Frontal-Wave–like Evolution in Some Mesoscale Convective Complexes

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ABSTRACT

In some mesoscale convective complexes (MCCs) the convective regions assume a configuration that resembles a frontal wave, while in other systems the configuration is not classifiable. We examined the evolution of the internal structure and flow of two MCCs that were part of an episode of systems that propagated along a stationary front, and in which the convective activity evolved from a chaotic to a frontal-wave–like pattern. The development of mesoscale vorticity and its possible relationship to the spatial patterns are explored.

Each system developed two dissimilar convective bands that together formed an open-wave pattern. The greatest number of convective clusters and the greatest midlevel convergence and high-level divergence were found in a core region (the “apex” of the wave) where the bands intersected. The north–south convective line was less enduring than the core convection or the east–west band; it evolved from outflow boundaries of the early storms preceding the mesosystem.

A conceptual model of a frontal-wave–like MCC is developed from Doppler radar and rawinsonde observations. Three airstreams are included in the model: a relatively warm and saturated, ascending flow from the apex into the rear half of the cloud shield; a dry midlevel inflow into the southern flank of the system, in which both ascending and descending motions are observed; and cool inflow into the northern flank of the stratiform cloud below 6 km.

It is found that the frontal-wave configuration is apparently not related to the development of vorticity on the scale of the mesoscale cloud system. Rather, the north–south convective line may evolve ahead of the outflow of the early storms as it propagates into the most unstable air.

It is hypothesized that the unidirectional profile of vertical shear led to an asymmetric distribution of stratiform cloud north of the convective core region. The asymmetry may have suppressed a stronger, more prolonged convergent wind, which inhibited a stronger response of the rotational wind.

1. Introduction

Ever since radar and satellite platforms provided a mesoscale perspective of convective storm systems and their setting, it was observed that cumulonimbus cells may develop in a variety of spatial patterns, yet generate a similar precipitating cloud system, the mesoscale convective complex (MCC) described by Maddox (1980) or the more general mesoscale convective system (MCS). One pattern has been recognized since the early days of radar: the squall line, described in the Glossary of Meteorology (Huschke 1959) as a band of active thunderstorms usually hundreds of kilometers long and 15–80 km wide. However, organized nonlinear patterns of convective cells, and even apparently unorganized patterns, also generate precipitating mesoscale convective systems. Some of these are suggestive of a rotating cloud system, and comma-shaped, frontal-wave–shaped, and spiral-shaped bands of convective cloud have all been observed in radar and satellite data.

Johnston (1982) reported comma-shaped clouds and apparent cyclonic motion in satellite images of dissipating mesoscale cloud systems. Spiral-like bands of enhanced radar echo were detected by Leary and Rappaport (1987) in the stratiform cloud of one MCS in apparent association with cyclonic flow in middle levels. Other studies classified mesoscale convective systems according to the dominant mode of organization. Clark et al. (1980) described two subtypes of MCCs that were not squall lines and were in a dynamically weak environment. Blanchard (1990) classified one-third of the 25 MCSs observed during the Preliminary Regional Experiment for STORM-Central (PRE-STORM; see Cunning 1986) as exhibiting either an occluding pattern or a chaotic pattern of convective centers within stratiform echo. Bluestein and Jain (1985) and Bluestein et al. (1987) classified springtime squall lines in Oklahoma into four modes of convective development, two of which (the “broken areal” and the “embedded areal” types) resemble the nonlinear MCC prototypes in this study. Six years of springtime...
rainstorms in Oklahoma were classified by Houze et al. (1990) on a continuum from "strongly" to "weakly" classifiable, or "unclassifiable," according to their resemblance to a prototype of a leading convective line with trailing stratiform echo. Fully one-third of their rain systems were deemed unclassifiable. At one point these authors employed the term "chaotic" to describe the organization.

It was expected that the mesoscale lower-level inflow and higher-level outflow of MCCs would adjust to gradient-wind balance (Cotton et al. 1989) and spin up an inertially stable warm-core cyclone (Zhang and Fritsch 1988). Cyclonic vorticity had been detected in the midlevels of tropical mesoscale systems by Leary (1979), Fortune (1980), and Gamache and Houze (1982). Evidence for mesoscale vortices in midlatitude MCSs has been reported by Johnston (1982), Ogura and Liou (1980), Bosart and Sanders (1981), Smull and Houze (1985), Leary and Rappaport (1987), Johnson and Hamilton (1988), Houze et al. (1989), Brandes (1990), and Bartels and Maddox (1991).

Maddox (1983) and Cotton et al. (1989) reported only a lackluster evolution of positive vorticity from the developing stage to the decay stages of the MCCs. Their results are composited from many events and from the standard rawinsonde network, which lacked the resolution necessary to clearly separate the mesoscale vorticity from the larger-scale anticyclonic environment. However, the mesoscale convective vortex has been explicitly simulated by Zhang and Fritsch (1987, 1988) and Tripoli and Cotton (1989a,b). A mesoscale vortex that is clearly evident in radar and satellite imagery developed in a weak dynamic environment but with frontal forcing in low levels (Menard and Fritsch 1989). A simulation by Zhang and Fritsch (1988) reproduced the rainfall pattern, the lack of movement, and other features of this MCS.

It might be expected that a mesoscale convective vortex would manifest circular symmetry or patterns of rotation in its clouds. Spiral and comma-shaped cloud bands have been reported in the postmature stages in several of the aforementioned studies. Less often, the convective elements were reported to develop in patterns reminiscent of a synoptic-scale cyclone—specifically, along bands that curved into a center just like warm fronts and cold fronts curve into a frontal wave cyclone. Some MCCs have been described as occluding systems in analogy to the later stages of an extratropical cyclone when a dry airstream intrudes into the core of the cyclone.

This paper examines the evolution and characteristics of two mesoscale convective complexes that originated with a chaotic distribution of convective cells and clusters of all sizes, yet later developed a wave-cyclone–like pattern of convective cloud embedded in a mesoscale stratiform cloud. We analyze the three-dimensional flows that were observed during this reorganization, and generalize the properties into a prototype of an MCC with frontal-wave–like organization. In another paper we compare this prototype with MCCs having leading line–trailing stratiform structure, and with systems having a mesoscale convective vortex.

2. Data sources and procedures

The Oklahoma–Kansas PRE-STORM field program was undertaken in 1985 to investigate the structure and dynamics of mesoscale convective weather systems (Cunning 1986). Fifteen MCSs passed through the observing network during the field program; of these, four were chosen for study because they exhibited some evolution from a chaotic to a more organized distribution of radar echo. Three of these propagated over a stationary frontal zone in quick succession on 3 and 4 June 1985; they are hereafter called systems A, B, and C. The fourth developed a comma-shaped cloud shield and a mesoscale vortex on 7 May 1985 and has been discussed by Brandes (1990).

Six radars of the National Weather Service were equipped with digitizing recorders in the radar data processor (RADAP-II) program. The digitized radar data were used extensively; the lowest elevation angle (the base scan) was nearly always used. Data were recorded at range spacing of 1.9 km and at increments of 2° in azimuth and elevation.

