Evolution of Environmental Conditions Preceding the Development of a Nocturnal Mesoscale Convective Complex

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ABSTRACT

A large nocturnal mesoscale convective complex (MCC) developed on 4 June 1985 during the Oklahoma-Kansas Preliminary Regional Experiment for STORM-Central (PRE-STORM) field phase. It occurred near the climatological center of the nocturnal maximum in warm-season precipitation situated over the central United States. In this study special rawinsonde and surface mesonet data have been used to examine how the environmental conditions, which supported MCC development, evolved at night over this region. The MCC of interest was the fourth in a series of MCCs, three of which propagated east-northeastward, 100–300 km north of a quasi-stationary surface front. The region where the MCC experienced its most intensive growth was initially characterized by dry and hydrostatically stable conditions (associated with the passage of the previous MCC) above the shallow, wedge-shaped cold air mass. In less than 3 h, interaction between the diurnally varying low-level jet and the frontal boundary led to a local increase in convective available potential energy (CAPE) of over 2000 J kg\(^{-1}\) for air parcels averaged through a 50-mb-deep layer immediately above the frontal surface.

Our analysis shows that the region north of the quasi-stationary surface front became a favored zone for nocturnal MCC development when 1) particularly high CAPE arose due to the transport of moist, high-\(\theta\), air northward above the frontal surface by the diurnally modulated low-level jet into a region of significantly colder midtropospheric conditions, and 2) adiabatic mesoscale ascent, which was particularly strong near the northern terminus of the low-level jet, resulted in significant cooling above the jet axis. The cooling acted together with the strong moisture advection to eliminate convective inhibition (negative CAPE below the level of free convection (LFC)), thus enabling air parcels over a mesoscale region to more easily attain their LFC. Strong and deep mesoscale ascent was absent south of the front. In this region the surface-based deep convection that was supported during the evening hours weakened overnight as the low-level jet veered to a southwesterly direction, resulting in less favorable vertical shear for the sustenance of convective updrafts, while diurnal cooling increased the convective inhibition and raised the height of the LFC.

1. Introduction

Maddox (1980) examined infrared satellite imagery and found that large and long-lived convective weather systems commonly occur during the late spring and summer months over the central United States. He noted that the cloud shields associated with these systems were nearly circular and, therefore, different from linearly oriented cloud shields typically associated with squall lines and, thus, classified such systems separately as mesoscale convective complexes (MCCs).\(^1\) Since the

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\(^1\)The Maddox (1980) MCC definition requires that the eccentricity of the cloud shield be at least 0.7 and that the IR temperature be less than \(-32^\circ\)C and cover an area greater than 100,000 km\(^2\) with an interior cold cloud region having a temperature less than \(-52^\circ\)C over a region greater than 50,000 km\(^2\) for a duration of at least 6 h.

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Wetzel et al. 1983; McAnelly and Cotton 1986) have shown that some nocturnal MCCs may evolve from convective systems that originate during the afternoon near the eastern slope of the Rocky Mountains, which later grow or merge into larger systems as they move eastward into the moist and unstable environment often situated over the central United States. Maddox et al. (1986) noted, however, that such developments account for only approximately one-fourth of the MCC events that are observed over the central United States. Annual MCC summaries (Maddox et al. 1982; Rodgers et al. 1983, 1985; Augustine and Howard 1988, 1991) have further revealed that most of these precipitation systems form in situ over the central United States.

By compositing observations from the conventional synoptic network, Maddox (1983) and Cotton et al. (1989) have shown that MCCs tend to occur in convectively unstable environments characterized by horizontal convergence associated with a low-level jet in the vicinity of a quasi-stationary east–west-oriented surface frontal zone. Maddox et al. (1979) found that a particular type of convective event that often leads to flash floods and also exhibits a distinct nocturnal bias, occurs when thunderstorms are initiated on the northern or “cool” side of a quasi-stationary surface front downstream from the axis of maximum low-level (e.g., 850-mb) flow. Kane et al. (1987) compositing rainfall characteristics associated with 105 mesoscale convective systems, of which the majority qualified as MCCs, and found that 45% occurred under the “frontal” conditions described by Maddox et al. (1979). Despite the recognition of the recurring synoptic-scale environments in which MCCs form, our understanding of the mechanisms that lead to the formation of MCCs and related nocturnal convective events over this region is incomplete (Sangster 1990; Maddox and Howard 1990). For example, Maddox and Howard state, “We still do not understand the interplay of the large and mesoscale processes that lead to nighttime thunderstorms and widespread rains over the Plains. The accurate prediction of them remains elusive.”

