Doppler Radar Observations of an Asymmetric Mesoscale Convective System and Associated Vortex Couple

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ABSTRACT

A study of the 28 May 1985 "asymmetric" mesoscale convective system (MCS) observed during PRE-STORM is presented. Dual-Doppler analysis revealed a well-defined cyclonic mesovortex in the northern portion of the stratiform region, in accord with previous studies on this type of circulation. At maximum extent, the closed cyclonic circulation had a diameter of approximately 80 km and a depth of 7 km. A smaller anticyclonic circulation was present to the south of the cyclonic vortex. Counterrotating vortices in asymmetric MCSs have been identified in recent modeling studies; however, prior to this study, observational confirmation of the anticyclonic vortex has been elusive.

1. Introduction

The study of mesoscale convective vortices (MCVs) (Menard and Fritsch 1989) associated with MCSs has greatly increased following the 1985 Oklahoma–Kansas Preliminary Regional Experiment for Storm-scale Operational and Research Meteorology (PRE-STORM) (Cunning 1986). Previous observations of such features were few and far between, since MCVs are generally too small and short-lived to be detected by conventional observations, such as the NWS sounding network. The PRE-STORM mesonetwork of sounding and surface stations was designed to resolve such mesoscale features. The detail provided by dual-Doppler radar analyses is often the best for resolving MCVs, yet these features can be too large and long-lived for complete observation, especially for ground-based dual-Doppler networks. These problems become even more apparent when single Doppler radar is used, but nonetheless, observations of MCVs have been obtained (e.g., Stirling and Wakimoto 1989). Brandes (1990) and Brandes and Ziegler (1993) used sounding data to document the kinematic structure of an MCV observed on 6–7 May 1985 during PRE-STORM. This method proved useful because of the large areal extent of the vortex, yet their analysis likely smoothed some of the finer-scale structure of the MCV. Compositing of sounding, dual-Doppler and wind profiler data has been used to increase coverage areas in situations where the MCV was not completely in dual-Doppler range (Biggerstaff and Houze 1991; Johnson and Bartels 1992). Satellite data has also been used to document MCVs (Johnston 1981; Bartels and Maddox 1991), yet this method is useful only in detecting mesoscale circulations after the parent MCS and upper-level cloudiness have largely decayed. Kinematic analyses of MCVs using dual-Doppler data have been presented as well (Houze et al. 1989; Verlinde and Cotton 1990; Johnson and Bartels 1992; Keenan and Rutledge 1993; Jorgensen and Smull 1993; Brandes and Ziegler 1993). The complex nature of the MCVs in these cases required that the dual-Doppler-derived kinematic fields be averaged over some fixed horizontal domain, with the exception of the Verlinde and Cotton (1990) study. The averaging in these cases can be subject to aliasing from convective scales, thus possibly diluting the kinematic depiction of the MCV. This study will present a kinematic analysis of an MCV using dual-Doppler data and an averaging method employing a variable horizontal domain, thus allowing the domain in which the MCV is contained to expand, contract, and change shape or orientation, even with height. Since MCVs are neither cylindrical nor rectangular, this technique provides a more realistic view of the average vortex kinematics. Verlinde and Cotton (1990) also allowed for a variable domain, allowing for variably sized circles with height.

Since even the best observational networks cannot often provide detailed kinematic and thermodynamic analyses of the entire life cycle of MCVs, numerical models have been employed to fill the gaps in our

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knowledge (e.g., Schubert et al. 1980; Hertenstein and Schubert 1991; Zhang and Fritsch 1987; Zhang 1992; Weisman 1992; Weisman 1993; Skamarock et al. 1994; Davis and Weisman 1994). These modeling studies provide the detailed kinematic and thermodynamic analyses of MCVs, which, along with observational studies, have lead to theories on the generation and maintenance of such features.

The motivation for the study of MCVs is driven by their longevity and their ability to regenerate convection after the parent MCS has dissipated. In a climatology of MCVs, Bartels and Maddox (1991) found that convection reforms in the vicinity of MCVs approximately 50% of the time. Furthermore, this convection is more likely to become organized into a new MCS than it is to remain disorganized. It has been estimated that MCVs contribute 30%–70% of the warm season rainfall over the United States (Fritsch et al. 1986). Also, MCVs have been known to produce flash floods (Zhang and Fritsch 1987), frequent lightning (Rutledge and MacGorman 1988; Rutledge et al. 1990), large hail, as well as occasional tornadoes. Thus, if MCVs are a frequent precursor to MCS regeneration, then a better understanding of how MCVs form and how they focus convective activity would have a large impact on the short-term forecasting of heavy rainfall and severe weather.

A brief synoptic setting of the genesis of the 28 May 1985 MCS and its associated MCV will be provided in the following section. Section 3 will present surface features identified by the PRE-STORM mesonet. Single radar data will be presented in section 4, whereas dual-Doppler analyses of the MCS as a whole, and the cyclonic and anticyclonic vortices in particular, will be the subject of section 5. We conclude with a discussion on how recent modeling studies relate to the observations described in this study.