The National Center for Atmospheric Research (NCAR) operated a pair of 5-cm Doppler radars (CP-3 and CP-4) northwest of Wichita. For system C, the reflectivity data were analyzed every 20 min, and dual-Doppler kinematic analyses were done for seven selected times spanning the mature and early decaying stages (the definitions of life-cycle stages of Cotton et al. 1989 were used). In this paper the kinematic analysis for one time (1042 UTC 4 June1) is discussed in detail; the evolution of the kinematic fields throughout the 3-h period spanning all seven analyses is given in a companion paper (McAnelly et al. 1992). The derivation of the horizontal and vertical wind fields from the dual Doppler data is described in the companion paper.

The National Weather Service (NWS) supplied visible and enhanced infrared images from the Geostationary Operational Environmental Satellite (GOES) every one-half hour. Data on cloud-top area and cloud-top temperature of MCCs were provided by J. Augustine, who has conducted an annual survey of MCCs; the survey of MCCs in 1985 included a discussion of system B on 4 June 1985 (Augustine and Howard 1988).

Hourly surface observations at airports were merged with the special observations from the Portable Automated Mesonetwork (PAM) and the Surface Automated Mesonetwork (SAM) deployed for the field program. The pressure observations underwent three ad-

1 All times are universal time coordinated (UTC).
justments before they were analyzed. The pressures were adjusted to the mean elevation of all stations (480 m MSL) to remove the effect of elevation; atmospheric tidal effects were removed; and the estimated instrument bias of each station was removed with a procedure described in Stumpf (1988). The adjustment technique is the same employed by Stumpf et al. (1991) for system B; details are given in that paper. The wind, pressure, temperature, and moisture fields were subjectively analyzed and overlaid on precipitation echoes.

The operational NWS rawinsondes were supplemented by special soundings launched at 12 sites at intervals of, usually, every 90 min during the latter part of the life of system A and all of system B. The radar wind-profiling stations at Liberal and McPherson, Kansas, obtained 1-h and 0.5-h averages, respectively, of the horizontal winds from 1.8 to 11 km above ground level. Wind data from profilers and rawinsondes were displayed as time–height sections, hodographs, and layer averages of shear. Temperature profiles of the raw data from the special soundings contained numerous superadiabatic layers, some of which are believed to be spurious because they were observed just above saturated layers. It is known that a wet temperature sensor, on emerging from a cloud, can experience rapid cooling as water is evaporated from the sensor. These spurious effects were corrected by retaining the temperature at the bottom of the apparent superadiabatic layer, but eliminating successive observations above that, until a potential temperature was reached that would give a nonnegative value of $\partial \theta / \partial z$ in the layer. Profiles were then analyzed on skew $T$–$\log p$ diagrams. The potential buoyancy of lifted parcels was calculated in the manner of Weisman and Klemp (1986).

Profiler, sounding, aircraft, and Doppler radar data were combined with radar reflectivity on surface and constant-pressure-level charts. Analyses were done both manually and objectively with the General Meteorological Package (GEMPAK) software. The movements of several features associated with systems B and C on satellite and radar images and surface charts were analyzed: the center of the cold-cloud shield was tracked; clusters of convective echoes with at least some echo having reflectivity > 45 dBZ and surviving for at least 1.5 h were also tracked. Mesoscale centers of high and low surface pressure were followed, as were the convergence lines that defined the frontal wave pattern. After the propagation of the mesoscale systems was determined, system-relative winds were charted.

Two NOAA P-3 Orion aircraft obtained wind, temperature, moisture, and radar data from the near vicinity of system C. Finally, the occurrence of hail, high winds, flooding, and tornadoes was documented from the Storm Data publication of NOAA.

3. The synoptic-scale setting

a. Overview

Mesoscale convective systems often recur in episodes lasting several days, during which excessive rainfall can accumulate (Wetzel et al. 1983; McAnelly and Cotton 1986). The recurring rain systems often propagate along a nearly stationary front (Maddox 1983). Such was the case of the three systems A, B, and C that propagated through the PRE-STORM network on 3 and 4 June 1985.

A stationary front on which convective and stratiform precipitation was anchored for some 36 h extended from a low pressure center over the Mexican plateau and southwestern states, through Kansas and Oklahoma, to the Ohio River valley (Fig. 1). The low pressure in New Mexico was associated with a slow-moving upper-level cyclone over southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2). East-northeast winds converged with much warmer southerly winds in the southern California and Arizona (Fig. 2).

At 850 mb, the highest mixing ratios were localized in the diffusent southerly air entering a col or deformation zone associated with a frontal zone in southern Kansas, at 1200 UTC 3 June when the episode began (Fig. 3a). The streamline and temperature analyses indicate that the front was south of Dodge City (DDC) and Topeka (TOP) at this time and level. Twelve hours later, the second MCC (system B outlined in Fig. 3b)
was reaching maturity in central Kansas, at the leading edge of the warm and moist air entering the confluent axis of the frontal zone. Although the additional soundings provide subsynoptic-scale detail not present in Fig. 3a, the warming and wind shifts at DDC and TOP suggest that the frontal zone was displaced northward at this level. (At the surface the front was displaced less than 60 km.)

At 500 mb the MCCs developed in southwesterly flow entering the broad ridge downstream of the closed low in Fig. 2. The absolute-vorticity analysis (dashed) indicates a mesoscale short-wave ridge-trough couplet.
The vorticity field was evidently influenced by the MCC, as can be shown by alternately excluding and then including the dense set of special soundings (plotted in the inset) available at 0000 UTC. The analysis in Fig. 2b was based on the NWS operational soundings, in order to compare large-scale features with Fig. 2a. A second analysis that included the special sounding data showed greater values of vorticity on the southwestern margin and upstream of the MCC, but decreased values over the downstream half of the cloud shield. It will be shown later that the MCC induced a convergent wind response in its vicinity at this level.

b. Environmental stability

Because a front lay through the region that the MCCs traversed, and because low-level moisture increased from west to east while vertical shear decreased from north to south, we characterize the MCC environments with five soundings in different regions. The first storms of system A developed in south-central Kansas on the cool side of the front, and the system remained on the cool side. The storms generating system B started on the warm side of the front (southwest of Amarillo, Texas) but propagated to the cool side as the MCC consolidated. (For the early development of systems A and B, refer to Stumpf et al. 1991.) The sounding at Dodge City (DDC on Fig. 3a) at 1200 UTC 3 June, 1 h before and 100 km west of the first storm of system A, was chosen to represent one genesis region (Fig. 4a). Clearly, surface-based air could not support convection, even after attaining the observed afternoon temperatures of 23°C. If it is assumed that parcels originating in the boundary layer of the warm side of the front were lifted to their level of free convection (LFC) near Dodge City, then strong convection was possible. Parcels originating at the bottom and the top of this high-θe source layer are depicted by the dashed lines in Fig. 4a.

The layer that occupied a depth of 100 mb just above a shallow (~10-mb-thick) nocturnal inversion at Oklahoma City (OKC) had a value of θe (346 K) representative of several surface observations south of the front. The lifting was evidently accomplished by overrunning of the cooler air mass, and perhaps enhanced by upper-level diabatic influence over northern New Mexico and the Texas panhandle (Fig. 2a) and low-level warm advection (Fig. 3a). These conditions prevailed during the early phase of system A.

