Evolution and Structure of the 6–7 May 1985 Mesoscale Convective System and Associated Vortex

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ABSTRACT

Observations collected during the Oklahoma-Kansas PRE-STORM experiment are used to document the evolution and structure of the mesoscale convective system (MCS) that occurred on 6–7 May 1985. The storm began when a short squall line developed in an area of preexisting thunderstorm activity. Thunderstorm updrafts along the squall line lifted warm, moist air with its southerly momentum to the upper troposphere. A broad region of convective outflow and a mesoscale updraft region with a mean vertical velocity in excess of $15 \times 10^{-3}$ mb s$^{-1}$ were created. A stratiform rain area with an embedded mesovortex formed behind the squall line. The vortex resided beneath the deepest upper-level outflow.

The mesovortex altered the wind field and consequently became the principal organizational feature within the MCS. A descending current from the storm’s rear that, depending on location, extended from 1 km to the upper troposphere was intensified and focused by the vortex. The descending rear inflow had a peak vertical velocity of $10 \times 10^{-3}$ mb s$^{-1}$ and concentrated into a jet that passed to the south of the vortex. The intruding flow caused the precipitation and cloud fields to develop comma-like shapes and determined the distribution of kinematic parameters within the MCS.

Mesovortex vertical vorticity was a maximum ($25 \times 10^{-3}$ s$^{-1}$) at middle-storm levels where environmental air converged into the mesoscale downdraft but was also strong at lower levels where the mesoscale downdraft dominated. Stretching of preexisting vorticity seems the primary amplification mechanism at middle levels. Tilting of horizontal vorticity generated by baroclinicity in the rear inflow is given as an explanation for the low-level vorticity.

1. Introduction

A distinct meteorological phenomenon, which comprises groups of thunderstorms and produces a significant proportion of warm-season rainfall and severe weather in the central United States, was first described by Maddox (1980, 1983). Dubbed mesoscale convective complexes (MCCs), such storms initially were defined and investigated using earth-satellite images. Today there is considerable effort to document other features and the life cycle of MCCs (e.g., Cotton et al. 1983; Wetzel et al. 1983; Smull and Houze 1985; Leary and Rappaport 1987; Bosart and Sanders 1981; Fritsch et al. 1986; Johnson and Hamilton 1988; Rutledge et al. 1988; Menard and Fritsch 1989; Johnson et al. 1989). MCCs have structural features in common with other midlatitude mesoscale systems, which do not explicitly meet the MCC size criteria (Maddox 1980), and also with tropical squall lines and cloud clusters (Zipser 1969, 1977; Houze 1977; Roux et al. 1984). Collectively, all of these systems are referred to as mesoscale convective systems (MCS’s).

Environmental factors favorable for the formation of an MCS have been summarized by Maddox (1983). The genesis region is usually characterized by mesoscale convergence and lifting that associates with low-level temperature advection. Midtropospheric forcing is weak and nearly equivalent-barotropic. Many MCS’s develop during the evening in areas of preexisting, unorganized convective activity and over a period of several hours transform to a highly organized system, often with a squall line at the leading edge. Prominent features also include mesoscale updraft and downdraft regions and a trailing area of enhanced stratiform precipitation (e.g., Zipser 1969, 1977; Houze 1977; Smull and Houze 1985; Leary and Rappaport 1987). A typical storm has dimensions of several hundred kilometers and may persist for 12 h or more (Maddox 1980; Velasco and Fritsch 1987).

MCS’s feed upon low-level air with high moisture content that may be transported by a low-level jet. The inflow rises in thunderstorm updrafts within the convective zone. A portion of the condensate produced falls in rainy downdrafts at the rear of the convective zone. The remainder is transported rearward by convective outflow in the middle and upper troposphere. Heat liberated by condensation and residual water vapor is carried along with the outflow. As the buoyant
air continues to rise, a small additional amount of condensate is produced and more latent heat is released. A mesoscale updraft region forms in the upper troposphere from which stratiform precipitation falls. Depending on the evolutionary stage, a substantial portion of the total MCS precipitation (≥40%) falls as stratiform rain (Houze 1977; Gamache and Houze 1983; Johnson and Hamilton 1988).

Zipser (1969) described a cool, dry mesoscale downdraft that filled the rain area of a tropical system and hindered the development of new convection in the region traversed by the storm complex. The studies of Zipser, Ogura and Liou (1980), Leary and Rapaport (1987), and others suggest that the downdraft air originates at middle levels, either from the front or the rear of the system. Consensus holds that the mesoscale downdraft is initiated by evaporative cooling as precipitation falls from the overlying mesoscale updraft. The melting of precipitation may also accelerate the downdraft (Leary and Houze 1979).

Later, Zipser (1977) described a distinctive “onion shape” sounding created by subsiding air on the trailing edge of the stratiform rain area. The sinking air, which warmed almost dry adiabatically and became several degrees warmer than its environment, coincided with a wake low-pressure center [see Johnson and Hamilton (1988) for a recent example and additional references]. Zipser determined that subsidence warming of 2°C in the 1000–700 mb layer could account for an observed 2 mb depression.