2. Synoptic setting

The large-scale features associated with the initial development of the MCS are examined using the conventional National Weather Service (NWS) surface and upper-air network. The high-resolution PRE-STORM sounding network was not operated for this MCS, which was not forecasted. A surface analysis at 0300 UTC 28 May (not shown) depicted a weak low pressure center over eastern Indiana with a trailing cold front extending through the Ohio Valley to southwestern Missouri. The front was stationary from the Kansas—Oklahoma border to the Front Range of the Rocky Mountains. Along the stationary front near the Kansas—Oklahoma border, the low-level flow was convergent and dewpoints were in excess of 16°C. The 0000 UTC 850-mb analysis (Fig. 1a) showed a pocket of warm, moist (high $\theta_e$) air being advected into the PRE-STORM region by a low situated in the central Rocky Mountains. At 700 mb (Fig. 1b) a weak short wave was moving out of the Rocky Mountains into a ridge of high-$\theta_e$ air. This feature appeared to be more of a weakening of the ridge than a short-wave trough at 500 mb (Fig. 1c). A climatology of MCVs by Bartels and Maddox (1991) states that MCVs generally form under long-wave ridges at 500 mb, which is consistent with this scenario.

3. Mesonet analysis

Locations of the PRE-STORM mesonet stations used in the following analysis are shown in Fig. 2. Lo-
locations of the National Center for Atmospheric Research (NCAR) CP-3 and CP-4 radars, and the dual-Doppler analysis domains used in this study are also indicated.

At 0935 UTC (Fig. 3a) the surface $\theta_e$ and wind fields indicated that the squall line associated with this MCS was entering the PRE-STORM mesonet. The synoptic flow was generally light southeasterly, but the northwestern PAM (Portable Automated Mesonetwork) stations were reporting northerly winds in excess of 10 m s$^{-1}$. In the vicinity of the wind shift there was also a tight gradient in the $\theta_e$ field. This appears to be the gust front associated with the convective line; the single radar analysis in section 4 will confirm that the convective system was moving through this region at this time. The synoptic flow was advecting high-$\theta_e$ air over the gust front, fueling the convection. Convergence between the southeasterly flow ahead of the MCS and the outflow from the MCS acted to focus upward motion at the gust front, intensifying the convection. At 1050 UTC (Fig. 3b) the gust front had moved to the southeast some 65 km. In the equivalent potentially cool air behind the gust front the winds were highly divergent, suggesting the presence of an organized downdraft. Doppler radar data presented below will confirm the presence of a mesoscale downdraft at this time. Note the gradient in $\theta_e$ was strongest on the southern end of the system. A recent modeling study by Skamarock et al. (1994) also identified a strong $\theta_e$ gradient on the southern end of a MCS, which resulted from the turning of the mesoscale outflow to the right by the Coriolis force, thereby strengthening the $\theta_e$ gradient and low-level convergence. The enhanced gust front acts to strengthen the convection on the southern end of the MCS, which in turn strengthens the gust front, creating a positive feedback situation. The radar analyses in sections 4 and 5 will illustrate that the convection was indeed strongest on the southern edge of the MCS. Figures 3c,d show the continued progression of the MCS to the east-southeast, with the strongest gust front surge on the southern edge of the system. By 1315 UTC (Fig. 3d), the mesoscale outflow was affecting the entire northern portion of the mesonet. At later times (not shown) the gust front appeared to slow in its southerly progression across the mesonet.

Surface pressure features are depicted in Fig. 4. At 0935 UTC (Fig. 4a), there was a localized high pressure area (mesohigh) entering the northwestern corner of the network behind the gust front. The mesohigh has been attributed to cooling due to rainfall evaporation (Sawyer 1946), as well as latent and sensible cooling due to melting ice particles (Fujita 1959). An area of low pressure (wake low) trailed the mesohigh and a trough of low pressure (presquall low) preceded the gust front. The presquall low (Hoxit et al. 1976) and the wake low (Zipser 1977; Johnson and Hamilton 1988) have both been attributed to mesoscale subsidence warming generated by the convective system, although the mechanisms of such subsidence generating both the presquall and wake lows may be quite distinct. Pressure gradients across the southern portion of the gust front were less than or equal to 4 mb (100 km)$^{-1}$. The mesohigh and wake low had intensified and moved to the southeast by 1050 UTC (Fig. 4b), but the presquall trough weakened somewhat. An hour later (Fig. 4c) the mesohigh had continued its east-southeasterly progression, now with a central pressure of 959 mb. The wake low was not as pronounced as earlier, yet still was evident to the rear of the mesohigh. The 1315 UTC analysis (Fig. 4d) shows the mesohigh exiting the northeastern domain of the mesonet, while maintaining a central pressure greater than 959 mb. The wake low was now slightly deeper, with a central pressure below 955 mb. A low pressure trough ahead of the gust front was again evident in the southern domain of the mesonet.

Individual time series at two different PAM stations illustrate the difference in the progression of surface features on the northern and southern extremes of the MCS. The northern portion of the MCS is represented by the PAM 6 station (Fig. 5). The pressure trace (Fig.
5a) showed the passage of a broad mesohigh between 1030 and 1330 UTC. Pressures increased 2–3 mb as the MCS moved over the station. The pressure rise occurs before the arrival of cold air, which is consistent with the nonhydrostatic pressure effect due to a collision between two fluids of different densities (Wakimoto 1982). A very pronounced wake low, accompanied by high wind (Fig. 5b), followed the mesohigh. The pressure dropped nearly 8 mb in 10 min; 1-min average winds were sustained at 18 m s⁻¹ and wind gusts of 25 m s⁻¹ were recorded. Surface temperatures increased and dewpoints dropped at this time as well (Fig. 5d), consistent with the theory that wake lows are a hydrostatic response to strong subsidence warming on the back edge of the stratiform cloud (Zipser 1977; Johnson and Hamilton 1988; Stumpf et al. 1991). The southern portion of the MCS had very different surface features. The pressure trace at PAM 27 (Fig. 6a) was markedly different from the records at PAM 6. A strong pressure rise (4 mb) occurred from 1000 to 1100 UTC, but the high pressure jump was quite localized and did not last very long. At PAM 6 the pressure rise was not as strong. There was also no apparent wake low passage at PAM 27 and the high winds (Fig. 6b) were associated with the passage of a mesohigh, not a wake low. Furthermore, the passage of the strong mesohigh at PAM 27 was associated with a much cooler and drier air mass than at PAM 6, as ev-
Fig. 4. Surface mesonet analysis of station pressure hydrometrically reduced to 480 m MSL (mb, solid contours), wind bars (full—10 m s⁻¹; half—5 m s⁻¹): (a) 0935 UTC 28 May 1985; (b) 1050 UTC; (c) 1140 UTC; (d) 1315 UTC.