The dotted line represents parcels that may have been ingested in the first storms of system B, which developed after 1800 when surface temperatures were 30°C and dewpoints were 14°C–18°C. The assumed parcel mixing ratio of 12 g kg⁻¹ was observed in the lowest 130 mb of the DDC sounding and the lowest 40 mb of the Amarillo (AMA) sounding, which was closer to the genesis region of system B. The lifting condensation level (LCL) is found to be high, at 750 mb, but the parcel has the same value of θe as the cooler.
and moister parcels ingested into system A. The diurnal heating evidently would have accomplished most of the lifting to the LFC, and the parcel had to overcome a rather small negative area between the LCL and LFC. Perhaps this is why the early storms in system B were more widespread than those in system A (Stumpf et al. 1991).

The sounding at Woodward, Oklahoma (WWR in Fig. 3a), represents the preconvective environment of system C (Fig. 4b). The dashed curve represents a parcel starting with the mean conditions of the 800–860-mb layer (arrow), postulated to have been lifted over the front in southerly flow. The parcel is found to be negatively buoyant up to 700 mb though the negative area is small; above that level considerable positive buoyancy is evident. The first storms developed in a weak trough on the cool side of the front.

After the systems consolidated into MCCs, we describe the thermodynamic environment in three regions relative to the system. The three representative soundings [Enid, Oklahoma (END), Pratt, Kansas (PTT), and Russell, Kansas (RSL)] are located by the letters E, P, and R on the 850-mb map in Fig. 3b at 0000, when precipitation was falling at the sites. The
soundings, though, in Fig. 4c–e, were taken just ahead of the system at about 2100 UTC. All severe weather reports came from the region south of the surface front characterized by a deep layer of conditional instability. The midafternoon sounding at END, at 2110 UTC represented this regime (Fig. 4c). The lowest 100 mb had an average mixing ratio of 17 g kg⁻¹, and after mixing, a parcel from the boundary layer would have 3900 m² s⁻² of convective available potential energy (CAPE), the highest value found in this episode. There was little directional shear in this sounding, and no vertical shear at all from 900 to 700 mb. Three hours after this sounding a tornado passed through Enid as a multicellular convective line extended south from the MCC and propagated east with it.

The second region was the swath along which the central convective cores of the MCCs propagated. Surface air would not have supported convection, but the stable air was only 300 m deep beneath a frontal inversion. The PTT sounding in Fig. 4d, ahead of system B, exemplified the regime. Advection of moisture in southerly flow above the inversion brought the 880–750-mb layer to near saturation just before this sounding was observed. This elevated layer would support the release of a moderate amount of CAPE, 2300 m² s⁻² in the discrete convective clusters.

The third region represented the northern extent of the precipitation in the MCC, where prolonged rain fell from light thundershowers (TRW— in surface reports). Russell, Kansas (RSL), reflects conditions 3 h after the first MCC departed and 3 h before continuous rain fell from system B. The profile (Fig. 4e) exhibited several moist layers between 870 and 670 mb that would support buoyant updrafts upon modest lifting. The available buoyant energy, though, was much less than in the southern or central regions. The boundary layer below the inversion at 900 mb was 3° cooler than the overlying air, in easterly (ground-relative) flow.

Moderate values of vertical shear characterized the regions where the systems developed and matured. The mean shear in the 0.5–6-km layer was 1 and 2.2 m s⁻¹ km⁻¹ in the southern and central regimes, respectively. Another measure of shear, the vector difference of the 850-mb and 300-mb winds, varied little over this region, being 25–30 m s⁻¹ from 250° to 260°. This agreed quite well with the propagation of the cloud shields of all three MCCs, and agreed in direction but was 20%–50% faster than the movement of the frontal-wave precipitation pattern. The vertical shear was only half as strong in eastern Kansas and Missouri where the MCCs passed through the postmature and dissipating stages.

4. Evolution of systems A and B

a. Overview

Systems A, B, and C tended to develop similar patterns of convective precipitation. System A developed during midday hours into a fairly small but long-lived system that propagated from western Kansas to West Virginia in 24 h; for 13 h it qualified as an MCC. During its development in Kansas, the active convective clusters were found on the western and southwestern margins of the cloud shield. Then by 2025 UTC in its growth stage, 4 h after initiation as an MCC and 2 h before the mature stage, the active clusters formed two bands, M and N, which intersected at point P (Fig. 5). This configuration was tracked for 4.5 h at a velocity of 15.5 m s⁻¹ from 285° into central Missouri. The pattern resembles the central portion of the line echo wave pattern described by Nolen (1959) except that Nolen described a wave on a much longer line.

System B attained the initiation criteria for an MCC only 2 h after system A did, and followed much the same path. It developed a larger cloud shield and exhibited the best example of the open-wave pattern in this study. The evolution of the precipitation echo, cloud shield, and surface pressure and wind fields has been depicted by Stumpf et al. (1991; hereafter called SJS-91; cf. Figs. 5, 6, and 7 of that paper). The transition from an irregular to a wavelike pattern of convective echo, during the mature stage only, is illustrated in Figs. 7 and 8 of Augustine and Howard (1988; hereafter called AH-88). The first storms appeared between 1500 and 1800 UTC 3 June (in the late morning hours) in west Texas and developed a continuous multicellular line that extended 400 km in a north–south direction at 1900, when the system qualified as an MCC (see Figs. 6a,b of SJS-91). It propagated downshear by incorporating convective clusters that had developed separately in western Kansas up to 300 km in front of the existing convective line. A “cluster” is here used to mean a collection of cumulonimbis, each identifiable by high radar reflectivity, that collectively span a region 20–100 km in diameter and jointly appear to generate a precipitating anvil cloud. Because the new clusters
developed simultaneously over a wide area far ahead of the older convective clusters, a mesoscale cold pool created by downdrafts from the latter are probably not responsible for initiating the new convection. Some, but not all, of the clusters appeared to form on the stationary front and on the outflow boundary left by system A.

A mesoscale open-wave pattern formed between 2300 UTC 3 June and 0030 UTC 4 June (see Fig. 6 for a surface analysis at 0100 overlaid on the radar echo and cloud-shield outline; also see Fig. 8 of AH-88 and Figs. 7c,d of SJS-91). This stage has been described by Meitin and Watson (1989) and Smull and Augustine (1989) as an evolution from an extremely irregular pattern of convection without stratiform echo to the open-wave pattern. Both Smull and Augustine (1989) and SJS-91 commented on the different character of the two bands that composed this wave pattern. The southern band consisted of a multicellular convective line that extended south from the MCC cloud shield and propagated eastward from the strongest of the early clusters in Texas. Unlike the discrete clusters mentioned in the previous paragraph, this line is at the leading edge of a low-level cold pool. It remained upstream of the stratiform cloud, so that debris from the convection was carried northward along the line toward the center of the open wave. The strongest cell of the line produced a tornado at END 15 min after the time of Fig. 8.

The northern half of the open wave consisted of one and, briefly, two bands of enhanced stratiform precipitation with embedded convective centers (AH-88) aligned parallel to the synoptic-scale stationary front. Smull and Augustine (1989), using evidence from dual-Doppler analysis, characterized the cellular convection in the northeast–southwest band as weaker and suppressed, but also broader, with ascent observed in a 40-km-wide band, from heights of 2 km to at least 16 km. Maximum convergence was found at the lowest level in the north–south band, which resembled a squall line, but somewhat higher (near 2 or 3 km) in the broader east–west band, which lacked a gust front.

