d, the vortex is situated along the gradient of maximum reflectivity. The T3 vortex extends vertically to 4.2 km and was located within the maximum reflectivity core. Visually at this time, the T3 vortex near cloud base was quite dark and surrounded by precipitation (see Fig. 1c). The cloud base itself was quite flat and uniform; no wall cloud was apparent.

6. Tornado parent circulations

Wakimoto and Martner (1992) and Wakimoto and Wilson (1989) reported that nonsupercell tornado widths, based on photogrammetric analysis, vary between 25 and 600 m. The radar range to the tornadoes is approximately 20 km. The half power beamwidth of FL-2 at this range is then about 350 m. Thus, it is apparent even before the synthesis of radar data that basic radar measurements cannot resolve features on the tornado scale. See the discussion later in this section and in the appendix on the scale of features actually resolved in the dual-Doppler analyses. However, the 3D wind field associated with each tornado parent circulation can be examined and are presented here to facilitate a clearer understanding of the horizontal and vertical motions leading to tornadogenesis.

a. T1 vortex

Figure 10 shows a time series at 2.5-min intervals, of 2.2-km plots of the T1 misocyclone. The 20-dBZ contour outlines the main body of the storm and heavy dark contour is the $10 \times 10^{-3}$ s$^{-1}$ vertical vorticity values. The black dot is the location of the radar TVS signature at about 0.10 km AGL. On quick inspection, it is obvious that this misocyclone is completely embedded within the upward moving air (red shading). Magnitudes of $4-10$ m s$^{-1}$ persist at $2.2$ km along the convergence zone until tornado time (Fig. 10c), when the continuous horizontal extent of the updrafts becomes broken and 2.5 min later is confined to the area of the tornado vortex (Fig. 10f). Perusal of the detailed 2.2-km plots indicates a slight strengthening of the updrafts in the vicinity of T1. Prior to tornado (Figs. 10a–d) the vortex lies along a gradient in upward velocity values with the maximum just east and southeast of the rotation center. From the onset of the tornado (Fig. 10e) to its cessation at 1609 (Fig. 10f), the maximum updraft core and the maximum vertical vorticity are collocated. Downdrafts (blue shading) of 4 m s$^{-1}$ intensity are apparent only during the tornado stage (Figs. 10e, f) at 2.2 km but fail to penetrate at this strength to the 1.2-km level. Analysis of air particle trajectories indicated that the source of this downdraft is in the northwesterly, low-$\theta_e$ air impinging on the storms at middle levels (4.2–6.2 km) along the gradient of storm reflectivity, clearly in contrast to the precipitation-loaded downdrafts observed in the Del City and Ft. Cobb supercell storms (Ray et al. 1981).

A time–height profile of maximum vertical vorticity values estimated from single-Doppler data is shown in Fig. 11 for T1. The single Doppler vorticity values were estimated from the maximum azimuthal Doppler velocity shear at each elevation angle from FL-2. The vorticity is estimated by dividing the maximum azimuthal velocity difference in the rotation by the distance between these extremes. Examination of FL-2 and UND showed that both radars detected similar magnitudes of azimuthal shear during the evolution of the three misocyclones. Thus, assuming cylindrical symmetry of the rotation, the vorticity estimate is equal to twice the azimuthal shear. It is evident from Fig. 11 that the misocyclone was confined to the lowest 1–2 km of the boundary layer initially, then intensified and built upward with time, coinciding with the rapid growth of reflectivity within the storm. A maximum estimated vertical vorticity of $200 \times 10^{-3}$ s$^{-1}$ occurs near the surface during the tornado, more than 10 times stronger than the dual-Doppler-derived value. The dual-Doppler vertical vorticity values were much smaller in magnitude due to the smoothing that occurs in processing the dual-Doppler winds (see the appendix) but were of the same order of magnitude as dual-Doppler values reported by Carbone (1983), whose data spacing was very similar to this case.