Recent analysis based on mesoscale observations taken during the Oklahoma–Kansas Preliminary Regional Experiment for STORM-Central (PRE- STORM) (Cunning 1986) has allowed further clarification of both the environment and structure of MCCs. Several studies (e.g., Augustine and Howard 1988; Stumpf et al. 1991; Fortune et al. 1992; Smull and Augustine 1993) have investigated aspects of a series of four MCCs that occurred over the PRE- STORM network on 3–4 June 1985. Of these, three matured and followed similar paths along the northern side of a quasi-stationary surface frontal boundary. The Smull and Augustine (1993) and Fortune et al. (1992) studies both found that warm, moist air transported northward above the frontal surface provided the conditional instability to support the convective activity within the MCCs. While these studies examined the internal structure and near environment of the precipitation systems, the objective of our study is to document the evolution of the mesoscale environment that led to the development of nocturnal convection during this episode, in particular the fourth MCC. Herein, we make use of measurements from the PRE-STORM special rawinsonde network and surface mesonet (Fig. 1) to further examine the processes by which convective instability develops over a mesoscale region preceding the growth of a nocturnal MCC. This analysis allows us to clearly demonstrate why the location north of a quasi-stationary frontal boundary is a particularly favored position for the development of large, organized nocturnal convective events over the central United States.

2. Large-scale environment and MCC overview

The locations of the initial formation and subsequent paths of the four MCCs that traversed the PRE- STORM measurement network during 3–4 June 1985 are illustrated in Fig. 2. Similar tracks of MCCs 1, 2, and 4 (Fig. 2), 150 to 250 km north of the mean position of the surface frontal boundary, resulted in widespread heavy rainfall. During this rainfall episode,
At 1800 3 June, a strong, moist southerly flow was present at 850 mb to the south of the surface front (Fig. 3a). This flow pattern was associated with pronounced low-level warm advection in the vicinity of the front. The composite studies of Maddox (1983) and Cotton et al. (1989) have indicated that large-scale low-level warm advection is a common precursor to MCC development. The concurrent 500-mb analysis (Fig. 3b) shows southwesterly flow with a ridge axis situated over the PRE-STORM network. A north–south temperature gradient over the network was also present at this level but was not nearly as pronounced as it was at 850 mb (Fig. 3a). The weak north–south midtropospheric baroclinicity is also characteristic of MCC environments (Maddox 1983). Favorable quasi-geostrophic forcing (e.g., Holton 1979) may be inferred by the presence of the low-level warm advection in the vicinity of the frontal zone (Fig. 3a). Stumpf et al. (1991) reported that additional quasigeostrophic forcing, associated with weak, positive vorticity advection by the thermal wind, may have aided the initiation of the earlier MCCs in this series.

The general synoptic pattern of a quasi-stationary surface frontal boundary, significant veering of the winds with height (indicative of warm advection), and the presence of strong and moist 850-mb-level southerly flow is similar to the composite conditions for frontal-type flash-flood events (Maddox et al. 1979). The considerable spatial variability in the speed of the anticyclonic 500-mb flow (Fig. 3b) was largely due to the presence of MCC 2 during its maximum extent over central Kansas and the incipient stages of MCC 3 over west Texas. The general presence of moderate-to-strong midtropospheric flow represented a departure from the prototypical "frontal event" conditions. The stronger background flow likely contributed to the rapid movement of the MCCs and may have been significant in preventing more serious and widespread flooding during the precipitation episode.

b. Overview of the 3–4 June 1985 precipitation episode

The four MCCs are depicted at various stages of their life cycles in an infrared satellite image (Fig. 4) at 2300 3 June. Their eastward progression within an anticyclonic 500-mb flow (cf. Figs. 2, 3b) was a common characteristic of these convective systems. The occurrence of large, organized mesoscale convective systems developing in succession and following similar paths in favored synoptic situations has been previously documented (e.g., Wetzel et al. 1983; McAnelly and Cotton 1986; Fritsch et al. 1986).

As mentioned in the Introduction, our specific interest lies in the mesoscale modification of environmental conditions that preceded and then accompanied the nocturnal convection that comprised the latter portion of the 3–4 June precipitation episode. The beginnings of MCC 4 can be seen at 2300 3 June as a

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2 Hereafter, all reference to time will pertain to central standard time (CST). To obtain universal time coordinated (UTC), add 6 h.
small area that constituted the northwest appendage of a large region of cold cloud situated near the Oklahoma–Texas border region (Fig. 5a). The main body of the cloud shield was associated with MCC 3, which had attained its maximum areal coverage by this time (Augustine and Howard 1988). The existence of MCC 3 and the developing MCC 4 as distinct precipitation systems is more clearly illustrated in Fig. 6a, which shows a separation of approximately 150 km between the low-level radar reflectivity patterns from the two systems. The radar echoes over the eastern edge of the mesonet (Fig. 6a) were associated with the extreme southwestern edge of the rapidly dissipating MCC 2 (Fig. 4).
By 0100 4 June discrete clusters of deep convection had developed over a broad east–west-oriented zone of surface equivalent potential temperature $\theta_e$ contrast (Fig. 6b), located 100–250 km north of the concentrated $\theta_e$ gradient that marked the position of quasi-stationary surface front over central Oklahoma. This development occurred to the east of the initial core region of convection that comprised the incipient stage of MCC 4, which itself had moved northeastward between 2300 3 June and 0100 4 June (cf. Figs. 6a,b). Meanwhile, the areal extent of both the radar reflectivity (Figs. 6a,b) and cold cloud shield (Figs. 5a–c) associated with MCC 3 had decreased dramatically during the late evening and early morning. A north–south-oriented reflectivity feature extended well south of the frontal boundary and was associated with a pronounced $\theta_e$ gradient that marked the leading edge of the precipitation associated with MCC 3 over southwestern Oklahoma (Fig. 6b).