Smull and Houze (1987a) have summarized environmental wind profiles for a number of MCS’s reported in the literature and found that, in a sizeable subset, a jetlike rear inflow began at middle or upper tropospheric levels on the back edge of the stratiform rain area and descended toward the front of the system. Subsidence was thought driven by pressure gradients at middle levels which resulted from hydrostatic adjustment to an elevated heat source in the upper troposphere and a heat sink in the lower troposphere (see also Zhang and Fritsch 1988a; their Fig. 20 and related discussion).

On occasion mesoscale vortices are observed within the stratiform rain region of MCS’s (Houze and Betts 1981; Johnston 1981; Smull and Houze 1985; Leary and Rapaport 1987; Johnson and Hamilton 1988; Stirling and Wakimoto 1989; Menard and Fritsch 1989). Numerical simulations (Zhang and Fritsch 1987; 1988a,b) suggest that warm core vertical circulations form as midlevel air converges to fill the void left by buoyant rising air in the mesoscale updraft and weak absolute vertical vorticity is amplified by stretching. The vortices may play a key role in organizing mesoscale convective systems (Zhang and Fritsch 1987; Velasco and Fritsch 1987; Menard and Fritsch 1989; Zhang et al. 1989). Inherently stable (Zhang and Fritsch 1987), mesovortices may persist for days—often with daily regeneration of convection.

An experiment, designed to increase the understanding of MCS’s, the Oklahoma–Kansas Preliminary Regional Experiment for STORM (or simply the Oklahoma–Kansas PRE-STORM Program), was conducted during the spring of 1985 (Cunning 1986). Here, PRE-STORM data are used to present a chronological description of a mesoscale convective system observed on 6–7 May. Features exhibited by the storm include a squall line, mesoscale updraft and downdraft regions, rear inflow, trailing stratiform rain, wake low, and a mesoscale vortex.

Much like in the 10–11 June 1985 MCS studied by Johnson and Hamilton (1988) and Rutledge et al. (1988), the rear inflow which fed into the mesoscale downdraft was of strong intensity (see Smull and Houze 1987a) and became a prominent storm feature. Subsidence warming, similar to that described by Zipser (1977) and by Johnson and Hamilton (1988), occurred as this current descended. But during the mature stage of storm development, sensible temperatures in the low-level unsaturated layer were actually cooler than in surrounding regions. As found by Johnson and Hamilton (1988), the wake low resided at the edge of the stratiform rain shield and was associated with the rear inflow.

The primary feature which distinguishes the 6–7 May mesoscale convective system from many studied previously is the strong mesovortex. Kinematic, thermodynamic, cloud, and precipitation fields were all altered by the presence of the vortex and were decidedly three dimensional.

2. Data sources and analysis procedures

The location of PRE-STORM data collection networks and sensors used in this study are shown in Fig. 1. The 80 station surface mesonetwork (Fig. 1a) had an average spacing of 50 km. The 40 northernmost sites were equipped with National Center for Atmospheric Research (NCAR) Portable Automated Mesonetwork (PAM II) stations. The remaining sites were furnished with National Severe Storm Laboratory (NSSL) Surface Automated Mesonetwork (SAM) stations. Dry- and wet-bulb temperatures, station pressure, rainfall, and wind velocity were recorded. Five-minute averages of the SAM data were generated from 1-minute samples to conform with the PAM II observations.

In preliminary analyses surface pressures were reduced to sea level using the methodology outlined in the Smithsonian Tables (List 1966; Table 48). This procedure resulted in fields with a number of permanent features—presumably due to inaccurate site elevation determinations and drifts in sensor calibrations. Consequently, 24 h mean surface pressures were computed at all sites for 7 May. Perturbation (deviation) pressures were then determined by subtracting the 24 h mean station pressure from the observed station pressure.
provided by the National Weather Service (NWS) array (Fig. 1b). Two sites in southeastern Oklahoma (HET and SLH) did not operate on 6–7 May. Radiosondes were released at 3 h intervals during the growth and mature stages of MCS development and at 90 min intervals during storm decline. Special serial soundings were also available from NWS sites at Oklahoma City (OKC), Oklahoma and Amarillo (AMA), Texas. Balloon positions were recorded for sondes released at the supplemental sites and OKC, and were used to place precisely each measurement. Positions were further adjusted using a mean mesovortex motion to account for differences in sounding release times, even though the motion of the vortex and its flow structure were not particularly steady. The time window for each analysis is small with respect to the life cycle of the MCS; consequently, this procedure is thought to have little influence on the conclusions drawn from the data.

Divergence, vertical velocity, and relative vertical vorticity were computed from the spatially adjusted rawinsonde observations via the kinematic method (Bellamy 1949). The wind is assumed to vary linearly along the legs of triangles formed by the rawinsonde observations. An O'Brien (1970) correction was applied to ensure that the vertical velocity was zero at the uppermost data level (usually 150 mb, ~14 km) and at the ground.