Fig. 5. Radar reflectivity pattern in system A at 2025 UTC 3 June, in southeastern Kansas. The four gray levels are keyed to the reflectivity (dBZ) at the left.

Fig. 6. Mesoscale surface analysis of the mature system B at 0100 UTC 4 June. (a) Pressure centers, streamlines, wind confluence lines (dotted), and synoptic-scale fronts superimposed on reflectivity pattern (shaded) from Wichita radar, and the –54°C temperature contour of the margin of the cloud shield (thin-dashed line). (b) Isobars of pressure reduced to a standard elevation of 480 m at 1-mb intervals; pressure troughs (thick-dashed lines), other features as in (a).
Convergence was maximized in the midlevels in the stratiform region. Leary and Bals (1989) also documented these structures and reported a second band of enhanced stratiform echo parallel to the first and longer-lived band associated with the stationary front.

In the analysis here, the relative southerly component of the winds, observed in the 3–8-km layer, accelerated during the hour prior to arrival of systems B and C. This may have been a "warm surge" of a prior short wave that participated in the development of the MCCs.

The intersection of the bands (which we denote as the apex of the open wave) had numerous, more intense convective centers and large, relative front-to-rear flow in low and middle levels, in Smull and Augustine's (1989) study. At high levels, the velocity field diverged from the apex region, especially toward the northwest where the stratiform precipitation zone was broadest. Rear inflow from the west was observed into the northern half of the latter region; rear inflow was not observed near the southern convective line.

Augustine and Howard (1988) added that the flow converged from both front and rear on a frontal-type feature associated with the north–south line, and that a significant difference of $\theta_e$ ($\sim 20$ K) existed across the front. Unsaturated mesoscale downdrafts propelled the feature south and east.

These features are reproduced in the composite analysis in Fig. 6 when the frontal wave was best manifested. The stronger mesohigh remained associated with the intersection of the two main precipitation bands and formed a ridge associated with rain-cooled, low-level outflow that extended 400 km to the south. The weaker mesohigh was associated with stratiform rain north of the primary east–west band of precipitation (continuous thunder was heard in this region even though the rainfall was light). The displacement of the mesohighs and the wind-shift lines was tracked for 3–6 h at a propagation rate of 20 m s$^{-1}$ to the east-northeast, the same rate as the radar echo pattern. That pattern propagated from 240° in the growth stage, and from 270° in the mature and late stages, in agreement with the large-scale flow in the 850–300-mb layer.

SJS-91 concluded that it was significant that surface mesohighs and mesolows were present in the northern stratiform region but not the southern convective line. The rear inflow was also associated primarily with the stratiform region by these authors, Smull and Houze (1987), and others. Relative inflow from the rear was observed by the Liberal, Kansas, profiler (LIB) very early in system B, 1 h after a strong convective band formed east of LIB. The inflow was observed in the 5.5–6.5-km layer for five more hours (see Fig. 9 of AH-88). Stumpf et al. (1991) attributed the genesis of the wake low pressure to dry-adiabatic warming in a narrow (10-km) zone where the rear inflow descended abruptly, at rates up to 6 m s$^{-1}$, on encountering virga falling from the stratiform cloud. In the precipitating region, the warming was offset by evaporational cooling of rain, but after the virga evaporated the warming was not offset. The maximum sinking was located in the region of largest reflectivity gradients (at the rear of the precipitation), and above the region of maximum gradient of surface pressure [2 mb (5 km)$^{-1}$], which led to sudden easterly acceleration of the flow (gusts to 15 m s$^{-1}$).

The authors noted that the surface pressure patterns did not mimic those of a synoptic-scale wave cyclone. Mesohigh pressure was located near the apex of the wave where low pressure would be found on the synoptic scale; mesolow pressure was located farther west at the rear of the MCC (see Fig. 6).

Thus, the open-wave pattern of precipitation echo and surface boundaries became well developed in system B, which can serve as an analog of system C. Since observations of B are more complete than those of C, evidence of the main features of the open-wave type of MCC will be drawn from both systems.

b. Cool inflow into and beneath the stratiform cloud

It was previously noted that cool air (with $\theta_e$ values near 325 K) and east-northeast wind prevailed in the boundary layer north of the surface front in advance of systems B and C. Evaporation of rain from system A stabilized further the cool surface layer as temperature and wind changed little during the passage of systems A and B. The dewpoints increased by 5–10 K while rain fell at locations north of the convective band. The onset of rainfall in system B (3–6 h later) then caused widespread fog. The boundary layer was not thermodynamically active in the stratiform area of these MCCs.

The middle levels of this region received an inflow of cooler air, however. For several hours prior to the disturbance, the temperature profile at RSL was like that at 2030 (Fig. 4e): close to moist adiabatic from 870 to 710 mb, but potentially unstable if lifted. Occasional sprinkles were observed after 2200 and continuous rain began at 2320; by 0000 the temperature had decreased in the 750–520-mb layer, and had increased above that (solid and dashed profiles, Fig. 7a). Temperatures had recovered to earlier values in the cooled layer by the time of the next sounding (0130, dotted), shortly after the cloud shield had exited. The layer below 700 mb progressively dried until by 0130 extreme warming and drying had depressed the low-level inversion to only 200 m above the surface. This "onion sounding" (Zipser 1977) is evidence of a mesoscale unsaturated downdraft. Substantial cooling in the 900–600-mb layer and warming in the 600–270-mb layer was observed at Fort Riley (FRI), Kansas, between 2250 and 0020 UTC (Fig. 7b, dashed and dotted profiles). (Stratiform echo surrounded the station from 2320 to 0345.) At the same time, the relative winds in the cooled layer acquired a component from the left side of the MCC, at both locations. Both evaporation of rain and cold-air advection appeared to be
It will be shown later that relative inflow from the northeast and sinking motion was observed by dual Doppler radar in the 3–7-km layer of the northern part of the stratiform cloud of system C. The midlevel cooling demonstrated for system B in Fig. 7 is postulated to have occurred in this region of system C.

Thunderstorms have been shown to occur regularly in environments without positive CAPE in the warm air above fronts (Colman 1990a,b). Although some layers of the FRI and RSL soundings did exhibit positive CAPE, as mentioned, the amount of available energy was small. In other respects the northern-regime soundings in Fig. 7 resembled the composite environment for elevated thunderstorms in Colman’s (1990a) climatology. In his composite, these storms tended to occur in regions of low-level warm advection, above a sharply defined front having a shallow slope, and north of the surface frontal position; they also were found in the right exit region of a 500-mb anticyclonic jet. All of these were characteristic of the 3–4 June episode. It is worth noting that his plotted frequency of elevated thunderstorms exhibits a distinct maximum over eastern Kansas (Colman 1990a, Fig. 1).

c. Kinematic evolution

The kinematic analysis technique of Bellamy (1949) can be applied to the special sounding dataset that sampled system B fairly well at 2230 (the growth stage) and 0000 (the mature stage). A polygon of nine triangles enclosed nearly all the precipitation echo at 0000 (Fig. 8). A different polygon, which enclosed less of the system, was drawn at 2230.