During the next 4 h the initial core region of convection north of the quasi-stationary front and the discrete convective clusters, located over southern and central Kansas (Fig. 6b), merged to form the large echo mass (Fig. 6c) that comprised MCC 4 at maturity (Fortune et al. 1992). By 0500 4 June, this precipitation system had attained its maximum areal coverage (Fig. 5d), exhibiting a large, cold cloud shield with $T \leq -52^\circ C$ covering 196 000 km$^2$ (Augustine and Howard 1988). At this time, precipitation associated with MCC 3 was no longer present (Fig. 6c). MCC 4 eventually dissipated east of the PRE-STORM network over eastern Missouri (Fig. 2) shortly after sunrise on 4 June. The reasons for the simultaneous growth of MCC 4 and decay of MCC 3 during the late evening and early-morning hours will become apparent from our analysis of environmental conditions to be presented in sections 3 and 4.

Surface data revealed major differences in the characteristics of convection located north and south of the frontal boundary. Changes accompanying the passage of the north–south-oriented precipitation feature marking the leading edge of MCC 3, south of the east–west-oriented surface front (Fig. 6b), resembled those often associated with squall lines. For example, at SAM 21 (Fig. 7, see Fig. 1 for location) a 1.8-mb pressure jump and a 5$^\circ C$ temperature drop in 10 min signaled the arrival of a 25-min period of heavy rain that totaled 13 mm. The sharp wind shift and an approximately 15-K $\theta_e$ drop at the onset of the heavy rain provide evidence that convective-scale downdrafts reached the surface. The sharp west–east $\theta_e$ gradient, extending for approximately 100 km along the line (Fig. 6b), is suggestive of gust-front lifting as the forcing mechanism for ascent of boundary-layer air parcels to their level of free convection (LFC).

A time series from PAM 19 (Fig. 8), located in south-central Kansas (see Fig. 1 for location), depicts the evolution of surface conditions 250 km north of
the surface front, in the path of the maturing MCC 4. While significant surface cooling was observed with the passage of the convective line south of the front (Fig. 7), the 35 mm of convective rainfall that occurred at PAM 19 north of the front during the period from 0220 to 0305 4 June were marked by a slight temperature rise of 0.6°C. The convective period was followed by a gradual 1°C–2°C cooling during a 2-h period of light stratiform precipitation. Although surface temperature changes during the passage of MCC 4 were not pronounced, a substantial mesohigh–mesolow pressure perturbation couplet developed as the MCC matured (Fortune et al. 1992). Temporal pressure changes at PAM 19, associated with the development of the MCC-induced pressure perturbation, are evident in Fig. 8. A similar dichotomy of convective types was also noted in analyses of the mature stage of MCC 2 (Stumpf et al. 1991; Fortune et al. 1992; Smull and Augustine 1993), which occurred during the early evening. These studies showed that convection near the center of the MCC was based above the frontal surface and, in general, lacked penetrative convective downdrafts due to the presence of the strong frontal inversion. In contrast, “squall-like” convection was found near the southern extremity of the MCC where there was little if any impediment to deep, convective downdrafts near the surface. The rainfall totals across the PRE-STORM mesonet associated with nocturnal convection produced by MCC 4 and the later stages of MCC 3 are displayed in Fig. 9. Although the rapid average translation speed of the larger convective clusters that comprised MCC 4 of 20–23 m s⁻¹ (Fortune et al. 1992) precluded a prolonged period of rainfall at individual locations, the
heavily rain rates that occurred during their passage contributed substantially to the total rainfall for the entire episode north of the surface front. The precipitation south of the frontal boundary associated with MCC 3 during its decaying stages appears to be lighter and less widespread; however, its spatial distribution may not be adequately represented due to missing mesonet data over southwestern Oklahoma.

3. Evolution of the surface front and low-level jet

The importance of the frontal zone as a focusing region for deep convection is evident from the location of the maturation of MCCs 1, 2, and 4 (Fig. 2). In this section, the structure of the frontal zone and low-level jet using vertical cross sections that are approximately perpendicular to the surface front at two different times are examined, concentrating on the conditions 1–2 h prior to the development of widespread convection within this zone in association with MCCs 2 and 4. PRE-STORM special rawinsonde measurements (with a spatial resolution of about 150 km), and surface data from the National Center for Atmospheric Research (NCAR) Portable Automated Mesonet (PAM) and the National Severe Storms Laboratory (NSSL) Stationary Automated Mesonet (SAM) (both having an average...
station spacing of approximately 50 km) were projected onto the cross sections and subjectively analyzed. Examination of the rawinsonde data revealed balloon drifts of up to 4 km at 800 mb and 10 km at 500 mb along the orientation of the cross sections. Since our interest lies in the mesoscale structure of the lower troposphere (i.e., in the vicinity of the frontal zone and low-level jet), and the balloon drifts at low levels were small in comparison to the rawinsonde station spacing, drift corrections were not performed.