enced by the equivalent potential temperature (Fig. 6c), air temperature, and dewpoint temperature traces (Fig. 6d). The profound differences in the surface features in the northern and southern portions of the MCS can be attributed to the asymmetry in the precipitation field. The weaker, broader mesohigh and intense wake low are more characteristic of stratiform precipitation (northern portion of the MCS), while the stronger, more localized mesohigh and absence of a wake low is more typical of convection (southern portion of the MCS; Rutledge 1991).

Skamarock et al. (1994) have shown how the presence of Coriolis effects in 3D simulations of long-lived squall lines can lead to this type of asymmetry. As the system matures, and Coriolis effects have had enough time to act on the flow (~6 h), the ascending mesoscale front-to-rear (FTR) flow is deflected to the right (north), which preferentially advepts hydrometeors from the convective line into the northern portion of the system. This develops a broad stratiform precipitating region to the north. At the same time, the descending mesoscale rear-to-front (RTF) flow is deflected to the right (south). This descending RTF flow contains air that has been diabatically cooled through the processes of melting and evaporation of precipitation. This RTF flow contributes to the outflow or cold pool at the surface. As the cold pool is deflected to the south by the earth's rotation, buoyancy gradients and
enhanced convergence at the leading edge of the cold pool enhance convection (under conditions of favorable shear) in the southern portion of the MCS (Rotunno et al. 1988; Weisman 1992). The enhanced convection strengthens the cold pool, which provides positive feedback to the convection. The surface features presented above strongly support these hypotheses. For example, the cold pool in the southern portion of the MCS (Fig. 6c) had much stronger gradients and higher amplitude compared to the northern portion of the MCS (Fig. 5c). This deflection of the cold pool to the south created stronger convection to the south and resulted in weaker convection with a broad stratiform region to the north, as will become apparent in the following radar analysis discussion. The intense wake low on the back edge of the northern stratiform cloud and absence of a wake low to the south may also be a result of the asymmetry of the precipitation field.

4. Single radar analysis

a. Wichita WSR-57 radar data

Digitized data from the National Weather Service WSR-57 radar in Wichita, Kansas, captured a large perspective view of the low-level reflectivity patterns associated with this MCS. At 0900 UTC (Fig. 7a) the MCS was in northwestern Kansas and appeared to have a largely symmetric structure. Two hours later (Fig. 7b), the squall line was in central Kansas and had begun to show signs of asymmetry (Houze et al. 1990). The northern portion of the system had developed a trailing stratiform region, while the southern portion was largely composed of convection. The MCS had clearly become asymmetric by 1200 UTC (Fig. 7c). The northern stratiform region had broadened and notching on the back edge of the stratiform echo (to the north of the short east–west convection line) suggests a strong rear inflow jet had developed. This notching developed into a well-defined hooklike echo by 1335 UTC (Fig. 7d). As will be shown in the dual-Doppler analysis, this feature is not only indicative of a strong rear inflow jet advecting in drier air into the MCS, but is also indicative of a strong low- and midlevel cyclonic circulation advecting hydrometeors from the stratiform region to the north of the dry air intrusion. Smull and Houze (1985) and Sterling and Wakimoto (1989) have previously noted the association of cyclonic midlevel shear zones with comma-shaped
stratiform echo features. Although speculative, the smaller notch seen in Fig. 7c, immediately to the south of the short east–west convective line, may be associated with the presence of the anticyclonic vortex (discussed below). In this case the anticyclonic circulation would advect drier air into the MCS from the southwest, causing a similar erosion of the echo.

b. NCAR CP-3 radar data

The NCAR CP-3 C-band radar (which was situated roughly 100 km northwest of Wichita) afforded a slightly smaller, yet more detailed view of the MCS. Radial velocities were thresholded and unfolded, and power (dBm) was converted to reflectivity (dBZ) using the NCAR software package RDSS (Research Data Support System). The processed radar files were then objectively interpolated to Cartesian grids, with a 1.5-km horizontal and 0.5-km vertical grid spacing using the NCAR REORDER package. A Cressman (1959) weighting scheme was used in the interpolation process, with a horizontal radius of influence of 2 km and a vertical radius of influence of 0.75 km. The grids were rotated such that the x axis was oriented parallel to the direction of storm motion (toward 120°), thus storm velocity (16 m s⁻¹) could be removed from the radial velocity data in a vertical cross section through the radar location.