The collocation of the updrafts with the maximum vorticity center in Fig. 10 and the broad region of upward motion within the vault (Fig. 9) illustrate the apparent importance of strong convergence in tornado-
Fig. 10. Time series of storm-relative winds, reflectivity, vertical velocity and vertical vorticity plots at 2.2 km associated with the T1 misocyclone. Updrafts (downdrafts) are the red (blue) color-filled areas (see legend). The area of vertical vorticity of at least $10 \times 10^{-3} \text{s}^{-1}$ is bounded by the heavy, black contour. The thin, black line is the 20-dBZ reflectivity contour outlining the storm area. The black dot is the location of the radar detected tornado vortex signature (TVS) at the lowest elevation scan, 0.3° (~0.10 km AGL). Scale vector is in the lower right-hand corner: (a) 1554–1557, (b) 1557–1559, (c) 1559–1602, (d) 1602–1604, (e) 1604–1607, and (f) 1607–1609 MDT.
genesis. Downdrafts appeared to play no role in the formation of the T1 tornado, that is, in causing an increase in low-level convergence of flow into T1. The single Doppler radial velocities were examined from both radars for the presence of small-scale divergence indicative of a downdraft that may not have been resolved by the dual-Doppler analysis. This inspection revealed there were none.

b. T2 vortex

Figure 12 illustrates the pretornado (a)–(e) and tornado (f) stages of flow around the T2 vortex. Vortices T2a and T2b, which were initially located along the western edge of the continuous line of updraft, merge by 1612. The process of vortex mergers have been documented in a numerical study by McWilliams (1984). He found that as the turbulent cascade begins around these incipient, local extrema in vorticity, the vortices resist deformations due to neighboring vorticity maxima and grow in circulation by mergers with weaker, like-sign vorticity extrema. Mergers by vortices having unequal strengths were found to be a mechanism for growth of the stronger partner. Here, the stronger vortex, T2a, resides along the gradient of the maximum updraft and downdraft centers during the pretornado stage, reminiscent of the location of these features in the Del City storm (Ray et al. 1981; Brandes 1984) and in Carbone’s cold frontal, tornado case during the tornado stages. Following the merger of the two circulations (T2a,b), a decrease in the horizontal winds and in the vertical motion intensity is first observed (Fig. 12d). By 1619, the motions have dramatically increased and the continuous line of updraft evolves into one large updraft collocated with T2. The updraft speed at 2.2 km near T2 increased from about 4 m s⁻¹ at 1557 (Fig. 12a) to 16 m s⁻¹ at 1624 (Fig. 12f). The increase may be in response to downdraft intensification of the low-level convergence on the southwest side of the vortex. Examination of the dual-Doppler derived divergence field indicated an increase in convergence (minimum values) at 2.2 km from $5 \times 10^{-3}$ to $9 \times 10^{-3}$ s⁻¹ between 1614 and 1619. The timing seems to be particularly significant since the increase in intensity appears immediately prior to the nascent tornado (at 1622) and start of light surface damage. The strength of this downdraft was 8 m s⁻¹ at 2.2 km and still 4 m s⁻¹ at 1.2 km. The downdraft and updraft orientations suggest an occlusion process could be setting up. However, no pressure-driven, occlusion downdraft was detected within the vortex itself during the tornado stage.

The single Doppler vorticity estimates shown in Fig. 13 indicate that the vortex was initially confined to the boundary layer. Upward growth began about 1554 at the time cumulus clouds increased in vertical extent. Between 1605 and 1610, the vortex strengthened considerably in conjunction with rapid intensification of the storm aloft. At 1612, vortices T2a and T2b merge with a corresponding dramatic decrease in vorticity. This decrease was accompanied by an increase in the diameter of the vortex. Within a couple of minutes of merger, the vortex reintensified apparently in response to the increased updraft noted previously between 1614 and 1619. A maximum vorticity value of $300 \times 10^{-3}$ s⁻¹ was observed near the surface during tornado occurrence; again, an order of magnitude higher than that retrievable by dual-Doppler synthesis.

c. T3 vortex

The evolution of the updrafts and downdrafts with T3 is considerably different than with T1 and T2 and erratic. In Fig. 14, from 1557 (Fig. 14a) until 1614 (Fig. 14c), the vortex is embedded within an updraft–downdraft pair very similar to observations of Carbone (1983), Ray et al. (1981), and Brandes (1984) and the modeling simulations of Klemp and Rotunno (1983). It is unclear if this downdraft can be considered an occlusion downdraft since the downdraft is present early on when the rotation is shallow and weak. Furthermore, the updraft weakens and the downdraft disappears just as the tornado begins (Fig. 14d). The downdraft later reappears 5 min into the tornado stage (Figs. 14e,f). It is not obvious whether these inconsistencies are real or are introduced by difficulties in resolving small-scale features.