a. Frontal structure during the midafternoon

Conditions across the surface front during the midafternoon, prior to the arrival of MCC 2, are shown in Fig. 10 (location denoted by line AA' in Fig. 1). At this time, the leading edge of the front was located near Enid, Oklahoma (END). Surface winds shifted from southerly to easterly across the shallow front (Fig. 10a), which extended to 150 mb above the surface at the northern extent of the PRE-STORM rawinsonde domain. Although shallow in depth, the front exhibited a strong surface potential temperature θ contrast of approximately 10 K (100 km)^{-1} beginning 100 km behind the leading edge of the surface wind shift. The displacement of the most intense surface θ gradient to a position well north of the surface wind shift arose from the modification and temporary erosion of the relatively shallow, cool postfrontal air mass by strong surface heating between END and Pratt, Kansas (PTT). This displacement remained intact several hours later in the vicinity of MCC 2 at maturity (Smull and Augustine 1993, cf. their Fig. 26). North of PTT, cool air (θ ≤ 294 K) was well entrenched as low clouds and fog precluded any significant surface heating.

South of the front, the boundary layer was about 200 mb (~2 km) deep and approximately well mixed in θ. Analysis of θ_e (Fig. 10b) does, however, indicate that the boundary layer was not uniformly well mixed in specific humidity as surface values during the midafternoon varied from approximately 360 K at END, in the vicinity of the front, to about 348 K at Norman,
Fig. 10. Vertical cross sections oriented from 346° (left) to 164° (right), whose location is given by AA′ in Fig. 1, in which (a) potential temperature θ is subjectively analyzed in 4-K increments (solid lines), with selected 2-K increments (dashed) added to help better delineate frontal structure. Horizontal winds are plotted with the standard meteorological convention used in Fig. 3, (b) equivalent potential temperature θ e is subjectively analyzed in 4-K increments, and the meridional component of the horizontal wind is subjectively analyzed in 5 m s⁻¹ increments using PRE-STORM rawinsonde and mesonet observations at 1500 CST 3 June 1985. The arrow underneath (a) represents the approximate location of the leading edge of the surface front. The locations of the projections of surface station and sounding sites are denoted at the bottom of (b).

Oklahoma (OUN), located approximately 150 km south of the front. A pronounced increase in θ and decrease in θ e with height above the top of the boundary layer (800–775 mb) in the southern portion of the cross section (at right), were associated with a persistent temperature inversion, characterized by dry air above it, that was not present to the north near the surface frontal boundary. Moderate low-level southerly flow was present south of the front and above the frontal surface during the afternoon (Fig. 10b); however, a concentrated maximum of the southerly wind component in the vertical was not present south of the front.

b. Frontal structure during the late evening

Examination of the frontal structure during the late evening, after the passage of MCC 2 and prior to the organization of the nocturnal MCC (system 4), reveals significant changes from the pre-MCC conditions that were present during the midafternoon. At 2230 CST 3 June (Fig. 11a; location denoted by line BB′ in Figs. 1, 6a), the surface frontal zone was characterized by a less intense horizontal potential temperature gradient than was apparent during the afternoon (Fig. 10a). A primary cause for the weakening of the potential temperature contrast across the front during the evening hours was the diurnal cooling present near the surface in the warm sector that was not occurring in the nearly saturated boundary layer north of the front. The development of a shallow surface-based stable layer within the warm sector (from diurnal cooling) and the subsidence warming above the front, in the wake of MCC 2 (Stumpf et al. 1991), are factors that also led to the observed shallower slope of the frontal surface and the more wedge-shaped appearance of the cold air mass at night (cf. Figs. 10a and 11a).

A second striking difference between the midafternoon and late-evening cross sections is the presence of a sloping maximum in the southerly wind component that originated near the surface within the warm sector.

Fig. 11. As in Fig. 10 but oriented from 360° (left) to 180° (right), with location of the cross section given by BB′ in Fig. 1, at 2230 CST 3 June 1985. The shaded region in (b) denotes locations where the southerly wind component exceeds 15 m s⁻¹.
and rose gradually to the 800-mb level above the frontal surface, approximately 300 km north of the surface frontal boundary (Fig. 11b). This low-level jet was responsible for advecting a tongue of warm, moist air ($\theta_e \geq 346$ K) toward the region that had been stabilized by pronounced drying associated with mesoscale downdrafts in the wake of MCC 2 (Stumpf et al. 1991; Fortune et al. 1992; Smull and Augustine 1992). Depressed values of $\theta_e \leq 314$ K were still present near 800 mb at the northern edge of the cross section near RSL (Fig. 11b), while the zone 100 to 200 km to north of the surface frontal position was beginning to moisten with the arrival of the low-level jet. Two hours later, deep convection developed locally near the terminus of the jet [between Woodward, Oklahoma (WWR) and PTT] as part of the rapid eastward expansion of MCC 4, within this zone north of the leading edge of the surface front (Figs. 5 and 6a,b).