Profiles of calculated area-averaged divergence and vorticity are depicted in Fig. 9. In the growth stage, convergence was established and vorticity was weak but positive below 400 mb. High-level divergence was becoming established.

At 0000 the convergence had increased and exceeded $5 \times 10^{-4}$ s$^{-1}$ in the middle layers from 825 to 450 mb, and anticyclonic, divergent outflow had become established in upper levels from 350 to 150 mb. This agrees with previous results from mesoscale convective systems. However, the vorticity profile was essentially zero from the surface to 400 mb. Further analysis showed that the area averages were sensitive to the number of stations that sampled the rear inflow—a strong inflow led to anticyclonic conditions on the right, southern flank of the inflow, and vice versa. The weak anticyclonic profile near 800 mb is also consistent with the veering of the southerly flow as it approached the deformation zone on the synoptic scale (see Fig. 3b).

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2 Triangle A was three times larger than any other triangle, and most of it was located outside the region of precipitation. The inclusion of triangle A weakened the signal in Fig. 9 considerably. Thus, the profiles in that figure are calculated without triangle A.

3 Data beneath the 800-mb level should be interpreted with caution because the absence of winds from a wind profiler meant that only one-half of the MCC echo area was sampled below that level.
Fig. 8. Locations of the Bellamy triangles marked by capital letters on the radar reflectivity pattern of system B at 0000 UTC. Reflectivity is shaded as in Fig. 6a. Vertical profiles of vertical motion \( \omega \) in six triangles are displayed around the periphery. The boxes span a range of \( \omega \) from \(-50 \times 10^{-3}\) to \(+50 \times 10^{-3}\) mb s\(^{-1}\) on the horizontal axis, and pressure from 1000 to 100 mb on the vertical axis.

Results from triangles B (the apex region) and F (where the rear inflow was strongest) are depicted in Fig. 9c. Convergence was pronounced from the surface up to 450 mb in the apex region and from 700 to 300 mb in the rear inflow; the latter region was divergent below 750 mb. Relative vorticity was strongly positive (exceeding \(10^{-4}\) s\(^{-1}\)) in the apex region from the surface to 825 mb, and weaker but positive in triangle F at

Fig. 9. Vertical profiles of horizontal divergence \( \delta \) and relative vorticity \( \zeta \) in the polygons containing the MCC (except triangle A), at (a) 2230 UTC 3 June and (b) 0000 UTC 4 June. The divergence (dashed) and vorticity (solid) are profiled in (c) for triangles B and F. The triangles are located in Fig. 8.
those levels. Vorticity was cyclonic in triangle $F$, but not $B$, in the middle troposphere. The outflow was divergent and anticyclonic above 300 mb for both regions.

The inset in Fig. 2b shows two wind observations in Kansas, at Pratt (PTT) and Wichita (IAB), with a strong southerly component. A line from PTT to IAB forms the southern base of triangle $B$ in Fig. 8. The inclusion of these winds affected the vorticity field at 500 mb, which became more anticyclonic downstream of triangle $B$ and more cyclonic upstream of it, relative to the background field (dashed lines) computed without the mesoscale soundings. Together with Fig. 9, these results suggest that the midlevel (800–400 mb) response to system $B$ was convergent but not (consistently) rotational.

The divergence was vertically integrated to yield vertical motion $\omega$, profiles of which were plotted for six triangles on the periphery of Fig. 8. The customary corrections, which are done to force $\omega$ to zero at the surface and at 100 mb, were not applied because data from eight of the nine triangles were missing at 100 mb.

**Fig. 10.** Evolution of surface features and precipitation echo at five times (indicated at upper right) in the life of system C. Surface high and low pressure centers, isobars, and fronts are drawn as in Fig. 6; the leading 9 has been omitted from the labels of the isobars (55 = 955 mb). Surface streamlines are thin-dashed, confluence axes are thick-dashed lines. Radar reflectivity (dBZ) levels are indicated by the legend at lower right in (a). The scalloped cloud symbols outline the area of the cloud shield colder than $-64^\circ$C in all panels [and also colder than $-54^\circ$C in (d) and (e)]. The rectangle locates the domain of dual-Doppler analysis on (c)–(e), and the black circles locate the two Doppler radars.
mb; thus, the profiles should be interpreted with caution. The strongest rising motion was found in the apex region (triangle B) from 500 to 200 mb, and at progressively lower levels from west to east in triangles C and D, which include the stationary front precipitation band. Triangle M, which included the north–south convective band, had stronger ascent at lower levels than in triangles C and B (data were absent above 500 mb). Descent predominated to the north of the stratiform region (A), and the rear-inflow region (F) exhibited an S-shaped profile of ascent above 5 km and equally strong descent below that.

To summarize, system B induced strong convergence in midlevels and strong divergence at high levels, but no systematic meso-α-scale vorticity was revealed, although cyclonic vorticity did increase in the core region and the rear inflow region in the low and middle levels.

5. System C: Evolution from a chaotic to a banded pattern

Only 10 h after system B passed through the observational network, a third convective complex delivered more than 7 cm of additional precipitation to the same area. The 24-h rainfall from the three mesosystems totaled from 7 to 15 cm in a band 80 km wide in eastern Kansas. The convective elements of system C were dispersed in no obvious pattern for the first 3 h, but then tended to organize in broken bands that resembled at first an open-wave pattern, and later a spiral pattern.

The first storms developed into a narrow convective line north of Amarillo, Texas, shortly after 0400 UTC 4 June. By the time that the cloud shield met the initiation criteria for an MCC at 0630 UTC, the line had evolved into an enduring, wavelike cusp at P (Fig. 10a).

South of P was a solid line of intense echo; the portion northeast of P was a broken band. The convection near P, which developed in surface easterly flow 100 km northwest of weak low pressure on the stationary front, evolved into the core convective region of the mature MCC, and the open wave configuration maintained continuity during the next 4 h. The coldest part of the cloud shield (within the scalloped boundary) was found above the core convective region and downstream of it at cirrus level.

New clusters formed in two well-separated areas. Some, like B and C, developed in a weak surface trough 100 km east of the core convection at A and P in Figs. 10a,b. These clusters formed well within the main cloud shield. Aridge of mesohigh pressure was beginning to be analyzed by 0800 in the expanding area of stratiform precipitation. Other clusters at D, E, and F developed some 300 km ahead of the core region in the direction in which the cloud shield was advancing (Figs. 10b,c). They generated distinct cloud shields that remained separate for 2 h until they merged with the main shield.

A convergent wind-shift line (dashed) formed between outflow from the mesohigh and the ambient easterlies. The line joined the low pressure on the front, with the preexisting outflow boundary from system B in the northeast part of Fig. 10b.

Over the next hour, the overall convective pattern became less chaotic as new cells formed between existing clusters, which were elongating and merging. By 0900, when system C was in its growth stage, a new convective line was observed from P to G, which was in part a result of faster-moving clusters such as A catching up with slower ones like B and C (Figs. 10b,c). Stratiform precipitation and an associated mesohigh were spreading out west of the northern, but not the southern, portion of this line.