Horizontal cross sections of reflectivity at 3.5 km (AGL) are depicted in Fig. 8. This time series is illustrative of the development of the rear inflow notch and the associated intensification of the cyclonic mesovortex circulation. At 1120 UTC (Fig. 8a) the notching of the reflectivity field was becoming apparent about 100 km due west of the radar. At 1148 UTC (Fig. 8b), the notching had broadened and penetrated deeper into stratiform echo. Also at this time, the line of convection on the southern side of the rear inflow intensified. Approximately 35 min later (Fig. 8c) the inflow notching began to deform the northern portion of the stratiform echo into a hooklike pattern. This feature is an indication that the cyclonic mesovortex circulation was intensifying, which will be confirmed by the dual-Doppler derived wind fields in the following section.

Vertical cross sections of reflectivity and storm relative horizontal flow are depicted in Fig. 9 (along AA’ in Fig. 8a). The horizontal flow patterns were derived from radial wind data with elevation angles less than or equal to 20° to minimize contamination from vertical motions. The convective region was evident in the re-
Fig. 7: Low-level (0.7° PPI) reflectivity from the National Weather Service WSR-57 radar located in Wichita, Kansas:
(a) 0839 UTC May 1985; (b) 1106 UTC; (c) 1201 UTC; (d) 1335 UTC.
Reflectivity pattern from $x = 25$ to $50$ km, with the 30-dBZ contour extending up to $8$ km AGL. It appears that there was a progression of cells from the leading edge rearward. A mature cell was at $x = 45$ km and a decaying cell was centered at $x = 30$ km. This indicates that new convection was forming out ahead of the older convection as the low-level convergence between the gust front and the environment overtook the previously leading convective cell. This is consistent with the conceptual model of a mature squall line from Houze et al. (1989). The dashed line at $z = 3.9$ km indicates the melting level estimated from the 1200 UTC Oklahoma City NWS sounding. To the rear of the convective line, a noticeable “brightband” signature ($>30$ dBZ) can be seen immediately below the melting level.

Strong FTR flow ($V_{ref} < -25$ m s$^{-1}$) embedded within the into the convective line can be seen near $x = 40$ km. The region of general FTR flow sloped up-
ward toward the rear of the storm. A rear inflow jet descended from 8 km through the stratiform region to the surface near the convective line. Negative buoyancy resulting from the evaporative and sublimative cooling was likely responsible for the descent of the rear inflow jet through the stratiform region. The ini-
tiation of the rear inflow jet above 8 km is consistent with the theory of Stensrud et al. (1991), in which they describe a mesoscale downdraft and subsequent rear inflow induced by a sublimational process. They hypothesize that sublimation beneath the rear anvil of the trailing stratiform region creates negative buoyancy, which transports higher momentum air from upper levels to midevels, thus initiating the mesoscale downdraft and an inertially unbalanced rear inflow. At this point another mechanism may contribute to the acceleration of the rear inflow jet. Smull and Houze (1987) propose that rear inflow jets are a dynamical response to midevel mesolows in the trailing stratiform region. The midevel mesolow is believed to be a hydrostatic response to a positive buoyancy anomaly above the melting level (due to latent heat release in the mesoscale updraft) and a negative buoyancy anomaly below (due to evaporative cooling and melting; Brown 1979; Leary and Houze 1979). The rear inflow is accelerated toward the convective line by this midevel pressure gradient and accelerated downward by cooling due to sublimation, melting, and evaporation. Strong low-level convergence was situated at $x = 50 \text{ km}$, where the rear inflow jet extended to the surface and met the oncoming environmental flow.

5. Dual-Doppler analysis

Dual-Doppler analyses of this case were performed using data from the NCAR CP-3 and CP-4 Doppler radars, which were separated by 60 km in PRE-STORM, along a baseline oriented north-northwest/south-southeast. The radar volumes were edited and interpolated to Cartesian space as described in section 4. The gridded CP-3 and CP-4 radar data were then synthesized and the 3D wind fields retrieved using an interactive version of CEDRIC (Mohr and Miller 1983). A minimum beam-crossing angle of $30^\circ$ between the two radars was maintained. An O'Brien (1970) adjustment was imposed to ensure the vertical velocity $w$ vanished at the ground, applied after imposing $w = 0$ constraint at echo top. We report on three dual-Doppler analysis times.

The first analysis captured the southern anticyclonic vortex, which was located in the "west" lobe of the dual-Doppler domain (Fig. 2). The second analysis carried out in the "east" lobe of the dual-Doppler domain captured the mesovortex couplet in the stratiform region and a portion of the mature squall line. The final dual-Doppler volume (also in the east lobe) depicts the mature cyclonic vortex dominating the stratiform circulation. Houze et al. (1989) previously analyzed dual-Doppler data for this same case and illustrated the existence of the mature cyclonic vortex.

a. Analysis at 1050 UTC

A horizontal cross section at $z = 2.4 \text{ km AGL}$ (Fig. 10) reveals a storm-relative anticyclonic circulation on the southern flank of the MCS. This feature is quite similar to the southern counterpart of the "hooked vortices" as simulated by Weisman (1993), Skamarock et al. (1994), and Davis and Weisman (1994). The closed anticyclonic circulation was relatively shallow, being confined between 2 and 3 km AGL, and had an average relative vorticity of $-10^{-3} \text{ s}^{-1}$ and a diameter of at least 85 km at 2.4 km AGL. The low-level reflectivity field at this time (Fig. 10) shows that the southern portion of the MCS was composed mainly of convective elements, with a small but developing trailing stratiform region. There was strong southeasterly system-relative inflow into the convective elements, supplying high-$\theta_e$ air to the storm. The anticyclonic circulation was located in the developing stratiform region. The relatively low reflectivities on the northwest side of the vortex seem to indicate drier air was being entrained from the environment on the western edge of this system. Additional drying may have been caused by subsidence in this region. Subsidence on the western extreme of the stratiform cloud is confirmed at later times in section 5c.