c. Structure and evolution of the low-level jet

Hodographs derived from rawinsonde data afford a more detailed depiction of the low-level jet structure. The evolution of the 0–5-km wind at Fort Sill, Oklahoma (FSI), located 100–150 km south of the surface front, is followed from early evening until 2 h before sunrise in Fig. 12. The first evidence of a low-level wind maximum in the vertical appears at 1810 3 June, with a 12 m s$^{-1}$ wind speed located at 900 m (539 m AGL, Fig. 12a). By midevening (2115 CST), the low-level jet had intensified to 15 m s$^{-1}$ (Fig. 12b) and had developed a considerably more pronounced vertical-shear profile as both the near surface winds below the jet and the 2.3–3.0-km winds above the jet had decreased markedly during the early evening. During the next 6 h (Figs. 12c,d) the jet maximum continued to increase in strength while it gradually veered from a direction of 172° at 2115 3 June to 202° by 0230 4 June. Rawinsonde observations from other stations within the warm sector of the 400-km-wide PRE-STORM network showed similar speed increases and veering overnight with maximum low-level wind speeds of 21 m s$^{-1}$ observed at Henryetta, Oklahoma (HET).

The 0000 4 June hodograph from END (Fig. 13, solid curve), located approximately 75 km north of the frontal boundary, indicates strong easterly flow in the 500-m layer above the surface. A southerly jet of 14 m s$^{-1}$ was located above the shallow easterlies, centered at 1350 m ($\sim 1000$ m AGL). The height of the maximum wind was 300 m higher than that observed south of the front at FSI (Fig. 12c). At PTT (Fig. 13, dashed curve), 200 km north of the front, the southerly jet maximum above the easterlies was weaker (11 m s$^{-1}$) and was located an additional 300 m higher at 1650 m (1050 AGL, 835 mb).

The comparison of the 0000 CST 4 June hodographs (Figs. 12c, 13) reveals several key features of the low-level jet prior to the local development of convection over southcentral Kansas. Specifically, 1) the axis of maximum wind within the jet gradually sloped upward at 2 m km$^{-1}$ from south to north above the frontal surface, 2) by midnight, the high $\theta_e$ (343–350 K) values within the core of the jet showed little variation from south to north, and 3) wind speeds within the jet core

![Fig. 12. Hodographs from Fort Sill, Oklahoma (FSI), at (a) 1810 CST 3 June, (b) 2115 CST 3 June, (c) 0010 CST 4 June, and (d) 0230 CST 4 June 1985. The wind speeds at heights (km MSL) are plotted in meters per second along with values of equivalent potential temperature $\theta_e$ in kelvins. The azimuth and speed of the peak wind associated with the low-level jet is indicated by the X.](image-url)
decreased significantly (3–7 m s⁻¹) from south to north above the frontal surface. The marked south-to-north decrease in wind speed with the jet has been discussed in earlier studies (e.g., Pitchford and London 1962; Bonner 1966) as often being associated with horizontal convergence that leads to nocturnal convective development. The role of horizontal convergence on the evolution of the environmental conditions above the frontal surface shall be examined in the following section.

The development, intensification, and pronounced veering of the jet during the night along with the relatively low height of the jet axis above ground level in the warm sector are characteristics that are consistent with the mechanism proposed by Blackadar (1957) in which a strong wind maximum develops near the top of a nocturnal inversion due to the cessation of strong turbulent mixing in the boundary layer. The veering of the jet was explained by Blackadar to result from an inertial oscillation of the ageostrophic wind component that arises through this process. The sounding from FSI at 2115 3 June (Fig. 14) shows that the surface-based nocturnal inversion existed during the period of most rapid jet intensification. At this time, the maximum winds associated with the jet (Fig. 12b) were located at 700 m AGL (895 mb), approximately 350 m above the top of the inversion. Although the nocturnal inversion exhibited spatial and temporal variability in both strength and depth, the jet continued to intensify throughout the remainder of the evening at all PRE-STORM rawinsonde stations south of the surface front. Previous studies (e.g., Holton 1967; Bonner and Paegle 1970; McNider and Pielke 1981; Fast and McCorcle 1990) have also indicated that jet development may be enhanced by the differential cooling of the sloping terrain of the Great Plains.