The time–height evolution of the single Doppler vorticity (Fig. 15a) is very similar to the latter two profiles shown for T1 and T2, illustrating a vortex in the boundary layer undergoing rapid intensification, and increasing in vertical extent as a storm develops swiftly overhead. The maximum single Doppler vortic-
Fig. 12. Same as Fig. 10 but for the T2 mesocyclone: (a) 15:57–15:59, (b) 16:04–16:07, (c) 16:09–16:12, (d) 16:14–16:17, (e) 16:19–16:22, and (f) 16:24–16:27 MDT.
superstorm tornadoes documented in literature. The intensification of the T2 downdraft may have precipitated a corresponding increase in horizontal convergence of flow into the T2 vortex, leading to vortex intensification. The role of the T3 downdraft is unclear and may be tied to unresolved, smaller-scale motions.

7. Vorticity production

The 15 June dual-Doppler data provide an opportunity to evaluate not only the vertical component (ζ) of the relative vorticity (ω), but also the horizontal components (ξ, η) of vorticity, and the terms contributing to the time rate of change of the vertical vorticity (δζ/δt) leading up to tornadogenesis. The relative vorticity ω is computed from the curl of the relative velocity ∇ × V; that is,

\[ \omega = \xi \mathbf{i} + \eta \mathbf{j} + \zeta \mathbf{k}, \]

where

\[ \xi = \frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}, \]
\[ \eta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}, \]

and

\[ \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}. \]

The latter term was discussed in the previous sections. Manipulation of the horizontal momentum equations and neglecting the contributions of the Coriolis parameter and the solenoid term, gives rise to the following relation for the time rate of change of the vertical vorticity used in this study:

\[ \frac{\partial \zeta}{\partial t} = - \left( u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} + w \frac{\partial \zeta}{\partial z} \right) \]
\[ - \zeta \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) - \left( \frac{\partial w}{\partial x} - \frac{\partial w}{\partial y} \right), \]

where the first term in the parentheses on the right-hand side is the horizontal (ζ_h) and vertical advection (ζ_v) of vertical vorticity, followed by the divergence (stretching) (ζ_DTV) and tilting (ζ_TLT) terms. The vertical derivative terms were computed using forward difference calculations. A vertical grid spacing of 0.5 km was used for T1 and T2, and 1.0 km for T3, which was located further out in range. The tilting term was very small near the ground since the vertical velocity values are close to zero. Comments on the relative accuracy of these derived quantities are included in the appendix.

a. Horizontal vorticity

The mechanism for vertical vorticity production at midlevels in supercells is believed to derive from ver-
Fig. 14. Same as Fig. 10 but for the T3 mesocyclone: (a) 1557–1559, (b) 1609–1612, (c) 1614–1617, (d) 1619–1622, (e) 1622–1624, and (f) 1624–1627 MDT.

Vertical tilting of horizontal vorticity in the sheared environment in the presence of vertical motion (Barnes 1970; Schlesinger 1975; Wilhelmson and Klemp 1978; Lilly 1982) and subsequent intensification due to stretching of the vortices by converging inflow air and intensification of the updraft with height (Rotunno
1981; Klemp and Rotunno 1983). Low-level vertical vorticity production is believed to result from different forcing mechanisms. Two such mechanisms are the vertical tilting of baroclinically generated, horizontal vorticity as it comes into contact with a strong updraft (Klemp and Rotunno 1983; Rotunno and Klemp 1985; Wilczak et al. 1992) or through the stretching of ambient vertical vorticity by local updrafts along a convergence line.