An east–west cross section (Fig. 11a) taken along the line $AA'$ (as indicated in Fig. 10) depicts a deep and intense cell on the southern end of the MCS, possessing a reflectivity core of 45–50 dBZ extending to 8 km AGL. Convergence existed at low- and midlevels in the reflectivity core. Strong divergence was present above 10 km AGL. Maximum vertical velocities exceeded 15 m s$^{-1}$ near 9 km AGL. This cell most likely
of the convective line revealed complex and highly three-dimensional reflectivity and kinematic structures. The convection further to the north (Fig. 11b, cross section from BB' in Fig. 10) was consistent with the mature squall-line conceptual model depicted in Houze et al. (1989). This cross section indicates a leading convective cell \( (x = -15 \text{ km}) \), a decaying cell following the leading cell \( (x = -30 \text{ km}) \) and a transition zone \( (x = -45 \text{ km}) \) separating the convective and stratiform regions. A mesoscale FTR flow emanated from the convective line and converged with a descending RTF flow at midlevels near the back edge of the stratiform region \( (\sim -80 \text{ km}) \). The fact that more intense convection was on the southern flank of the MCS and that the less intense convection was situated further to the north is consistent with modeling studies of long-lived convective systems that include Coriolis effects (Skamarock et al. 1994).

The presence of cyclonic rotation on the northern flank and anticyclonic rotation on the southern flank of bow–echo-type storms was first postulated by Fujita (1978). Studies discussing the cyclonically rotating MCVs have been numerous in the past 10 years (e.g., Brandes 1990; Bartels and Maddox 1991; Brandes and Ziegler 1993; Jorgensen and Smull 1993), but confirmation of the anticyclonic counterpart to the cyclonic MCV has remained elusive. Recent modeling studies (Weisman 1993; Skamarock et al. 1994; Davis and Weisman 1994) have suggested that this anticyclonic feature may be a common component of long-lived convective systems. It may be that the anticyclonic vortex is shorter lived or smaller in scale compared to its cyclonic counterpart, thus rendering it more difficult to observe, as has been previously suggested by Smull and Weisman (1993).

Previous observational studies of (cyclonic) MCVs have relied on applying areal averaging techniques to get a more simplistic view of the vertical profiles of kinematic properties within an MCV (Jorgensen and Smull 1993; Brandes and Ziegler 1993; Johnson and Bartels 1992; Keenan and Rutledge 1993). This present study is no different in this respect, but a somewhat more sophisticated averaging technique is applied. Instead of defining a rigid boundary at midlevels encompassing the closed circulation and using the same boundary for averaging at all levels, or simply averaging over the entire horizontal domain, a variable horizontal domain based on a vorticity threshold is used. The selection of the vorticity threshold for the averaging domain is, admittedly, subjective, although it is a good way to isolate the closed circulation from the rest of the data. A vorticity threshold of less than \(-3 \times 10^{-4} \text{ s}^{-1}\) was sufficient to isolate the anticyclonic circulation in this case (as evident in the 1050 UTC dual-Doppler analysis). This value of vorticity was chosen because the closed circulation in the horizontal wind field was best represented by the area enclosed by this threshold. Although the wind vectors revealed

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**Fig. 11.** Vertical cross sections of reflectivity (dBZ, gray shading) and storm relative wind vectors (m s\(^{-1}\), wind vectors 4.5 km long indicate a 20 m s\(^{-1}\) wind) from the 1050 UTC dual-Doppler analysis: (a) along AA' from Fig. 10; (b) along BB' from Fig. 10.
that a closed circulation was confined between 2 and 3 km, the vorticity analysis showed that a strong anticyclonic vorticity signature extended above the closed circulation up to 7 km (presented below). We begin discussing the anticyclonic vortex by noting convergence (areally averaged over the vortex at each height; Fig. 12a) was situated above 3 km, with divergence below. The areally averaged vertical velocity (Fig. 12b) indicated strong subsidence below 5 km, with upward motion above this level. The lack of areal coverage in the dual-Doppler analysis below 2 km precluded any analysis of the closed circulation below this level. The average vorticity values within the vortex ranged from $-6.3 \times 10^{-4}$ s$^{-1}$ at $z = 1.9$ km to $-12.2 \times 10^{-4}$ s$^{-1}$ at $z = 4.4$ km (Fig. 12c). Vorticity tendencies (Fig. 12d) were also averaged over the vortex using the following equation:

$$\frac{D}{Dt} (\zeta + f) = - (\zeta + f) \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)$$

stretching

$$- \left( \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} + \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} \right) .$$  (1)

tilting

The solenoidal term could not be calculated due to the lack of thermodynamic data, thus the "total" tendencies indicated here assume the solenoidal generation of vorticity was small compared to the convergence and tilting terms; in other words, the total tendency we refer to is merely a summation of the stretching and tilting terms. This assumption is supported by the studies of Skamarock et al. (1994) on
simulated MCV circulations, who found the solenoidal generation of vorticity to be two orders of magnitude smaller than the generation of vorticity by stretching. The vertical profile (Fig. 12d) shows that the tilting term had a positive tendency (i.e., it was destroying the anticyclonic circulation) throughout the depth of the vortex at this time. The positive contribution by the tilting term was largest below 4 km. The snapshot of the vortex presented here shows that stretching was acting to locally destroy the vortex below 3 km because of the strong divergence in the wind field. It should be noted that stretching alone could not have produced the anticyclonic vortex (stretching cannot change the sign of the vorticity, assuming the initial absolute vorticity was positive), so it is possible tilting occurred at an earlier time to produce the initial vortex, which was then enhanced by stretching. Above the melting level, the stretching term dominated the tendency equation, contributing to the spinup of the anticyclonic circulation. The total contribution due to stretching and tilting was to destroy the vortex below and enhance the vortex above the melting level. The deep layer of convergence (Fig. 12a) contributed to the negative tendency (spinup of the anticyclonic circulation) in the stretching term at this instant in time.