Other studies (e.g., Uccellini and Johnson 1979; Uccellini 1980) have shown that the low-level jet may occur as part of a coupled circulation associated with the passage of an upper-tropospheric jet streak. During the 3–4 June episode, a 35–45 m s⁻¹ jet streak moved through the quasi-stationary 300-mb trough, located over the southwestern United States (Fig. 15). The increase of the low-level southerlies began as the exit region of the upper-level jet approached the western portion of the PRE-STORM network at 1800 3 June (Fig. 15b); however, the observed acceleration of the low-level flow toward the upper-tropospheric jet streak ceased by 2100 3 June, before the low-level jet had reached maximum intensity. As discussed earlier, the low-level jet exhibited pronounced veering overnight so that by sunrise the low-level winds were nearly parallel to the major axis of the upper-tropospheric jet (Fig. 15c). This evolution is in contrast to the development of a strong component perpendicular to the upper-tropospheric jet expected in cases where coupling of upper- and lower-tropospheric jet streaks (Uccellini and Johnson 1979) is observed. During the 3–4 June case, the juxtaposition of the subtropical high located over the eastern Gulf of Mexico and the cutoff low situated over the southwestern United States (Fig. 3) provided a synoptic environment of strong southerly flow. Hence, it appears that boundary-layer processes (e.g., surface radiative cooling) were of primary importance in the development of a strong nocturnal low-level jet within this preexisting, strong low-level flow regime.

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**Fig. 13.** As in Fig. 12 but for Enid, Oklahoma (END), at 2356 CST 3 June (solid) and Pratt, Kansas (PTT), at 0000 CST 4 June 1985 (dashed).

**Fig. 14.** Skew T-logp diagram of temperature and dewpoint curves (°C) for Fort Sill, Oklahoma (FSI), at 2115 CST 3 June 1985. The winds are plotted at the side with the standard meteorological convention used in Fig. 3.
terminus of the low-level jet, in which the nocturnal MCC (system 4) later experienced its most explosive growth.

\textit{a. Northward transport of warm, moist air by the low-level jet}

The hodographs and $\theta_e$ data demonstrate the importance of the low-level jet in transporting high-$\theta_e$ air, heated in the daytime boundary layer south of the front, northward above the frontal surface. At the time of the low-level jet onset south of the frontal boundary, $\theta_e$ at the jet level ranged from 347 K at FSI (Fig. 12a) to 351 K at Sulphur, Oklahoma (SUL, not shown). Assuming a mean speed of 15 m s$^{-1}$, this air would be expected to arrive 350 km to the north in about 6.5 h. Although this simple calculation requires the assumption of steady conditions and no turbulent mixing in the region of low static stability above the front (both of which are violated to some degree), the values of $\theta_e$, observed in the jet core at PTT 6 h later (at midnight), of 343–351 K (Fig. 13) are remarkably similar to those observed earlier at the jet level over southern Oklahoma and are in significant contrast with the 330-K values below the frontal surface. This transport led to dramatic local changes in potential instability ($\partial \theta_e / \partial z < 0$) above the frontal surface. Vertical time series of

4. Thermodynamic destabilization over the MCC growth region

The data presented in the previous section were used to document the mesoscale structure and evolution of the shallow quasi-stationary front and the diurnally varying low-level jet. In this section, we further examine how several factors, including 1) the transport of high-$\theta_e$ air by the low-level jet, 2) mesoscale lifting, and 3) the presence of midtropospheric baroclinicity, acted together to produce the environment north of the quasi-stationary, surface frontal boundary, near the northern

\textbf{Fig. 15.} 300-mb analysis with geopotential height contours (solid) analyzed in 12-dam intervals with the 35 and 45 m s$^{-1}$ isotachs (dashed) at (a) 0600 CST 3 June, (b) 1800 CST 3 June, and (c) 0600 CST 4 June 1985. The bold arrow schematically illustrates the approximate streamline of the low-level jet in (b) and (c).

\textbf{Fig. 16.} Vertical profiles of equivalent potential temperature $\theta_e$ and horizontal winds (plotted with the standard meteorological convention used in Fig. 3) from soundings taken at Pratt, Kansas (PTT), at 2100 CST 3 June, 2230 CST 3 June, and 0000 CST 4 June (2400 CST 3 June) 1985.
\( \theta_e \) from PTT (Fig. 16) [in the path of the precipitation maximum associated with MCC 4 (Fig. 9)] shows that an increase of 10–20 K occurred in a 130-mb-deep layer above the front in a time period of 1.5 h or less, prior to the local development of deep convection.

A comparison between vertical profiles of temperature and dewpoint (Fig. 17a) for 2100 3 June (dotted) and 0000 4 June (dashed) reveals that the low-level \( \theta_e \) change was primarily due to large increases in moisture. This evolution is consistent with the pronounced 850-mb moisture advection that was occurring in the vicinity of PTT at 2230 3 June (Fig. 18a). The warm and dry conditions present from 700 mb to just above the surface at 2100 3 June (Fig. 17a) resulted from the previously discussed mesoscale unsaturated downdraft that occurred in the wake of MCC 2. The anomalously warm conditions occupied a northeast–southwest swath at 850 mb that, in addition to PTT, included the stations of IAB and Fort Riley, Kansas (FR1) (Fig. 18a). Under these warm and dry conditions the most unstable air parcel above the surface front possessed only 20 J kg\(^{-1}\) of convective available potential energy (CAPE) and had substantial (340 J kg\(^{-1}\)) convective inhibition to overcome beneath its LFC. Three hours later, after the arrival of the moist air, CAPE for air parcels lifted from within the 100-mb layer above the surface front ranged from 1200 to 3570 J kg\(^{-1}\), while convective inhibition had decreased to only 20–40 J kg\(^{-1}\). For elevated thunderstorms that occur north of quasi-stationary or warm fronts, it is often difficult to be certain of the exact source level of air parcels that constitute the deep convective updrafts. Averaging the temperature and mixing ratio through a 50-mb layer immediately above the frontal surface yields a CAPE of 2260 J kg\(^{-1}\) and a lifted index (Galway 1956)\(^3\) of \(-7^\circ\)C, the latter representing a decrease of 8^\circ\)C during the 3-h period!