Composite radar precipitation maps were prepared with radar reflectivity measurements from NWS stations at OKC, AMA and Wichita, Kansas (near IAB, McConnell Air Force Base). Doppler radar observations from NSSL's radars at Norman (NRO), Oklahoma and Page Field (CIM), near Yukon, Oklahoma (Fig. 1a) were used to monitor the movement of the mesovortex and to reconstruct the three-dimensional flow within the MCS. The dual-Doppler analysis procedure is based on the methodology described by Brandes (1977) but incorporates downward integration of the continuity equation and a variational adjustment that forces the vertical velocity to be zero at ground and at the uppermost data level (Ray et al. 1980; Ziegler et al. 1983). Single-Doppler observations from the NRO radar were used to determine vertical profiles of the mean divergence, mean vertical velocity, and mean horizontal velocity. The methodology, similar to that described by Browning and Wexler (1968), assumes that the wind varies linearly over the computational domain (a circle of 25 km radius). Other data examined in the study include IR (infrared) images from the GOES-West satellite and measurements obtained with a radar wind profiler near Liberal, Kansas (LBL).

3. Environment and formation (6 May 1800 CST to 7 May 0000 CST)

Evolution of the 6–7 May 1985 mesoscale convective system closely follows the typical midlatitude North American situation described by Maddox (1980) and
by Velasco and Fritsch (1987). An organizational stage is defined as the period 2100–0000 when a short squall line developed in the Texas panhandle and moved rapidly eastward sweeping away scattered, moderate-to-severe thunderstorms that had persisted from early afternoon hours. (All times are Central Standard Time; all heights are above mean sea level.) The preexisting activity had resulted in numerous reports of large hail in northwestern Oklahoma and the Texas panhandle. A tornado had touched down briefly near Woodward (WWR), Oklahoma. (The location of all sites mentioned in this report are shown in Fig. 1.)

a. Genesis region

Synoptic scale conditions at 1800 on 6 May 1985 are similar to those described by Maddox (1983) for a composite of MCC genesis regions. Environmental features include a weak frontal zone, low-level warm-air advection, an abundant supply of moisture transported by a low-level jet, and a lower-tropospheric short-wave trough.

The front (Fig. 2a) crosses the northern Texas panhandle and passes south of WWR and over Enid (END), Oklahoma. Surface temperatures and dewpoint temperatures in the MCS genesis region (the eastern Texas panhandle and northwestern Oklahoma) are 23°–29°C and 15°–20°C, respectively. A 1017 mb surface high pressure center over northern Missouri and 1006 mb low in the extreme southeastern corner of New Mexico create a weak pressure gradient across the genesis area. Surface winds are light (<6 m s⁻¹) and from a southeasterly (east-northeasterly) direction to the south (north) of the front.

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**Fig. 2.** Surface (a) and upper-air analysis (b)–(d) of horizontal winds in ground-relative (absolute) reference frame and temperature (dashed lines) at 1800 CST on 6 May 1985. A half wind barb is 2.5 m s⁻¹, a full barb is 5 m s⁻¹, and a flag (when present) is 25 m s⁻¹. Pressure contours (continuous lines, panel (a)) are reduced to sea level and are at 4 mb intervals. Height contours (continuous lines) are at 15 m intervals in panels (b) and (c) and at 30 m intervals in panel (d). Isotherms are at 5°C intervals. Hatched regions on panels (a), (b) and (c) represent dewpoint temperatures greater than 15, 10, and 0°C, respectively.
Standard pressure surface analyses, redrawn with the PRE-STORM supplemental soundings added, are presented in Figs. 2b–d. The 850 mb map for 1800 (Fig. 2b) shows a sharp trough across west-central Texas, northeastern New Mexico, and central Colorado. Warmest temperatures are over the high terrain to the rear of the trough. Warm-air advection in the Texas panhandle and extreme western Oklahoma contribute to the destabilization of the air mass in the MCS genesis region. Dewpoint temperatures >10°C at 850 mb delineate a surface-based moist tongue that covers central Texas, much of Oklahoma, and western Kansas. Peak dewpoint temperatures (>14°C) are aligned from OKC to Dodge City (DDC), Kansas. The top of the moist layer increases from slightly above 850 mb at Stephenville (SEP), Texas to more than 600 mb in the vicinity of the front. (The front is indicated by the wind shift and confluent zone at the Oklahoma–Kansas border.) Lifted indices along the moist air axis are −3° to −8°C. As the MCS developed, the axis of highest dewpoint temperatures and deepest surface moisture moved westward to a position slightly ahead of the squall line.

The 850 mb winds at IAB, OKC, WWR, END, and Clinton (CSM), Oklahoma imply that a low-level jet lies across central Oklahoma and central Kansas. With the jet strengthened, particularly in the region ahead of the squall line.