b. Analysis at 1140 UTC

This particular analysis indicates the MCS contained a pair of counterrotating vortices at this time. Two low-level horizontal cross sections of storm-relative velocity and reflectivity are shown in Fig. 13. A closed cyclonic circulation is revealed in the northern portion of the trailing stratiform region. A closed anticyclonic circulation, nearer to the convective line, was present in the southern portion of the MCS. Above the 2.4-km level (not shown), the anticyclonic circulation was no longer closed and the cyclonic circulation appeared to tilt northward to the dual-Doppler domain. The relative vertical vorticity in the southern vortex was on the order of $-2 \times 10^{-5}$ s$^{-1}$, while the northern vortex was slightly weaker ($\sim 10^{-5}$ s$^{-1}$). The vortex pair was separated by a strong rear inflow jet (10–15 m s$^{-1}$) that penetrated to the convective line.

The cross section along $CC'$ (Fig. 14a) is representative of the southern, intense portion of the MCS, and the cross section along $DD'$ (Fig. 14b) is representative of the northern portion of the MCS, containing weaker convection and more extensive stratiform rain. The southern portion of the system was also characterized by much stronger system relative FTR flow than found in the northern portion of the MCS. The low-level RTF flow was stronger along $CC'$ than along $DD'$, which may have contributed to enhanced low-level convergence at the leading line, resulting in stronger forcing of the convection in this region. Simulations of long-lived convective systems by Skamarock et al. (1994) show that the deflection of the low-level cold pool to the right of the storm motion by the rotation of the earth creates enhanced low-level convergence and

Fig. 13. Low-level horizontal cross section of reflectivity (dBZ, gray shading) and storm-relative wind vectors (m s$^{-1}$, wind vectors 4.5 km long indicate a 20 m s$^{-1}$ wind) from the 1140 UTC dual-Doppler analysis: (a) 1.9 km; (b) 2.4 km.
c. Analysis at 1314 UTC

By 1314 UTC the intensity and depth of the cyclonic circulation had increased dramatically and the anticyclonic circulation was almost nonexistent. The cyclonic circulation was closed from 2 to 9 km, with the strong-

FIG. 14. As in Fig. 11 except from the 1140 UTC dual-Doppler analysis: (a) along CC' from Fig. 13; (b) along DD' from Fig. 13.

subsequently stronger convection on the southern side of eastward-propagating convective systems. The stronger $\theta_e$ gradient in the surface mesonet data in the southern portion of the dual-Doppler domain (see Fig. 3c) supports this hypothesis.

FIG. 15. As in Fig. 13 except from the 1314 UTC dual-Doppler analysis: (a) 2.9 km; (b) 6.9 km.
est vorticity located in vicinity of the melting level (3.9 km). At 2.9 km (Fig. 15a) the cyclonic vortex dominated the wind field in the northern portion of the stratiform region, while a remnant of the anticyclonic circulation was still present in the southern portion of the domain, \((x = 65 \text{ km}, y = 27 \text{ km}; \text{note the same relative position between the two vortices at this time and at 1140 UTC, Fig. 13b)}\). This weaker anticyclonic circulation was identifiable between 2 and 4 km at 1314 UTC. In sensitivity studies with Coriolis effects, Skamarock et al. (1994) found that the anticyclonic circulations in their bow-echo simulations were weakened considerably as the positive buoyancy anomaly in the mesoscale updraft (FTR flow) was deflected to the north by the earth's rotation over time. A similar effect may have contributed to the spindown of the anticyclonic circulation from 1140 to 1314 UTC. The reflectivity pattern at 1314 UTC (Fig. 15a) had become highly asymmetric due to the extended presence of the cyclonic circulation. On the northern side of the cyclonic vortex the reflectivities were relatively high (30–35 dBZ), because streamlines in this region originated from the decaying convective elements of the system; thus hydrometeors were being advected into this portion of the vortex, thereby elevating the reflectivities. Growth of hydrometeors in the mesoscale updraft would have also contributed to higher reflectivity in this region. On the southern side of the vortex there was a pronounced lowering of the reflectivity field where drier environmental air was drawn into the storm by the strong circulation, thus causing the hydrometeors to sublime and evaporate. Additional lowering of the reflectivity in this region can be attributed to subidence warming, enhancing the sublimation and evaporation processes in the mesoscale downdraft. This notchings in the reflectivity field was present in the single-Doppler analyses as well (Fig. 8), suggesting that the cyclonic circulation was present as early as 1120 UTC (dual-Doppler analysis could not confirm the presence of the cyclonic vortex as early as 1120 UTC, because the area of the MCS that contained the MCV at later times was over the baseline of the two radars at 1120 UTC). The circulation at 6.9 km (Fig. 15b) was still quite strong, but considerably weaker than the circulation at lower levels.