While the region north of the quasi-stationary frontal boundary in advance of the incipient nocturnal MCC (system 4) was rapidly becoming more susceptible to deep convection, conditions were becoming less favorable south of the frontal zone. Diurnal cooling of the boundary layer during the evening at SUL (Fig. 17b), along with low-level drying of unknown origin, was responsible for a decrease in the CAPE from 2830 to 1910 J kg\(^{-1}\) for air parcels averaged through the lowest 50 mb. Since a CAPE of 1910 J kg\(^{-1}\) is still sufficient to support intense convective activity, it is perhaps more important to note that the low-level cooling and drying led to an increase of convective inhibition for the ascent of boundary-layer parcels from 35 to 80 J kg\(^{-1}\) while raising the LFC 40 mb to 790 mb during the 3-h period. By 0252 4 June (3 h later), the LFC had risen an additional 45 mb.

The line-normal vertical wind shear in advance of the north–south-oriented squall-like convective feature that comprised the leading edge of MCC 3 (Fig. 6a,b) decreased from approximately 11 to 7 m s\(^{-1}\) in the lowest 2.5 km AGL at FSI as the low-level jet veered during the 3-h period from 2115 3 June to 0010 4 June (Figs. 12b,c). Line-perpendicular shear of less than 10 m s\(^{-1}\) was considered by Weisman et al. (1988) to constitute only weak forcing for updrafts at the leading edge of a storm-induced cold pool or gust front. Farther east at SUL, the magnitude of the line-perpendicular shear dropped more significantly from 10.5 to 2.5 m s\(^{-1}\) in the lowest 2.5 km AGL during the 6-h period from 2052 3 June to 0252 4 June. Thus, the likely occurrence of the lifting at the leading edge of the squall-like convection, becoming less favorable with time, together with the increase of the LFC height (i.e., deeper lifting required for free convection to be realized), is consistent with the observed dissipation of convection south of the surface front overnight (Figs. 5 and 6).

b. Horizontal convergence and vertical motion

Mesoscale horizontal convergence and ascent have previously been shown to be present in precursor MCC environments (e.g., Maddox 1983; Cotton et al. 1989). We now examine the influence of mesoscale vertical motion on convective development in the regions both north and south of the surface front. Average horizontal divergence for the mesoscale region north of the surface front bounded by END, PTT, and Wichita, Kansas (IAB) was obtained using the technique described by Bellamy (1949) on rawinsonde wind data interpolated to constant pressure surfaces every 25 mb. Kinematic vertical motion in pressure coordinates \( \omega = dp/dt \) was subsequently obtained by integrating the continuity equation and applying an adjustment to the horizontal divergence (Fankhauser 1969; O'Brien 1970), which ensured that \( \omega = 0 \) at both the lowest (940 mb) and highest (140 mb) pressure levels in the vertical profiles. The adjustment was linear with the largest corrections applied at the top of the profile where pressure was lowest. Vertical profiles of \( \omega \) for the 3-h period preceding the local development of deep convection, within the region near the northern terminus of the low-level jet (Fig. 19), indicate that the advection of high-\( \theta_e \) air above the frontal surface was accompanied by mesoscale lifting in the lower and middle troposphere. Peak ascent of \(-9 \text{ to } -11 \text{ } \mu \text{b s}\(^{-1}\) \((11-14 \text{ cm s}\(^{-1}\)) was situated in the lower troposphere between 800 and 750 mb, approximately 50 mb above the jet axis.

The lower-tropospheric minimum of \( \omega \) prior to MCC development is consistent with Maddox's (1983) composite results. The level of maximum horizontal convergence for this case was located in the vicinity of the vertically sloping frontal surface between 850 and 900

\(^3\) Galway (1956) defines the lifted index as the difference between the ambient 500-mb temperature and the 500-mb temperature of an undiluted air parcel that has been lifted from a modified layer within the lowest 3000 ft AGL. Here we make no such restriction for the source level of the parcel.
mb. The additional 100-mb depth of convergence located above the front was largely associated with the meridional variation of the wind speed near the terminus of the low-level jet (Figs. 11b and 13). An elevated peak in environmental horizontal convergence prior to MCC development was not resolved in the composite study by Maddox (1983). This difference may be due to a combination of factors such as coarser spatial and temporal resolution, the averaging inherent in compositing, and the possibility that MCC environments other than the frontal type described by Maddox et al. (1979) and Kane et al. (1987) may have also been incorporated into the composite. The onset of this MCC near midnight, 2–3 h later than average (see, e.g., Maddox et al. 1986, Figs. 17.4 and 17.5), may also contribute to the differences in pre-MCC convergence profiles since a well-developed nocturnal inversion was doubtfully present in the Maddox (1983) pre-MCC composite that utilized observations taken earlier in the diurnal cycle at 1800 CST (0000 UTC). An elevated peak in horizontal convergence was present in the Cotton et al. (1989) MCC composite study during the period 12 h prior to MCC development. However, maximum horizontal convergence in their study was situated at 800 mb, 50–75 mb above the location of the peak in this case.