The 700 mb trough (Fig. 2c) is reduced in amplitude from that at 850 mb and is difficult to detect in western Texas. Wind flow over the region of subsequent MCS genesis and growth is diffusive. Velocities have veered with height and are weaker than at 850 mb. The veering wind is consistent with the warm-air advection that extends to 700 mb. High dewpoint temperatures (>0°C) over northern Oklahoma and southern Kansas straddle the front and mark the region of deepest surface-based moisture. At 500 mb (Fig. 2d), a weak shortwave trough can be seen near the eastern boundaries of New Mexico and Colorado; a weak ridge exists over Oklahoma and Kansas.

b. Early development

By 2100 thunderstorms in the western Texas panhandle began to align in a short squall line that moved rapidly eastward. At 0000 (Fig. 3a) the line extends northeastward from 50 km east-northeast of AMA nearly to the Oklahoma border and then turns northward. Convection intensity decreases from strong thunderstorms in the southwest to weak showers in the north. A broad area of stratiform rain, bounded on the east by the squall line, spreads northward into Kansas. The preexisting thunderstorm activity, along and to the north of the front, decreases in intensity from a large hail storm to the south of WWR to light showers in southeastern Kansas. IR satellite images show that the cloud canopy had expanded rapidly between 2100 and 0000. At this early stage, coldest temperatures, indicating highest cloud tops, are displaced slightly behind the north end of the squall line (Fig. 4a). A small, second area of cold cloud-top temperatures, 50 to 60 km farther to the east, is associated with the severe thunderstorm south of WWR. Warmer temperatures in Kansas indicate that these storms are less developed.

There is evidence that a mesovortex was already present at 0000. Figure 5 shows a time versus height section of mesovortex-relative winds recorded by the radar wind profiler near Liberal, Kansas (LBL). The observations hint that a closed cyclonic wind system in the lower troposphere (below 4 km) passes to the south of the site between 2300 and 0030. Unfortunately, the available data do not establish whether the vortex is related to the preexisting shortwave or represents new development. Based on radar observations (0200 to 0700), the mesovortex and squall line moved from 275° at 8.3 m s⁻¹ and from 287° at 19 m s⁻¹, respectively.

c. Deepening of the moist layer and intensification of the low-level jet

Changes in the environment ahead of the squall line are recorded in soundings from CSM (Figs. 6 and 7). At 1851 (Fig. 6a), the height of the surface-based moist layer reaches 2.6 km. The sounding then turns dry to 8 km whereupon a relatively moist layer of convective outflow is found. From 1851 to 2106, dewpoint temperatures at all levels below 5.7 km increase markedly and dry-bulb temperatures at nearly all levels between 2.6 and 8 km cool (cf., Figs. 6a and 6b). Established trends continue until 0007. This sounding (Fig. 6c) shows additional cooling between 4 and 7 km and further moistening in upper regions of the surface-based moisture layer (near 4.4 km). The considerable potential instability present was released in new convection that first appeared on radar at 0030 when a short eastward protruding line of thundershowers developed ahead of the primary squall line. This activity can be seen (at a later time) in Fig. 3b as it is overtaken by the faster moving squall line.

The CSM wind profile for 1851 (Fig. 7) reveals a deep, low-level wind maximum of 15–17 m s⁻¹ in the vortex-reference frame. The wind maximum sharpens to a jet of 20 m s⁻¹ (1.2 km elevation) at 2106. Between 2106 and 0007, as the squall line moves within 80 km of CSM, a further intensification to 25 m s⁻¹ occurs. Although the height of the wind maximum is relatively high, the 2106 profile resembles onset conditions of low-level nocturnal jets commonly observed in the central United States (Bonner 1968). The intensification of this flow between 1851 and 0007 is twice the normal nocturnal jet increase during these hours; and the strong veering associated with the evolution of a typical low-level jet is not apparent here (see Bonner’s Figs. 18 and 19). A similar speed increase occurred between 2330 and 0230 as the MCS approached OKC. The jet focuses the low-level moisture transport and
serves as a persistent source of high equivalent potential temperature ($\theta_e$) air. The jet seems analogous to the “warm conveyor belt” described by Browning (1971) for synoptic waves.

4. MCS evolution

a. Mature stage (0200–0300 CST)

The 6–7 May mesoscale convective system never achieved a steady state condition; storm decline began immediately when growth ended. Consequently, a mature stage is defined by the maximum extent of the stratiform rain area and the overlying cloud canopy. Areal coverage of stratiform rain was greatest between 0200 (Fig. 3b) and 0300, whereas the coverage of cloud top temperatures ≤−52°C was slightly greater at 0200 (Fig. 4b) than at 0300. After 0300, the stratiform rain area began to erode on its western boundary; and a steady increase (warming) in IR temperatures indicated a decrease in the cloud canopy height.
1) WIND FLOW AND THERMODYNAMIC STRUCTURE

Low-level mesovortex-relative winds, determined from the special rawinsonde observations, turn cyclonically across the entire MCS (Fig. 8a). Inflow is from the south-southeast at 10–12 m s⁻¹ but is as strong as 23 m s⁻¹ at lower levels (0.75 km, OKC). Strong winds also exist in the diffluent flow within the stratiform rain region. The wind field does not explicitly portray a vortex; placement of the mesovortex center is based upon Doppler radar measurements. A vortex appears at 3 km (Fig. 8b) but not at 6 km (Fig. 8c). At both 3 and 6 km a broad region of confluent winds feeds into a rear downdraft from the west.