In Fig. 16, the horizontal vorticity field at 1600 and 0.2 km AGL is plotted showing the orientation of the vectors relative to the misocyclones and associated updraft regions. Based on the Szoke and Rotunno (1993) hodograph at 1600, where the shear vectors were oriented to the south below 2 km AGL, the environmental vorticity vector pointed to the east at 90° in the 0–2-km AGL layer. This agrees with the vector orientation in the relatively quiescent regions away from the misocyclones and updrafts in Fig. 16a. Horizontal vorticity magnitudes are on the order of $25 \times 10^{-3}$ s$^{-1}$. Immediately southeast of the vortices in Fig. 16a, the horizontal vorticity vectors are oriented to the northeast, parallel to the surface convergence zone (see Fig. 6) and shaded updraft regions. The same orientation of vectors was observed at 2.2 km (not shown) with slightly higher magnitudes on the order of $30 \times 10^{-3}$ s$^{-1}$. Examination of the vorticity field in the vicinity of each of the misocyclones with time showed little change in vector orientation from that shown in Fig. 16a. Only small areas immediately to the northeast and southwest of the vortices intermittently had vectors oriented normal to the misocyclones; areas that would likely be more favorable for upward tilting of horizontal vorticity.

Davies-Jones (1984), expanding upon previous related studies by Rotunno (1981) and Lilly (1982), defined the component of the mean vorticity along the storm relative wind, when the storm relative winds veer with height, as streamwise vorticity. At those levels where there is streamwise vorticity, he found that linear theory yields a positive correlation between vertical velocity and vertical vorticity. As seen previously (Fig. 10), a high correlation exists between the vertical vorticity and velocity fields associated with the T1 misocyclone. In Fig. 16b, vorticity vectors and storm relative streamlines encompassing T1 are nearly parallel along the east-northeast and southwest quadrants of the misocyclone. Plots of the streamwise vorticity component (not shown) showed large, positive values in these regions, and near-zero values in the other quadrants. Streamwise vorticity was found in similar locations, and at heights above 0.2 km, surrounding T2 and T3 during the evolution of these events. The existence of streamwise vorticity in the presence of preexisting surface vortices indicates regions of continued enhancement of vertical vorticity. The orientation of this flow and the horizontal vorticity vectors directed into the misocyclone in Fig. 16b, is consistent with flow features presented by Klemp and Rotunno (1983) and Rotunno and Klemp (1985), especially in the latter's Fig. 8. They emphasize the importance of this flow into the surface misocyclone, along the forward-flank outflow, for the development of near-surface vertical vorticity. In Fig. 16, enhancement of vertical vorticity is not occurring along the forward-flank of the storm, but rather along the north side of the storms where the only surface convergence boundary exists. Thus, while environmental horizontal vorticity did not apparently contribute much to the formation of the misocyclones, some positive contribution to increased vortex development would be expected due to tilting of horizontal vorticity in those areas where positive streamwise vorticity was observed.
shown are vertical vorticity advection ($\zeta_v$), stretching ($\zeta_{\text{DIV}}$), and tilting ($\zeta_{\text{TILT}}$). At 0.2 km (not shown), $\zeta_{\text{DIV}}$ is large, reaching a maximum value of $100 \times 10^{-3}$ s$^{-1}$, centered within the $10 \times 10^{-3}$ s$^{-1}$ vortex region. The intensity of this term decreases with height (see Fig. 17). The $\zeta_v$ and $\zeta_{\text{TILT}}$ terms could not be computed at 0.2 km. At all heights shown in Fig. 17, $\zeta_{\text{TILT}}$ and $\zeta_v$ clearly contribute little to production and redistribution of vertical vorticity when compared with the magnitude of the $\zeta_{\text{DIV}}$ term. Positive tilting does occur on the SW side of the vortex where the northerly influx of flow (see streamlines in Fig. 17) encounter the storm updraft, but its spatial extent is only a fraction of the area of positive $\zeta_{\text{DIV}}$. Not shown in Fig. 17 is the horizontal vorticity advection ($\zeta_h$). The positive maximum for this term was located on the east side of the vortex and a negative maximum was located along the west side, consistent with the movement of T1 toward the east. The magnitude of this term is equivalent to the divergence term at low levels. This advection term increased by a factor of 4 from 0.2 to 3.2 km. The distribution and comparative magnitudes of the three vorticity terms at other time periods during the pre-tornado stage is very similar to the distributions shown in Fig. 17. Thus, within the T1 mesocyclone at low levels (0.2–2.0 km), vorticity divergence and horizontal redistribution of vertical vorticity dominated vorticity production. Above 2.0 km, vortex intensification was attributed to a redistribution of positive vertical vorticity from low altitudes to upper levels.