The stationary convective band at D lengthened in both directions until a continuous line of echo extended to P. This line may have been loosely associated with the east–west wind-shift line (dashed), though the association was not as close as in system B. If the east–west line PD is considered together with the north–south line PG, then a configuration of convective echo resembling an open wave became established on the meso-α-scale by 0900. Unlike system B, new, relatively intense clusters like the earlier E and F continued to develop in a chaotic arrangement in a pressure trough south of the wind-shift line and east of the convective line PG. This configuration soon changed into a spiral alignment of echo (heavy curve on Fig. 10d) that developed inward of the open wave. The spiral pattern apparently did not influence the surface wind or pressure fields, which were increasingly dominated by the mesohigh located west of the convectively active core of the MCC. A wind confluence line (dashed) and wake pressure trough were forming just west of the stratiform echo.

A dual-Doppler wind analysis (section 6) was performed at 1040 UTC in the boxed domain on Fig.
10c, just before MCC maturity at 1100. As the cloud shield was advected northeastward, its colder portion (<−64°C) no longer was located over the western half of the stratiform echo. At the surface, a now pronounced trough extended from the active convective clusters in the eastern half of system C to the wave on the stationary front in Oklahoma (where no precipitation was occurring). A second "wake trough" hugged the rear margin of the stratiform echo and extended southward. It formed the western boundary of low-level outflow that remained after the precipitation departed.

The motion of the convective clusters differed greatly in the 4-h period encompassing the growth and mature stages. Clusters such as E and F, located east of the advancing cloud shield until they merged with it (Figs. 10b–d), advanced at an average velocity of 8 m s⁻¹ from nearly south (200°). They may have been rooted in warm sector air disconnected from the surface layer (see section 7e). Clusters in the core convective area, such as A, advanced at 20 m s⁻¹ in step with the open-wave pattern of echo. Band PG advanced from due west at 15 m s⁻¹ and lagged behind the rest of the precipitation after 1100.

It is difficult to select a single overall propagation velocity because two regimes were observed. The cloud shield and the surface mesohigh were displaced at an average velocity of 29 m s⁻¹ from 245°, in close agreement with the wind shear in the 850–300-mb layer. The mesohigh was closely associated with the stratiform precipitation region in both systems B and C. In contrast, many of the larger convective clusters, and the open wave configuration of echo, propagated at 20–23 m s⁻¹ from 235° before 1000 and from 255° after 1000. A reference motion of 20 m s⁻¹ from 235° was chosen for the purpose of computing system-relative winds.

6. Internal flow of system C

Velocity fields were synthesized from the Doppler data at several times from 0900 to 1200 UTC in the maturing phase of system C. Figures 10c–e locate the rectangular dual-Doppler domain within system C. The propagation vector of system C was oriented at an angle of 8° from the long axis of the domain. Several features were noted in these wind fields, which are depicted relative to the moving MCC.

1) Convergence in the main convective band. The midlevel flow converged strongly into the main convective band, identified by the spiral line at 1002 (Figs. 11 and 10d). Forty minutes later, when this band was

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**Fig. 11.** Relative wind vectors and reflectivity field at the 4-km level at 1002 UTC, before the mature stage. The rectangular domain of Figs. 11–13 is located on Fig. 10. The heavy solid curve marks a line of convective cloud development: the dashed line from Z to R marks an axis of horizontal wind shear. The reflectivity contours are 10 dBZ (thin solid), 20 dBZ (thin dashed), 30 dBZ (light shading), and 40 dBZ (dark shading). Black circles locate the two Doppler radars.
less distinct, the flow was still convergent at the 4-km level in the northeastern corner of the domain in Fig. 12. More convergence was found in the band than in isolated clusters like K and L in Fig. 11.

2) Confluence in the front-to-rear flow. Confluence in the high-reflectivity band through R seemed to accelerate the flow northward toward P, especially at the later times (Fig. 12). Confluence in a westward direction was found along \( y = 65 \), not only near the convective band, but also behind the band as far as T (Fig. 11). North of \( y = 65 \), the flow had a northerly component at these levels; confluence with the south-southeast flows led to strong front-to-rear (FTR) motion just north of the dashed lines on the figures.

3) The front-to-rear–rear-to-front interface. Relative, rear-to-front (RTF) inflow from the southwest converged with FTR flow on the dashed line, especially at later times near the 4-5-km level. Along this line (DR in Fig. 13, TR in Figs. 11 and 12) there was pronounced horizontal wind shear and high values of cyclonic relative vorticity (\( > 5 \times 10^{-5} \text{s}^{-1} \), shaded in Fig. 12). High values of deformation (\( > 10 \times 10^{-5} \text{s}^{-1} \)) were also found on the axis.

Two cross sections, BC and EF (located on Fig. 13 and shown in Fig. 14), depict the vertical structure of the flow in the front-to-rear sense. The rear-to-front (left-to-right) speed component \( u' \) is contoured. The deep region of front-to-rear flow is better depicted on section BC (Fig. 14b), where it is associated with upward motion from heights of 4 to 11 km, on the right side. Two narrow convective-scale updrafts at \( x = 80 \) and \( x = 110 \) km were feeding into the mesoscale, midlevel updraft. The cross section EF (top) contained a strong updraft from \( x = 90 \) to 120 km, associated with higher reflectivity and strong relative southerly flow on Fig. 13, in the remnant of the spiral convective band. Behind the updraft region, the flow descended from a height of 8 km at \( x = 80 \) km to the lowest observed levels at \( x = 40 \) km. This outflow from the convective updrafts apparently descends first into a zone of lower reflectivity before it ascends into the stratiform region, and is similar to the transition zone found in squall lines (Smull and Houze 1987a). A smaller transition zone is seen at \( x = 100 \) km in section BC.

A distinct tongue of descending, rear-to-front flow originated behind the precipitation in section EF at heights of 4-7 km, and proceeded forward and downward to RR. The interface between this flow and the front-to-rear flow above it is well defined by the gradient of the \( u' \)-component wind. Below the tongue, from the lowest observed level of 1 km up to 2 or 3 km, front-to-rear outflow prevailed in the stratiform region behind R. The descending tongue of air is apparently a form of rear inflow as described by Smull and Houze (1987b) for squall-line MCCs. However, other cross
sections, including the BC section, exhibited nonuniform descent or even regions of ascent within the rear inflow, and the boundaries of the inflow were less well defined. While the prevailing flow southwest of the TR interface was rear inflow, there was much variability in its magnitude. The stronger inflow tended to be located behind the more intense precipitation—a point discussed in McAnelly et al. 1992.

A cross section GH, perpendicular to the rear inflow (see Fig. 13 for location), exhibits contours of the same u component as in the previous sections, in this case normal to the page (Fig. 15a). The interface DR was well defined by the strong horizontal shear in the zone that sloped steeply northward from D at $y = +40$ km. To the right, northeasterly inflow descended from the 4 to at least the 1-km level and perhaps to the surface. For the purpose of comparing the properties of the northerly inflow in systems B and C, we have drawn a vertical line at $y = +80$ to approximate the system-relative location of the Russell sounding at 0000 UTC (plotted in Fig. 7a, located in system B in Fig. 8). At that location, northerly relative flow changes to southerly flow at the 4–5-km level. Another interface sloped southward with height in the vicinity of $y = 30$ km, $z = 6$ km. The rear inflow, which had a southerly component in system C, occupied a greater depth towards the south.