Particularly striking in Figs. 8a and 8b are the warm-air advection ahead of the mesovortex and the cold-air advection at the rear. Temperature and wind patterns bear close resemblance to those found by Carr and Millard (1985) for comma cloud systems. Processes important on the synoptic scale, such as vorticity and temperature advection, may be operating here (in
Fig. 5. Time versus height cross section of storm relative winds recorded by profiler near Liberal (LBL), Kansas on 6–7 May 1985. Wind bars are scaled as in Fig. 2. Heights are above mean sea level (MSL).

addition to evaporative cooling and latent heat release) and may also influence the distribution of vertical velocity in MCS’s.

Soundings released at the rear of the stratiform rain region show considerable drying at low levels. The CSM sounding (0307, Fig. 6d) exhibits an “onion” shape below 4.4 km that is much like the sounding described by Zipser (1977) for the trailing stratiform region of a tropical squall line. The nearly dry adiabatic lapse rate and constant water vapor mixing ratio between 2.6 and 4.4 km are consistent with a kinematic analysis, yet to be described, that shows this air is sinking. Although the subsiding air may have warmed dry adiabatically, sensible temperatures at nearly all levels below 3.9 km are either the same or slightly cooler than those observed in the prestorm environment (cf., Figs. 6c and 6d).

Subsidence warming is seen below 3.1 km in the WWR sounding for 0300 (Fig. 9). Temperatures in the highly unsaturated layer are much cooler than observed in neighboring regions or in the prestorm environment (Figs. 8a and 8b).

The stratification of equivalent potential temperatures provides insight concerning the origin of air parcels within the MCS. In the lower dry layer of the CSM sounding (Fig. 6d) equivalent potential temperatures range from 322.6 K (1.5 km) to 324.7 K (3 km). Temperatures colder than 325 K were found in upwind sections of the rear inflow, e.g., between 3.6 and 7 km (a minimum of 319 K, 4.9 km) at AMA (0200), but also in the prestorm environment, e.g., between 4.5 and 7.3 km in an earlier sounding at CSM (a minimum of 317 K, 5.7 km; Fig. 6c). Streamlines computed with a gridded version of the winds in Fig. 8 (not shown) indicate that the air in the lower portion of the unsaturated downdraft (near 1.5 km) originated ahead of the MCS while the air at ∼3 km and higher came from the west. Air parcels within the entire low-level unsaturated layer of the WWR sounding originate in front of the mesovortex. In situ equivalent potential temperatures of 320.4 to 321.7 K again indicate descent from midlevels.

Winds above 7.5 km in the CSM 0307 sounding have a strong southerly component. The flow is quite moist, nearly saturated with respect to ice, and has equivalent potential temperatures ranging from 329 K (7.5 km) to more than 332 K (above 9.5 km). This air is thought to originate in the lower troposphere, ascend in convective updrafts along the squall line, and form the deep convective outflow layer in the upper troposphere.1 An undiluted ascent would have produced outflow temperatures of 335 to 340 K. In this

1 The appellation “convective outflow” as used here identifies the source of the air. Far more than a passive outgrowth of the convective zone, this southerly current feeds the mesoscale updraft and may be the most significant feature of an MCS (Maddox 1980).
region of the mesoscale updraft at least, some mixing in the convective updrafts, presumably with air originating above 2.3 km, has taken place. Nonetheless, the observed upper tropospheric temperatures are several degrees warmer than the environment and constitute the warm core that drives the mesoscale updraft. Winds recorded with the LBL wind profiler (Fig. 5) reveal that the upper-level outflow was deepest when the mesovortex passed to the south of the site (at ~0000) and the lower boundary of the southerly current descended to 8 km. The convective outflow is fairly uniform in direction and speed. Sublimation of precipitation debris falling from the outflow layer and resultant cooling are thought to accelerate downdrafts in the dry and potentially cold rear inflow (e.g., the air at 6.4 km in the CSM sounding with $\theta_e = 323$ K and a sensible temperature of $-17^\circ$C).

A surface perturbation pressure analysis for 0300 is presented in Fig. 10. Two low pressure areas exist ahead of the squall line. The minimum in north-central Oklahoma ($<-2$ mb) is a transient feature that was first detected near WWR at 0000 (Fig. 11). This low moved southeastward initially, passed east of the convection between 0030 and 0130, and then turned northeastward. Its role in the evolution of the 6–7 May MCS is unknown. The inverted pressure trough of
gument has been made that the highly unsaturated portion of the WWR sounding is composed of air that begins at the front of the system. Although the CSM sounding exhibits rear-to-front flow at elevations $\geq 3$ km, this sounding is actually between the thunderstorm high and the wake low. The only obvious warming at WWR and CSM is associated with convective outflow in the upper troposphere. On the other hand, the 29 m s$^{-1}$ translation speed of the wake low, more than 20 m s$^{-1}$ faster than the mesovortex, approaches that of the fastest ground-relative wind in the rear-inflow jet.