2) T2 VORTEX

The vertical vorticity terms, excluding the horizontal advection term, within the vicinity of the T2 rotation are shown in Fig. 18 at 1619, just prior to tornadogenesis. As observed with T1, the intensity of the divergence term is much stronger than the other terms near the surface and maximum, positive values are located within the vortex with height. The maximum value of $\zeta_{\text{DIV}}$ at 0.2 km (not shown) within T2 was greater than the T1 divergence value. The tilting term has a positive maximum along the southwestern side of the vortex from 0.8 to 2.0 km. A significant region of positive tilting, comparable in extent to the positive $\zeta_{\text{DIV}}$ area, is apparent at 1.4 km. This maximum straddles the region between the updraft and downdraft maxima and during the stage when the T2 downdraft intensifies (see Fig. 12e). Large, positive values of $\zeta_v$ at 2.0 km associated with the updrafts within the circulation, indicates that near-surface vorticity was redistributed vertically prior to tornado onset. Thus, at this time period, the stretching and tilting terms appeared to be the greatest contributors to vorticity production below 2.0 km, while stretching and vertical advection of surface vertical vorticity provided most of the contribution to vorticity production at 2.0 km.

It is perhaps more informative to look at the trend of these vorticity terms with time and at a fixed height

b. VORTICITY PRODUCTION TERMS

1) T1 VORTEX

Figure 17 shows the horizontal distribution of three of the vorticity terms in (2) associated with the T1 vortex at 0.8-, 1.4-, and 2.0-km heights and volume time 1600, just prior to tornadogenesis. The terms
Fig. 17. Distribution of vertical advection, stretching and tilting vorticity terms at three heights within the vicinity of the T1 vortex at 1600 MDT, 4 min prior to tornadogenesis. The three heights shown are 0.8, 1.4, and 2.0 km. Vertical vorticity values of 10 and 20 x 10^{-6} s^{-2} are the heavy black contours. Smaller magnitudes of vertical vorticity are not shown but were confined to the gradient in vorticity surrounding the T1 vortex (arbitrarily defined as the area within the 10 x 10^{-6} s^{-2} contour). Shaded (hatched) areas are positive (negative) contributions to vertical vorticity production. Storm relative streamlines (thin gray curves) are also overlaid.

during the pretornado stage to assess which terms, if any, were dominant in the formative stages of the T2 vortex. Figure 19 shows the $\zeta_d$, $\zeta_t$, and $\xi$ terms at 0.8-km and 5-min increments, starting at 1604, 18 min prior to tornado onset. It can be seen from the storm relative streamlines that inflow into the vortices (T2a and T2b at 1604) is initially from both the north-northeast and southern directions, evolving into circular flow around T2 as the vortex intensifies (1614). What is most dramatic in Fig. 19 is the lack of any appreciable positive tilting during the early stages of vortex development and intensification with either T2a or T2b or the just merged T2 at 1609. It is clear that initial vortex development is due to stretching and vertical advection.
of vorticity. As T2 intensifies immediately prior to tornadogenesis, the $\zeta_{SHV}$ term is still strong, but the patchwork of positive tilting found in the northerly inflow region is now contributing to the enhancement of the vertical vorticity at 0.8 km and more significantly at 1.4 km (Fig. 18). Not surprisingly, the maximum positive values $\zeta_{DIV}$ and $\zeta_{TILT}$ at 0.8 km are located in the region of storm updraft and strongest confluence of surface inflow (see Fig. 18).

3) T3 VORTEX

The magnitudes and signs of the vorticity terms near the T3 vortex present quite a contrast to those observed with T1 and T2. Figure 20 shows these terms at 1614, just prior to tornadogenesis. The maximum value of the divergence term is significantly stronger at 0.2 (not shown) and 1.2 km than was observed with T1 and T2. Storm-relative inflow
from the north-northeast converges into the T3 vortex along its southern side, where the largest values of $\zeta_{\text{DIV}}$ are found. The intensity of $\zeta_{\text{DIV}}$ decreases in magnitude with height. A region of negative $\zeta_{\text{DIV}}$ is observed at 1.2 km within the $10 \times 10^{-3}$ s$^{-1}$ vortex, similar to observations within the T2 vortex, decreasing the overall contribution to positive vorticity production by this term. Large, negative values of $\zeta_{\text{TILT}}$ are also evident at 1.2 and switch to positive values at 2.2 km. The negative tilting at 1.2 is counterbalanced by a region of large, positive $\zeta_{\text{V}}$ at 1.2 km that was not observed with T1 and T2. Once again, the divergence and advection terms appear to provide the greatest contribution to vorticity production below.
2.2 km, with divergence and tilting both contributing to vortex intensification aloft.