Of interest is the large area of rising motion from $y = -15$ to +15 km and from heights of 2–7.5 km in the rear-inflow region (left side) of Fig. 15a. The rising motion was centered on the intersection of lines BC and GH in Fig. 13. At the time of this figure, a notch of low reflectivity was located there, but in the previous hour a strong convective cluster was observed upwind of the region. This cluster (J on Fig. 11) was located at the extreme rear of the stratiform rain at 1002 UTC, and weakened while propagating south of the intersection by 1042. Many other vertical sections do not show such a rising-motion feature at the back of the stratiform region; along section KL (Fig. 15b) the vertical motions are predominantly downward along the interface between forward-directed and rearward-directed motion ($y = 30$, $z = 6$ km; and $y = 50$, $z = 4$ km). Some upward motion is observed from 2 to 3 km in the rearward-directed, cool outflow at $y = 30$ km. This air may be lifted by a surge of stronger outflow arriving from the northeast.

Isolated convection, then, was still observed on the southern rear margin of this MCC, even though the convection was surrounded by descending dry inflow. Although the convection may have been assisted by low-level outflow boundaries, there must have been regions beneath the stratiform cloud where the supply of positive CAPE was not exhausted.
Wind profiler and aircraft data agreed with the radar-derived rearward flow north of TR at the 5-km level. West of the stratiform echo, where Doppler data did not exist, other data indicated forward-directed flow toward the MCC in the 4.5–7-km layer. The aircraft flew through a sloping interface between the two flow regimes somewhere near the rear margin of echo.

The horizontally averaged divergence and vorticity fields were analyzed at each height in the domain of Doppler analysis. The vertical profiles can be compared with the results from system B (Fig. 9) with the caveat that the Doppler domain never contained all of the echo of system C, and was sampling different portions at different times. Convergence was established at all levels below 6 km in the growth stage of system C, with a peak value at 3 km (Fig. 16a). The level of peak convergence rose and the lowest level became divergent at the mature stage (Fig. 16b), which is typical for MCCs.

Unlike system B, area-averaged vorticity did increase in system C at all levels below 8 km, between the growth and the mature stages. The lowest level exhibited the greatest cyclonic vorticity, and values of $10^{-4}$ s$^{-1}$ extended from 3 to 7 km. The increase may result in part from the fact that the Doppler domain contained more of the rear inflow region and less of the eastern half of the system at 1042 than earlier. The rear margin of system B exhibited more vorticity than other portions; however, it was not as pronounced as in system C, nor did it occur in a layer as deep (compare Figs. 9c and 16b). It is felt that the generation of midlevel vorticity did, in fact, occur in system C.
Fig. 15. Vertical cross sections along lines perpendicular to the system motion. Lines GH and KL are located on Fig. 13. The relative u component (positive out of the figure) is contoured as in Fig. 14, and represents the flow in the direction of system motion. Vectors depict the relative flow perpendicular to system motion. The approximate location of the 0000 UTC RSL sounding (Fig. 7a) relative to this system is indicated by the vertical line at $y = 80$.

7. Discussion

a. Prototype of a frontal-wave–like MCC

The precipitation and flow structure of a mesoscale system that develops a wavelike pattern of convective echo can be inferred from the analysis of systems B and C. The effects at the surface are less pronounced than the effects of a squall line MCC. The strong winds and temperature changes of a squall line are mitigated in the wavelike systems by a layer of stable air that is always present north of the surface front. When MCCs recur in an episode, the outflows of the earlier systems may stabilize the surface layer south of the front, so that the subsequent systems obtain their potential buoyant energy from the low-level jet above the stabilized layer. The precipitation consists of extensive light rain showers with embedded thunderstorms north of the existing front and numerous, discrete thundershowers in the vicinity of the front, although a squall line may extend south from the southern margin of the MCC.

The relative flow in the prototype MCC is depicted in Fig. 17 at three levels: the source layer for the convective updrafts, the layer of midlevel convergence, and the upper-level outflow.

At the level of the low-level jet (850 mb), inflow having a typical $\theta_e$ value of 350 K is accelerated from the southeast into convective and mesoscale updrafts (Fig. 17a). A distinct inflow from the northeast advects lower temperatures into the stratiform rain area; west of the surface mesohigh at $P$ this flow is accelerated rearward by the strong pressure gradient. It is also accelerated southward, but not as strongly; where this outflow converges with ambient flow south and southwest of the cloud shield, isolated convection may be observed.

At the level of maximum systemwide convergence (500 mb), mass is added by detrainment of updrafts in the convective regions (shaded, Fig. 17b). This warm airstream is accelerated rearward, especially in a ~30-km-wide jet through $P$ toward the rear of the stratiform precipitation. Inflow from the south and the north of the MCC converge on the flanks of this jet at the dashed interfaces. The inflow from the south and southwest is dry and most of it descends rather abruptly beneath the warm airstream where it encounters precipitation. However, areas of upward motion, believed to be remnants of convective clusters that had been in the apex region, are found in the vicinity of $B$. They are surrounded by a mesoscale downdraft that contributes to onionlike soundings along the rear margin of the precipitation. The relative flow is light and variable in the southeast quadrant.

At 200 mb, the system outflow diverged (somewhat anticyclonically) from the apex region (Fig. 17c). A heavy black line denotes a confluent axis along the western and northern margins of the cloud shield, where the ambient flow (15–30 m s$^{-1}$ relative to the cloud) encountered ice-saturated outflow from the cloud.

An idealized cross section through the rear half of the cloud system west of $P$, along a line $BB'$, shows the warm ascending airstream completely elevated above the surface (Fig. 18). Inflow in the 3–7-km layer converges on the warm jet from both south and north; the relatively warm but dry inflow from the south descends in unsaturated mesoscale downdrafts that do not normally reach the surface. Properties of the cool inflow are inferred in part from the soundings in Fig. 7, the approximate location of which is marked by the vertical line at $r$. The cooler inflow descends beneath an ill-defined interface; stronger downdrafts reach the surface in the band of enhanced precipitation.

b. Spatial pattern of convection and rotational tendency

The rainfall patterns that resemble a wave cyclone were established roughly midway between the initiation
and the mature stages of these MCCs (in the growth stage of Cotton et al. 1989). If rotation developed on the meso-α scale, it did so after the systems left the observation network after the mature stage. Cyclonic vorticity was observed in low-to-middle levels particularly in the apex region, and on the shear interface between the warm airstream and dry inflow, but it was not consistently detected on the scale of the entire cloud system: one system exhibited much more vorticity than the other. Thus, a cyclonic circulation cannot be causing the wave-cyclone organization of convection. Could the configuration of convective bands enhance the mesoscale vorticity? Perhaps, in a limited way, the intersection of two bands enhances the meso-β-scale convergence, which further favors convective cluster activity. If maintained long enough, that may foster eventual cyclonic adjustment on a larger scale.