2) Kinematic analysis

A kinematic analysis of MCS flow properties near the mesovortex center was performed with rawinsonde observations from END, WWR, and CSM (Fig. 12a). The shallow, ground-based convergent layer ($\delta < 0$), which appears in nearly all triangles, is capped by the subsidence inversion and is apparently related to the general lowering of surface pressure in the vicinity of the MCS (Rutledge et al. 1988). The mean flow is divergent between 0.7 and 2 km, convergent from 2 to 8 km, and divergent above 9 km. The mesoscale downdraft ($\omega > 0$) resides between 1 and 4.5 km and has a peak velocity of $10 \times 10^{-3}$ mb s$^{-1}$. Hence, the kinematic analysis supports the earlier notion that the low-level unsaturated layers at WWR and CSM were created by subsidence. The mesoscale updraft begins above 4.5 km and has a peak velocity of $-17 \times 10^{-3}$ mb s$^{-1}$. The updraft slows and the flow diverges just below the tropopause.

The CSM sounding (Fig. 6d) shows that the upper boundary of the intruding westerly current is 7.5 km, whereas the kinematic analysis puts the upper bound of the mesoscale downdraft at 4.5 km. This suggests that the depth of the downdraft is not uniform and deepens toward the west. The maximum downdraft ($10 \times 10^{-2}$ mb s$^{-1}, 2.5$ km) resides roughly midway between the freezing level and the base of the subsidence layers at CSM and WWR. The melting of precipitation may have accelerated the downdraft (Leary and Houze 1979).

Figure 12a shows that vertical relative vorticity ($\vec{\zeta}$) is weakly negative in the stable surface layer but becomes strongly positive from 1 to 7 km, i.e., at levels principally dominated by strong rear inflow. The maximum at 6.6 km responds to a southwesterly jet of 39 m s$^{-1}$ at CSM that lies between the dry layer centered near 6 km and the upper-tropospheric moist layer (Fig. 6d). Vorticity decreases markedly in the upper troposphere (above 7 km), where the upper-level convective outflow associated with the mesoscale updraft prevails. Apparently little vertical vorticity associated with the strong rear inflow is transported upward by the mesoscale updraft. Except at the interface, there is probably little mixing between the two currents. The potentially cold rear-inflow air would tend to sink when chilled by the evaporation of either precipitation or
Fig. 8. Temperatures, dewpoint temperatures, and winds aloft in mesovortex reference frame for 0300 CST on 7 May 1985. Heights are (a) 1.5, (b) 3, and (c) 6 km above mean sea level (MSL). Temperatures and winds are plotted as in Fig. 2. Isotherms (dashed lines) are at 1°C intervals. The data window is 0230–0330 CST except for the Amarillo (AMA), Texas sounding, which is 0200 CST.

cloud. Above 8 km the flow turns divergent and the vorticity is anticyclonic.

Although approximately an order of magnitude larger, the profiles in Fig. 12a are remarkably similar to those determined from synoptic scale data by Bosart (1986) for the stratiform region of the convective system that produced the Johnstown, Pennsylvania flood of 19–20 July 1977. The vertical velocity structure also agrees with that found for a tropical squall line by Gamage and Houze (1982, their Fig. 13) and with that reported for a midlatitude squall line by Smull and Houze (1987b, their Fig. 12).

A kinematic analysis with the sounding triad IAB-END-WWR (Fig. 12b) reveals that a weak updraft has replaced the mesoscale downdraft in northern quadrants of the MCS. This finding is consistent with the warm-air advection seen in Figs. 8a and 8b. The deep updraft layer may account for the persistent heavy precipitation in the northern quadrant of the MCS (e.g., Fig. 3b). Updraft strengths, however, become suspi-
ciously large above 4 km, a result attributed to the obtuse shape of the computational triangle at midtropospheric levels (see Fig. 8c).

b. Early decline (0430–0505 CST)

1) WIND AND THERMODYNAMIC PROPERTIES

By 0430 the MCS has moved into central Oklahoma and the mesovortex center is very close to END (Fig.

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Fig. 10. Surface pressure perturbations from 24 hour mean pressures at 0300 CST on 7 May 1985. The contour interval is 0.5 mb. Radar-indicated precipitation areas are hatched.

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Fig. 11. Tracks of wake-low, thunderstorm-high, and transient-low pressure systems on 7 May 1985. Positions are given at 30 min intervals. The average speeds of the thunderstorm high and the wake low are given.
Mesovortex flow has decreased in speed (Figs. 13a and 13b), but the closed wind circulation reaches from ground to above 6 km (Fig. 13c). Strongest winds reside in the rear-to-front jet that begins in the confluent region at the rear of the MCS and passes to the south of the vortex center. Individual soundings reveal that in the west the intruding flow extends above 9 km.

The isotherm pattern at 1.5 km continues to show that the low-level mesovortex core is relatively cold with respect to its local environment. Consequently, cold-air advection continues in western quadrants of the MCS, especially from Pratt (PTT), Kansas, to WWR, and to OKC; however, pronounced warming now accompanies the drying at the rear of the precipitation shield, e.g., near WWR and OKC. Lowest surface pressures (Fig. 11) are close to the vortex center at 1.5 km and not with the warmest temperatures. The vortex core is also cold at 3 km, but cold advection is reduced from that at 1.5 km and from that observed at 3 km earlier (0300, Fig. 8b). The vortex core is warm above 5 km (Fig. 13c).