8. Summary

The parent storms associated with the three tornadoes studied here were from a multicellular line of storms that initiated above the collision of two surface gravity currents, formed in an environment characterized by conditional instability, modest vertical wind shear, and high CAPE. Both outflows caused 2°C drops in surface temperature following their passage, with the coldest air located several kilometers south of the collision area. Collision occurred after convective temperature had been reached and the boundary layer had become well mixed in the vicinity of the dissipating DCVZ. Two of the storms, T1 and T2, exhibited characteristics often associated with supercell storms: a WER, a BWER, a pronounced vault with T1, and hook appendage on the back side of the storm. Surface rotation and updraft centers were preferentially located in the regions of strong surface convergence and along the northwest sides or right flanks of the storms, consistent with the wind shear profile and also with supercell modeling studies. No midlevel mesocyclones resulting from tilting of ambient horizontal vorticity were observed prior to the tornado genesis process within any of the storms. Low-level inflow into the storms came from the southeast and from the northerly flow riding up over the collision zone, creating a broad line of 10–20 m s⁻¹ updrafts parallel to the convergence line. This line of strong updrafts controlled storm development and mesocyclone evolution. We believe that it is the collision of the two outflow boundaries in the vicinity of the DCVZ, coupled with the stronger than normal atmospheric instability and vertical wind shear, that likely caused these nonsupercell tornadoes to be more intense (F2–F3) than normal. No forward-flank, precipitation-driven downdrafts and resultant outflows, believed to be important in supercell tornadoogenesis, were observed prior to tornadogenesis for the three tornado cases presented here.

A few minutes prior to the collision of the two surface outflows, apparent barotropic shearing instabilities were initially detected along the southern outflow. At least two of these surface circulations persisted for 10
min as apparent 2D Helmholtz instabilities before being affected by convection overhead. Coincident with the rapid storm growth overhead, four of these circulations were observed to build in height and intensified with time into vortices of tornado strength. Two of these vortices, T2a and T2b, were observed to merge at all heights about 10 min prior to tornadoogenesis.

Horizontal rolls were not detected by the radars. Analysis of the dual-Doppler, derived horizontal and streamwise vorticity fields showed that positive streamwise vorticity was observed along the northeast and southwest sides of the preexisting surface vortices, coinciding with the updraft regions, indicating areas of continued vortex enhancement and potential regions of upward tilting of horizontal vorticity.

During the pretornado stage, the T1 vortex was located along the gradient in upward vertical velocities. During the tornado stage, the maximum updraft core and maximum vertical vorticity were collocated. Downdrafts appeared only during the tornado stage, were located outside the main storm area and failed to penetrate to the surface. The vorticity divergence term was large and positive in the lowest 2 km subcloud layer. It dominated contributions to vertical vorticity production by the tilting and advection terms during the period leading up to tornadoogenesis. Immediately prior to tornadoogenesis, vertical advection of vertical vorticity also became significant above 2.0 km, redistributing positive vertical vorticity to higher altitudes. This case illustrated that downdrafts and associated outflows need not play a role in tornado formation. Tornadoogenesis resulted from convergence and stretching of the surface vortex by the storm updraft.

The merging T2 circulations were observed first outside the primary updraft region and with time the T2 vortex straddled the maximum updraft and downdraft centers within the storm during the pretornado stage. Immediately prior to tornadoogenesis, the updraft intensified significantly coincident with a developing downdraft and outflow that acted to strengthen the low-level (2.2 km) convergence into the vortex. Corresponding single Doppler vorticity estimates showed a dramatic intensification of the vortex at this time following the merger of T2a and T2b. The downdraft and resultant outflow were obviously important in the formation of the T2 tornado. During the pretornado and tornado stages, the divergence term dominated the other two forcing terms, contributing large positive values to vertical vorticity production in the 0.2–2.0 layer. Redistribution of surface vorticity vertically also contributed to the enhancement of the T2 vortex during its initial stage of development. Tilting of vorticity was not present during the formative stage of the misocyclone but was observed to contribute positively to vorticity production immediately prior to tornadoogenesis, enhancing the strong contributions arising from vertical stretching of the vortex. Tornadoogenesis appears to have been mainly a consequence of stretching and vertical transport of the preexisting surface vorticity by an intensifying storm updraft. The intensification of the downdraft and resulting outflow acted to increase the horizontal convergence into the misocyclone, thereby further tightening the vortex to tornado intensity.