What causes the wave-cyclone pattern of organization? The convective complex propagates along a frontal zone and enhances convergence into rainbands parallel to the front. The orthogonal convective bands were traced back to clusters initiated on the High Plains. Their cold pools tended to merge and grow to meso-α-scale size, and an outflow boundary advanced southward and eastward on the downshear side. Sometimes the boundary came to resemble a squall line or mesoscale cold front; other times the convective line was broken. In any case, new convective cells increasingly

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![Fig. 16. Vertical profiles of horizontally averaged divergence (dashed) and vorticity (solid) in the dual-Doppler domain, in the (a) growth and (b) mature stages of system C.](image)

![Fig. 17. Idealized relative flow at three levels in a prototypical frontal-wave-like MCC. Dark shading indicates convectively active areas; light shading, stratiform precipitation reaching the surface. Numbers in (a) indicate typical observed values of equivalent potential temperature θe. Dashed lines indicate confluent interfaces of different airstreams. The location of cross section BB' is indicated here. The apex region is denoted by P.](image)
developed on the two pseudofronts and especially at their intersection.

Although important at the mature stage, the pseudocold front was not an enduring component of the MCCs. In systems A and B the front lagged behind the cloud shield and became broken; in system C it never was continuous, and disappeared at the mature stage. But the core region of clusters tended to endure longer than the convective lines.

If the convective line along the pseudo-cold front is broken as in system C, then pools of high-valued $-\theta_e$ air remain intact as the line passes eastward. Thus, new convective clusters often develop on the rear margin of the cloud shield, in an environment that normally is unfavorable for convection in low levels. Surges of cool air from the mesohigh to the north may provide a mechanism to lift such pockets of conditionally unstable air to their level of free convection. Isolated clusters were observed along the southwestern margins of systems A, B, and C, but were more evident in C.

c. Early chaotic patterns of convection

In systems A, B, and C, the early convective clusters were organized in no obvious spatial pattern for some 4 h before the meso-$\alpha$-scale convective bands became apparent. This lack of a classifiable pattern has been reported in many MCCs (see Blanchard 1990 and Houze et al. 1990). Why did dozens of discrete cumulonimbis erupt nearly simultaneously over a 300-km-wide region? Analysis of the soundings in the wake of systems A and B (as in Fig. 7 and subsequent soundings) supports the following hypothesis. In the wake of an MCC an extreme example of an onion sounding is observed, with a cool, saturated, and foggy layer on the surface, beneath an inversion some 15 K strong. The synoptic-scale wind profile is observed to return 3 h after the MCC departs, and the low-level jet steadily moistens the layer above the inversion. (The return of the moisture could be tracked at the 850-mb level.) When the layer attains a critical mixing ratio it becomes conditionally unstable over a wide region, at which time numerous convective cells are observed to develop in the region. Their downdrafts have densities similar to that of the still-present, cool surface layer. Thus, gust fronts do not readily form, and the convective clusters may remain relatively unaffected by neighboring outflows for some time. The primary bands eventually develop on the leading edge of a deeper, meso-$\alpha$-scale outflow in the case of the north-south convective line and in the preexisting stationary frontal zone.

d. Shear profile and rotational tendency

The direction in which the precipitating stratiform clouds were blown may have affected the evolution of vorticity. At the 11-km level, prevailing winds were 35 m s$^{-1}$ from the southwest. As the broken convective line (G on Fig. 17) propagated eastward, the prevailing winds transported the ice cloud northward along the line and through the apex region. Consequently, a stratiform cloud shield did not surround the southern convective regions; instead, it was displaced to the north. It was noted earlier that a stronger dry inflow was observed in the northern half of these MCCs.

Our work is consistent with the hypothesis of Stumpf et al. (1991) that the stratiform precipitation is responsible for initiating midlevel inflow from the rear. Numerical experiments of Chen and Cotton (1988) and Schmidt and Cotton (1990) suggest that in-cloud latent heating and cloud-top radiative cooling (in both convective and stratiform clouds) create high pressure at cloud-top level (which diverts the upper flow downward
and around the cloud tops) and low pressure at middle levels, which drives the convergence at that level. In the MCCs of 3 and 4 June, the inflow converged from all directions: north, east, south, and west. If the stratiform cloud is sheared north of the convective region, as it was by the unidirectional wind shear above the 3-km level, then inflow beneath that cloud will be displaced from the lower-level inflow into the active convective regions. It is hypothesized that a weak-shear profile permits an ice cloud to accumulate around the core region, so that latent heating and radiative cooling from cloud tops will enhance a deeper and stronger convergence into the core.

8. Summary

A large fraction of mesoscale convective systems are observed to have a spatial pattern of convective activity that is not readily classifiable, or that has been classified as chaotic. In other MCCs, convective bands assume a configuration resembling a synoptic-scale frontal wave. We examined the evolution and internal flow of two MCCs that evolved from a chaotic pattern to an organized, frontal-wave–like pattern of convective activity. They were part of an episode of recurring MCCs that propagated along a stationary front and thus caused large accumulations of rainfall. The internal flow fields were analyzed for evidence of development of mesoscale vorticity and its possible relationship to the frontal-wave pattern and other patterns suggestive of rotation.

System B of the episode of 3 and 4 June 1985 developed a mesoscale wave consisting of two dissimilar bands of convective echo associated with low-level convergence boundaries. One band resembled a precipitating stationary front with convective echoes embedded in a sloping updraft in enhanced stratiform echo. The other band was a broken multicellular convective line, oriented north–south, that propagated eastward through the warm air. The open wave evolved in the mature stage from an irregular pattern of echo.

System C, on the other hand, exhibited a chaotic pattern of convective clusters through most of its life. For 1–2 h at maturity, the clusters tended to align along intersecting bands or along a spiral-shaped band. Even at this stage some clusters were distributed chaotically.

In both systems the intersection of the convective bands (the apex) harbored the most numerous convective clusters and exhibited the greatest low- to midlevel convergence and high-level divergence. The broken convective line that extended south from the apex was a less enduring phenomenon than the convection in the apex or the stationary band; it originated from the merger of low-level outflows from the early storms in the genesis region of the MCCs. This line tapped the most unstable air and generated a severe thunderstorm in system B.

A conceptual model of a frontal-wave–like MCC derived from Doppler radar data and rawinsonde data is presented in Fig. 17. The dashed lines separate a potentially warm ascending airstream from a potentially cool descending airstream with \( \theta_e \) values some 10–20 K lower. In midlevels the potentially warm airstream exits the convectively active regions in a rearward-directed jet, while potentially cooler air converges on the jet from both flanks and descends beneath it (Fig. 18). Cold advection is found below the 6-km level in the flow converging on the north flank. Vertical motions are not uniform in the inflow on the southern rear flank, with regions of meso-\( \beta \)-scale ascent embedded in the descending flow. A convective line from \( P \) to \( G \) may sometimes be broken, which would allow pockets of untapped, high-valued \( -\theta_e \) air to remain behind the line. These pockets have access to the low-level jet at the 2-km level and could support convective development on the southern rear margin of the cloud shield.

The shear profile affects where the stratiform cloud will be transported as it is generated in the convective regions. In this episode a unidirectional profile of shear carried the ice cloud toward the north side of the convective regions on the stationary front. The asymmetry of the convective–stratiform structure in the 4 June episode may have inhibited a greater response of the rotational wind to the disturbance.

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