An OKC sounding (Fig. 14) depicts a low-level unsaturated layer that has an almost dry adiabatic lapse rate from 1.5 km to slightly above the freezing level. Equivalent potential temperatures in the layer are 2°C warmer than found earlier at this relative location in the MCS (0300, Figs. 6d, 9). The OKC sounding shows a deep layer of westerly winds, the rear inflow, from below 1.5 to above 6 km. Precipitation falling into the intruding current from the overlying convective outflow is swept rapidly eastward as it evaporates, giving a comma-like appearance to the low-elevation reflectivity field (e.g., Fig. 3c).

Warm temperatures at WWR (Fig. 13a) associate with air that begins ahead of the storm. Equivalent potential temperatures have risen 5°C. Either the flow now originates from a different elevation or the environment has changed.

The mesovortex was positioned for dual-Doppler analysis only for a short time. A reconstruction of the horizontal wind flow during this period (0500) is shown in Fig. 15. The vortex center at 2.4 km ($x = -30, y = 102$ km) is near the tip of a hooklike region of enhanced stratiform rain > 30 dBZ. The circulation center coincides with a radar-indicated vertical vorticity maximum of $15 \times 10^{-4}$ s$^{-1}$. Strongest horizontal winds are displaced from the vortex center and are configured in a belt that approaches from the west (near $x = -55, y = 105$ km) and then passes within 50 km to the south and east of the center. The OKC sounding (Fig. 14) was released just south of the analysis domain ($x = -12, y = 18$ km).

2) VAD AND KINEMATIC ANALYSES

Vertical profiles of mean divergence and mean vertical and horizontal velocity, determined by VAD technique (Browning and Wexler 1968) with measurements from the NRO radar at a radius of 25 km, are given in Fig. 16. The data, from the rear of the stratiform rain shield, clearly shows that the intruding westerly flow is subsiding. In the mean, the wind is divergent from ground to 4 km, and convergent from 4 to 7.5 km. The corresponding vertical motion shows a deep downdraft from ground to 7 km and a shallow updraft layer above. Strong convergence at midtropo-
Fig. 13. Temperatures, dewpoint temperatures, and vortex relative winds as in Fig. 8, except for 0430 CST. The data window is 0426-0505 CST.

Mesovortex properties are now examined by kinematic analysis using the OKC–WWR–IAB rawinsonde triad (Fig. 17). Release times are not synchronous; hence, results must be interpreted cautiously. Salient features observed earlier (Fig. 12a) reappear in this 0440 analysis—but at different altitudes and decreased intensities. The latter is thought to be due in part to a larger computational domain (Fig. 1b). The considerable depth of the mean downdraft layer, which may correspond to the transition from rear inflow (west-southwest winds) to convective outflow in the mesoscale updraft (south-southwest winds). This relationship is similar to that in the 10–11 June MCS (Rutledge et al. 1988; their Figs. 9 and 11). A large portion of the rear-inflow air must enter the mesoscale downdraft. Strongest downdrafts (−35 cm s⁻¹, 27 × 10⁻³ mb s⁻¹) are close to the freezing level (3.7 km, Fig. 14) and possibly have been accelerated by the melting of precipitation. Although an areal average, the horizontal wind profile (Fig. 16) agrees quite well with the OKC sounding (Fig. 14) released within the analysis domain.
Fig. 14. As in Fig. 6, except for the Oklahoma City (OKC), Oklahoma sounding at 0505 CST on 7 May 1985.

Fig. 15. Mesovortex horizontal flow reconstructed from dual-Doppler radar observations collected at 0500 CST on 7 May 1985. Winds are relative to the mesovortex. The height is 2.4 km. A vector 4 km in length is equivalent to 20 m s⁻¹. Distances are from the NRO radar.

Fig. 16. Vertical profiles of mean vertical velocity, divergence, and horizontal velocity in mesovortex reference frame, as determined by VAD analysis with data from the NRO radar at 0500 CST on 7 May 1985. Wind barbs are scaled as in Fig. 2. The computational radius was 25 km.
be overemphasized in the analysis, is related to the accelerated flow at OKC. Vertical vorticity continues to concentrate at levels with strong rear inflow (Fig. 14). Relative vorticity decreases rapidly in the convective outflow layer and is weak in the lower stratosphere. All three soundings indicate a reduced convective outflow in the upper troposphere.

c. Dissolution and epilogue

Surface perturbation pressures from 0600 (Fig. 18) show an intense low that is near the constricted portion of the precipitation field and is tied to the intruding westerly inflow. The radar-indicated mesovortex center is displaced 80 km to the northwest of the low-pressure center and lies within the decreasing area of stratiform rain in northern Oklahoma (also Fig. 3d). By 0700 the separation between the leading edge of the intruding flow and the vortex center had increased even farther, and two distinct low-pressure centers could be seen (Fig. 19). The most intense low persists near the tip of the intruding flow. Although a connection with subsidence warming cannot be established, the relationship between the low and the rear inflow appears similar to that in the 10–11 June storm (Johnson and Hamilton 1988). The trailing weak low may be related to the mesovortex core or to the warming seen at this relative location earlier (e.g., Fig. 13a).