Vertical cross sections along \(EE'\) are shown in Fig. 16. The storm-relative velocity vectors (Fig. 16a) indicate deep front to rear flow in the trailing stratiform region of the MCS, with a rear inflow jet encroaching on the rear edge of the system. The rear inflow entered the storm as high as 9 km but was strongest near the melting level (3.9 km). The convergence field (Fig. 16b) indicated a maximum \((\gtrsim 2 \times 10^{-3} \text{ s}^{-1})\) in the same region where the rear inflow was strongest. There was also strong divergence below the convergence maximum, both of which contributed to the strong downward vertical velocity \((\lesssim -4 \text{ m s}^{-1})\) shown in Fig. 16c. Upon comparison of the vertical velocity and convergence features with the relative vertical vorticity field (Fig. 16d), it is seen that they were nearly coincident with the maximum in relative vertical vorticity \((\gtrsim 4 \times 10^{-3} \text{ s}^{-1})\). Strong, convective-scale downdrafts have also been identified along the trailing edge of the stratiform region by Stumpf et al. (1991), Hunter et al. (1992), and Bernstein and Johnson (1994). In these cases and the present case, the strong downdrafts lead to warming in this region, resulting in hydrostatic pressure falls and the subsequent formation of the wake low.

In the vorticity analysis that follows, a vorticity threshold of \(10^{-4} \text{ s}^{-1}\) was used to isolate the closed cyclonic circulation. The vortex area defined in this manner was then used to produce areal averages of relative vertical vorticity, vertical velocity, convergence, as well as terms in the vorticity tendency, Eq. (1). The resulting horizontal averages were then used to construct vertical profiles. This analysis was performed only on the levels where a closed circulation was evident. The areally averaged convergence profile (Fig. 17a) shows the existence of strong divergence at low levels followed by strong convergence at mid- and upper levels. The convergence within the vortex was deep, extending over 6 km. This deep layer of divergence was in part caused by the confuence of the opposing FTR and RTF flows in this region. Divergence at low levels existed within the vortex, even though there was strong rotation and possibly lower pressure. This can be explained by the fact that there were strong negative buoyancy effects (melting and evaporation) occurring below the melting level (3.9 km), which contributed to strong downward motion. The mass in the downdraft was forced to diverge as the effects of the ground began to feel. As would be expected with upper-level convergence and low-level divergence, the average vertical motion (Fig. 17b) within the vortex was downward. The effects of melting and evaporation can be seen in the layer between 2.9 and 3.9 km, where the areally averaged downward velocities exceed 30 cm s\(^{-1}\). The average vertical vorticity profile (Fig. 17c) shows a steady increase in vorticity from 1.9 km to a maximum of \(1.16 \times 10^{-3} \text{ s}^{-1}\) (or 13 times greater than the local Coriolis parameter) at the melting level, and then a steady decrease in vorticity to the top of the closed circulation.

The areally averaged vertical vorticity tendency profiles (Fig. 17d) show that at this time, the vortex was spinning up (a positive tendency) at nearly all levels. However, below 2.4 km, the tendency was strongly negative, due to the strong divergence creating negative stretching. The contribution of the tilting of horizontal vorticity into the vertical (below 2.4 km) was positive, yet not strong enough to counteract the negative tendency due to stretching. Throughout the depth of the vortex, the tilting term was generally opposite in sign and smaller in magnitude than the stretching term. The vorticity tendency maximum was just above the melt-
Fig. 16. Vertical cross sections along EE' (as shown in Fig. 15) from the 1314 UTC dual-Doppler analysis of reflectivity (dBZ, gray shading) and (a) storm relative velocity vectors (m s\(^{-1}\), wind vectors 4.5 km long indicate a 20 m s\(^{-1}\) wind); (b) convergence (solid) and divergence (dash) (10\(^{-5}\) s\(^{-1}\)); (c) positive (solid) and negative (dash) vertical velocity (m s\(^{-1}\)); (d) positive (solid) and negative (dash) relative vertical vorticity (10\(^{-5}\) s\(^{-1}\)). Note the heavy solid contour denotes the \(u' = 0\) contour.

...ing level (near 4.4 km). The stretching term was maximized at this level, in a region of high vorticity and strong convergence, while the tilting term was relatively small. This analysis suggests that during the mature phase of the MCV, stretching of existing vertical vorticity was the primary mechanism for the mainte-
Fig. 17. Vertical profiles of areally averaged quantities over the cyclonic vortex at 1314 UTC: (a) convergence (10^{-5} s^{-1}); (b) vertical velocity (cm s^{-1}); (c) relative vertical vorticity (10^{-5} s^{-1}); (d) relative vertical vorticity tendency (10^{-5} s^{-1} h^{-1}): tilting (TILT, small dash), stretching (STR, large dash), total (TOTL, solid).

nance of the circulation. The modeling studies of Weisman (1993), Davis and Weisman (1994), and Skamarock et al. (1994) propose that tilting of horizontal vorticity into the vertical by differential vertical motion is the primary mechanism for the genesis of MCV circulations, which are subsequently enhanced by horizontal convergence. It should be noted that this same genesis mechanism was also proposed by Brandes and Ziegler (1993), based on analyses of the vorticity budget in the 6–7 May 1985 PRE-STORM MCS. Since the MCV in this case was already mature at the time it was completely in the dual-Doppler analysis domain, we can speculate only on the mechanisms involved in the generation of the cyclonic circulation.