The vertical motion field encompassing the T3 vortex was quite variable. The vortex initially straddled the maximum updraft and downdraft centers during the pretornado stage. At tornado time, the downdraft disappeared only to reappear again later in the tornado stage. This downdraft was observed early on, at least 20 min prior to tornadoogenesis, when the storm was in its earliest growth stage and prior to any appreciable vortex intensification. Limitations in being able to resolve motions on the actual tornado vortex scale using dual-Doppler analyses might explain the variability in motions observed in this case. The vorticity production terms also exhibited more variability in magnitude than observed with the T1 and T2 events. The general trend, however, showed the divergence term clearly dominating the contribution to positive vorticity production at heights at and below 2.2 km. Rapid intensification of the divergence magnitudes were observed leading up to tornadoogenesis. Vertical advection of vorticity was generally weak during the pretornado stage, but again showing a noticeable increase in magnitude at 1.2 km just prior to tornadoogenesis. While tilting values were mostly negative in the vicinity of the T3 vortex at 1.2 km, just prior to tornadoogenesis, large positive values were observed at 2.2 km and above, collocated with the maximum vertical vorticity and updraft centers. While the sequence of events leading up to T3 tornadoogenesis exhibit more variability, it seems clear that vorticity stretching by the storm updraft was crucial to tornado development. The variability in downdraft detection, especially the lack of a detection at tornado onset, makes its significance in tornadoogenesis uncertain. Vertical tilting of vorticity appears to have enhanced vortex intensification in this case.

9. Discussion

This study has shown that all three tornadoes originated from misocyclones in the boundary layer. These misocyclones intensified and extended vertically in association with rapid storm growth overhead. They were clearly associated with the maximum storm updrafts. The dual-Doppler results presented herein combined with the single Doppler studies of Brady and Sjokke (1989) and Wakimoto and Wilson (1989) indicate two necessary conditions for nonsupercell tornadoogenesis: 1) the existence of a boundary layer misocyclone generated along a convergence line and 2) the juxtaposition of the misocyclone with a strong updraft of a convective cloud. Downdrafts did not play a significant role in the genesis of the first tornado in this study. Thus, a downdraft is not a necessary condition for the formation of a nonsupercell tornado. However, in any
one case, a localized downdraft like an occlusion downdraft that is dynamically induced, may play a significant role in whether a tornado develops by enhancing the convergence and updraft in the vicinity of the surface vortex.

The natural question to ask is whether the above necessary conditions for nonsupercell tornadoes hold for supercell tornadoes. Some observations seem to argue that the tornado vortex descends to the ground. This is certainly true when visually observing the condensation funnel; however, it is not necessarily the case for the vortex. As reviewed in section 2, there is both observational and theoretical evidence that tornadogenesis cannot be simply explained as the descent of the mesocyclone to the surface. Most of the supercell tornado literature show the tornado occurs along the boundary layer convergence line generated by the forward flank downdraft and storm inflow. This is a vorticity rich environment where surface mesocyclones are likely to form, as postulated by Brandes (1977). Because of the small-scale nature of this vorticity, there simply has not been adequate observations to determine if a boundary layer mesocyclone is a necessary condition for a supercell tornado. Answers to this question will have to await more detailed observations.

Downdrafts and attendant outflows are always present within supercell storms, however, it is unknown if they are a necessary condition to trigger a tornado. Klemp and Rotunno (1983) noted that rear-flank downdrafts and occlusion downdrafts prevail in many long-lived severe thunderstorms that do not produce tornadoes. And yet, as with nonsupercell tornadoes, localized downdrafts may be instrumental in determining the timing and intensity of a tornado through their affect in increasing low-level convergence and companion updrafts, as was observed by Brandes (1984). A higher density of surface mesonet stations, finer mesh numerical models and/or very high spatial resolution, rapid scan radars are needed to address these flow interactions on the scale of the tornado vortex.