The satellite IR image for 0700 (Fig. 4d) shows that a comma shaped region of cloud top temperatures $<-32^\circ C$ was beginning to evolve, presumably as the subsiding rear inflow developed upward. By 1000 there were two distinct regions of cloud top temperatures $<-32^\circ C$. Although the precipitation had largely dissipated, the mesovortex persevered. In fact, a spiral band cloud system was tracked for the next several days as it continued to move eastward.

5. Discussion and conclusions

Evolution of the 6–7 May 1985 storm typifies nocturnal mesoscale convective systems observed in the central United States. Genesis occurred along a stationary front in a region of preexisting thunderstorms that had been triggered initially by daytime heating and had persisted into the evening hours. By 2100 a squall line, which presumably developed in response to an advancing lower-tropospheric shortwave, began sweeping away the lingering convection. An extensive attached area of heavy stratiform rain and associated mesoscale updraft formed behind the fast-moving squall line. From early on, there were indications of an embedded meso-$\beta$-scale vortex within the stratiform area. Expansion of the cloud canopy and precipitation shield continued until 0200–0300 when convection along the squall line weakened and a slow decline began. Both the squall line and stratiform rain area had largely dissipated by 1000 on 7 May.

Conspicuous low- and midlevel MCS features, at maximum development, are summarized in Fig. 20. In northern and eastern sections of the mesovortex, weak ascent averaging $-10 \times 10^{-3}$ mb s$^{-1}$ existed below 3 km. The updraft region extended upward as the
ward, while pressure forces associated with hydrostatic effects in the mesoscale updraft accelerate the flow forward into the precipitation region. Additional middlelevel air converges to replace that which subsides in the downdraft. This air also chills and descends—a continuing process until the mass of precipitation falling from the convective outflow no longer provides sufficient cooling to overcome the dry-adiabatic warming.

Kinematic analyses show that MCS vertical vorticity was intense at midstorm levels where strong rear inflow and convergence existed; consequently, stretching may be the dominant vorticity-generating mechanism. A midlevel convergence ($\delta$) of $-10 \times 10^{-7}$ s$^{-1}$ (an average value in Fig. 12a) and an earth rotation vorticity ($\zeta_R$) of $8.6 \times 10^{-5}$ s$^{-1}$ (at 36° latitude), combine to amplify vertical vorticity at a stretching rate ($-\zeta_R$) of 8.6 $\times 10^{-9}$ s$^{-2}$. A mesovortex having a mean relative vertical vorticity of $15 \times 10^{-5}$ s$^{-1}$ could be produced in 1.5 h. Preexisting relative vorticity, such as that of the shortwave trough, would only accelerate the process. [The maximum stretching rate in Fig. 12a is $30 \times 10^{-9}$ s$^{-2}$ (4.4 km elevation).] Short spin-up times for mesovortices indicate that they are likely to be common in MCS's—provided that the system has persisted for several hours, that the mixing does not diminish the latent heat released or diffuse the vorticity, and that the system is not sheared apart by the larger scale flow.

Note that the strong midlevel vorticity arises largely from convergence into the mesoscale downdraft rather than convergence into the mesoscale updraft. This conclusion is based on the observation that the rear inflow has low equivalent potential temperature and hence sinks when chilled by evaporation.

Strong vertical vorticity also exists in regions of the MCS dominated by the mesoscale downdraft. The pri-
mary source of the low-level vorticity may originate with horizontal vorticity generated by baroclinicity in the rear inflow. To the west and south of the mesovortex center (Figs. 8b and 13b) warm air lies to the right and cold air lies to the left when facing downwind. A potential temperature gradient $\nabla T = 2 \, K$ over a horizontal distance $\Delta x_h = 100 \, km$ (a conservative value) coupled with a mean potential temperature $\theta_0 = 310 \, K$ and a gravitational acceleration $g = 9.8 \, m \, s^{-2}$ relates to a horizontal vorticity generation rate, $g\nabla T / (\theta_0 \Delta x_h) \approx 6 \times 10^{-7} \, s^{-1}$. This vorticity generation is manifest as the large vertical wind shear over CSM ($7.5 \times 10^{-3} \, s^{-1}$ at 3 km, Fig. 6d). The associated horizontal vorticity vector points to the right of the temperature gradient and is very nearly opposite the velocity vector. The streamwise vorticity component ($\eta_x$) is $7 \times 10^{-3} \, s^{-1}$. The observations presented (Fig. 16) indicate that the speed of the downdraft is $\sim 0.4 \, m \, s^{-1}$. If the streamwise vorticity associated with the rear inflow experiences a downwind vertical velocity gradient $\Delta \omega / \Delta x = 5 \times 10^{-6} \, s^{-1}$ (0.5 $m^2$ per 100 km) as it approaches the center of the mesoscale downdraft, the vertical vorticity could be generated by the tilting of the horizontal vorticity at a rate of $\eta_z \Delta \omega / \Delta x = 3.5 \times 10^{-9} \, s^{-2}$. This rate is comparable to the largest stretching amplification rate in Fig. 12a. Hence, the generation of horizontal vorticity by temperature gradients and the subsequent tilting of that vorticity (an effect that is not quasi-geostrophic) could play an important role in the distribution of vorticity in this MCS. Heuristic studies are planned to better quantify all mechanisms whereby vertical vorticity is amplified.

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