The vortex averaged profiles of convergence, vertical velocity, and vertical vorticity in this study are quite similar in structure, but dissimilar in magnitude to those found by Brandes and Ziegler (1993) using dual-Doppler data. Similar to this study, Brandes and Ziegler found strong mid- and upper-level convergence and low-level divergence in vicinity of the vortex, which contributed to strong downward motion. The magnitudes of the divergence and vertical velocities calculated from dual-Doppler data were two to three times stronger in the Brandes and Ziegler study compared to those found in this study. The vertical vorticity profile in Brandes and Ziegler was also quite similar, with the vorticity maximum near the melting level. However, the magnitudes of the average vorticity in their study were nearly an order of magnitude smaller than those found in this case. Their analysis of the vorticity tendency had a distinctly different profile. The tilting term dominated the stretching term and the peak vorticity production occurred at higher levels (6.5 km) than in this case. The peak amplification rates in Brandes and Ziegler (1993) were also about three to four times
smaller than those found in this study. These comparisons should be viewed with caution because of the rather ad hoc choice of averaging techniques, and the inherent resolution differences between the dual-Doppler and radiosonde data.

Jorgensen and Smull (1993) analyzed an MCV using airborne-Doppler radar data. The vortex in their study was shallower (~5 km) and weaker than the 28 May 1985 case. The peak relative vertical vorticity, again found near the melting level, was on the order of $3 \times 10^{-3} \text{ s}^{-1}$. The flow within the vortex was again convergent and downward. The amplification rates of the vortex were found to be an order of magnitude smaller than in this study.

This observational evidence shows the consistent presence of both convergent and downward air motion within MCVs, as well as peak rotation rates near the melting level. These findings correspond well with theories developed from numerical simulations that suggest mesolows near the melting level and stretching of vertical vorticity by convergence, are responsible for the strong low- and midlevel rotation found in the stratiform regions of MCSs (Bosart and Sanders 1981; Zhang and Fritsch 1987, 1988; Hertenstein and Schubert 1991).

6. Conclusions

Radar analysis documented the evolution of this convective system from a squall line, to a squall line with a trailing stratiform region (symmetric MCS), to an asymmetric MCS with a broad stratiform region to the north and stronger convection to the south. Recent observations (Loehr and Johnson 1995) and modeling studies (Skamarock et al. 1994) suggest that this is the typical life cycle of long-lived convective systems in the absence of large-scale forcing. This asymmetry in the radar echo pattern became strongly evident as the cyclonic circulation intensified, which advected hydrometeors from decaying convection northward and westward, while at the same time advecting drier environmental air from the rear of the system southward and eastward. The depth (~7 km) and intensity ($>4 \times 10^{-3} \text{ s}^{-1}$) of the MCV created a distinct hooklike appendage in the reflectivity field during the mature stage of the MCS. This leads to the question of how and why the MCV began to dominate the mesoscale flow characteristics of this MCS. The dual-Doppler analysis hints at some mechanisms that would feed back on each other, promoting the rapid growth of the circulation.

The presence of a strong and deep rear inflow jet provided three important mechanisms. First, the rear inflow jet provided mid- and upper-level convergence as it collided with the system relative FTR flow. Second, the rear inflow advected drier air into the system, producing negative buoyancy through the processes of sublimation, melting and evaporation. Both the mid- and upper-level convergence and negative buoyancy effects lead to strong downward motion and perhaps more importantly, strong gradients in vertical velocity at the back edge of the stratiform cloud. Third, the rear inflow jet had a maximum westerly velocity of 15 m s$^{-1}$ near the melting level (3.9 km), which created strong westerly shear below this level, augmenting the ambient shear. The horizontal vorticity created by the rear inflow jet and the ambient flow could then be tilted into the vertical by the strong gradients in vertical velocity. The areal-averaged vorticity tendencies showed that stretching contributed to positive vorticity at the same levels where the rear inflow jet was present, and was maximized just above the melting level (where the rear inflow was strongest). Thus, it appears that the presence of the rear inflow jet was instrumental in the amplification of the vortex circulation in this particular case.

It has been stated earlier that the rear inflow jet is believed to be a dynamic response to the midlevel pressure gradients induced by positive buoyancy anomalies aloft in the mesoscale FTR flow (Smull and Houze 1987). It has also been suggested that the rear inflow jet may produce vertical vorticity through the above mechanisms. The presence of the rotation itself can act to lower the pressure (see section 5c), which would intensify the midlevel mesolow and accelerate the rear inflow jet. The stronger stretching and tilting of vorticity by the rear inflow jet would contribute positively to the lower pressure at midlevels. Lower pressure itself could also intensify the circulation by way of a cyclostrophic response. This is how rapid amplification of the MCV could occur.

The hypothesis on the rapid intensification of the cyclonic vortex and the demise of the anticyclonic vortex in this case was based on the observations that correspond well with theories developed in numerical modeling studies. Unfortunately the dual-Doppler data were not available during the period in which the cyclonic vortex was intensifying (1148–1303 UTC), as other scanning strategies were being followed during this time period. Thus, the above hypothesis cannot be unequivocally supported with this dataset. Ideally, dual-Doppler and supporting thermodynamic data throughout the developing, mature, and dissipating stages of an MCS would be needed to identify all the mechanisms involved in the generation and maintenance of the counterrotating circulations found in long-lived convective systems. However, the inherent evolutionary nature of these systems makes them very difficult to observe (system lifetimes exceed 10–12 h), even when airborne Doppler radar is used (Smull and Weisman 1993). Numerical simulations, however, will continue to provide high-resolution data to develop theories for the evolution of circulations within MCSs. These numerically derived theories should then be validated by observations whenever possible.
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