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APPENDIX

Accuracy of the Dual-Doppler-Derived Fields

The accuracy of the dual-Doppler-derived wind field depends first upon the quality of single Doppler radial data collected. Accuracy of the radial velocity estimates relies on collection of a representative number of (radial) samples per pulse volume. Using the known radar parameters, number of radar samples, and the mean spectrum width and signal-to-noise values within the tornado-producing storms, the radial velocity estimates within the storms are likely accurate to within ±1 m s⁻¹ for FL-2 and UND (see Fig. 6.6 in Doviak and Zrnić 1984). Errors due to radar geometry and beam effects also affect the accuracy of both the radial and reflectivity estimates. Because of the small radar baseline, sampling of the low-altitude divergence was possible to within a few hundred meters above the ground and necessarily assumed to be fairly representative of surface magnitudes. The close distance of the two radars also helped to minimize radar geometry effects on vertical velocity estimates. These errors were computed to be generally less than 1 m s⁻¹ within the boundary layer. Radar data was edited for removal of sidelobe echo, ground target return, second trip return, and for dealing with data exceeding the velocity Nyquist interval, minimizing errors due to the radar beam and radar processing.

Synthesis and retrieval of the wind and derived fields is constrained by the temporal and spatial resolution of the data collected. To account for storm advection during the 2.5-min volume collection and temporal differences in data collection between FL-2 and UND, average storm motion vectors for 20-min periods were used to adjust nansultaneous data to a central time. Storm motion was calculated by tracking the storm reflectivity core and updraft regions with time. Due to the often rapidly evolving nature of convective storms, some error is implicit in the storm structure and wind
field collected in any given 2.5-min volume. For the data presented here, consistency in specific features over time provided confidence that the overall error due to storm evolution was minimal.

The dual-Doppler data are analyzed to a grid with 400-m spacing; 400 m being slightly larger than the largest dimension of a radar pulse volume in the area of interest and thus the distance for independent samples (Gal-Chen and Wyngaard 1982). Specification of an appropriately sized influence radius for the Cressman technique used to interpolate data to a Cartesian grid helped minimize the effects of noisy data (Carbone et al. 1980), as did the two-step Leise filter employed, but also limited the ability to resolve features having wavelengths of less than 1.2 km. However, six independent samples per wavelength will ensure that approximately 75% of a 2.4 km wave can be resolved, following the arguments of Carbone et al. (1985). The vertical velocity values are smoothed because they are obtained from finite-difference derivatives of the horizontal winds on the 400-m grid. Other first-order derivatives, such as divergence and vorticity, were derived using centered finite-difference calculations over at least three grid points, and the error in these fields is likely small if errors in the horizontal winds are fairly small. However, the tendency for these errors to magnify exists when computing second derivative fields such as the vorticity production terms.

The most potentially serious errors in the analysis are likely due to upward integration of the continuity equation. Lack of an upper boundary condition for $w$ due to incomplete coverage of storm top precluded the ability to integrate downward and obtain a variationally adjusted estimate of $w$ (Kessinger et al. 1987). Under the assumption that the divergence field was constant between the surface and 0.2 km (the first level of our analysis), an initial lower boundary value of $w$ was set equal to a fraction (0.2) of the integrand value, as a first estimate of $w$ above the surface. The close proximity of the storm to the radars meant that computations of horizontal winds at midstorm levels likely contained large contributions of $W$ (Carbone 1985), where $(W = w + V_y)$. This contribution of $W$ to $u$ and $v$ estimates is removed using the iterative method of computing $w$ in CEDRIC. Weighing all the above factors the $u$ and $v$ wind components are likely accurate to $\pm 1-2$ m s$^{-1}$. The accuracy in $w$ is estimated to range from 1 to 2 m s$^{-1}$ near the ground to as much as 5-7 m s$^{-1}$ at 6 km. Equally important to our case study of tornadoes is the accuracy in the location of $w$. Given the good temporal resolution of the data and high spatial resolution of the horizontal winds below 6 km AGL, the locations of various features in the data are felt to be sufficiently resolved. However, resolution of vertical velocity and vorticity values on scales less than approximately 1.5 km are not well resolved.

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