While the low-level mesoscale convergence diagnosed from the END–PTT–IAB rawinsonde triangle north of the surface front (location given by the northern triangle in Fig. 1) was 200 mb deep, the low-level mesoscale convergence obtained from the FSI–HET–SUL rawinsonde triangle south of the surface front (location given by the southern triangle in Fig. 1) was only 50 mb deep and was overlaid by a 200-mb-deep layer of horizontal divergence (Fig. 20). The horizontal divergence in the 900–700-mb layer and associated descent above 850 mb is consistent with the pronounced diurnality (Fig. 18a) and the 900–700-mb drying (Fig. 17b) observed just above the low-level jet in the warm sector.

The horizontal distribution of low-level (775 mb) \( \omega \) prior to the passage of MCCs 3 and 4 across the PRE-STORM rawinsonde network is presented in Fig. 21. The 45 Bellamy triangles used in the preparation of this analysis were required to have areas less than 25 000 km\(^2\), and possess no angles less than 15\(^\circ\). These constraints eliminated both extremely large and highly oblique triangles that could yield unrepresentative values of the horizontal divergence used to calculate \( \omega \). The termination of several of the soundings below the tropopause necessitated that \( \omega \) be calculated without employing the adjustment scheme used in Fig. 19. Although errors in horizontal divergence that accumulate during an upward integration of the mass continuity equation can significantly affect \( \omega \) by the middle and upper troposphere, the errors in \( \omega \) are expected to be small at 775 mb (150–200 mb AGL). A check of \( \omega \) values diagnosed from triangles that were composed of complete soundings revealed that the differences in magnitude between adjusted and unadjusted profiles at this level did not exceed 1.6 \( \mu \)b s\(^{-1}\) and were in general less than 0.5 \( \mu \)b s\(^{-1}\).
The analysis shows a broad zone of ascent straddling the Oklahoma and Kansas border with minimum values of ω located over south-central Kansas in the vicinity of the END–PTT–IAB triangle centroid (Fig. 19). South of the surface front, descent was widespread over central Oklahoma above the low-level jet. A second area of descent over the northeastern portion of the rawinsonde domain extended rearward from the region of precipitation associated with MCC 2 (see inset of Figs. 19a,b). It should be noted that the westward extent of this zone of descent could possibly be overestimated by up to approximately 75 km, since at this time, only the easternmost rawinsonde stations in Kansas [FRI and Chanute, Kansas (CNU)] were still being influenced by the low-level divergent flow present near the back edge of the dissipating MCC. A separate analysis based on 12 independent triangles (not shown) yielded the same three zones of ascent and descent with only minor differences in scale and extrema.

Although there was significant moistening at stations north of the surface front along and above the axis of the jet during the period of thermodynamic destabil-
zation, the air remained subsaturated during this period. Thus, the kinematic diagnosis of mesoscale ascent north of the surface front is qualitatively consistent with the two-dimensional depiction of the southerly jet core sloping northward approximately parallel to the isentropic surfaces in cross section BB' (Fig. 11). However, a more detailed analysis reveals that the mesoscale ascent just above the frontal surface prior to the outbreak of convection did not correspond to simple upglide along a stationary pattern of isentropes as is often assumed in cases of warm-frontal overrunning. Estimates of $\omega$ based on the assumption of upglide were made along several sample isentropes located at and slightly above the jet level along BB' between WWR and PTT. For these estimates, we assumed that $w = V(\partial H/\partial s)$, where $V$ is a mean horizontal wind speed along the streamline, with $\omega \sim -\rho g w$. The $\omega$ values of $-1$ to $-3 \mu b \, s^{-1}$ yielded by these calculations were of the same sign but significantly weaker than the kinematically diagnosed values, which fell within the range of $-4$ to $-6 \mu b \, s^{-1}$ at similar levels in the vicinity of WWR and PTT (e.g., Fig. 21).

This comparison implies that during the period of thermodynamic destabilization prior to convective development, the actual trajectories of air parcels moving northward above the frontal surface were more steeply sloped than the instantaneous isentrope pattern within this region near the northern terminus of the low-level jet. South of the surface frontal boundary, the slope of the isentropic surfaces near the jet level exhibited greater east-to-west variation, presumably due to local differences in the strength and depth of the nocturnal inversion. However, in general, the weak ($-1$ to $-2 \mu b \, s^{-1}$), kinematically diagnosed ascent just above the nocturnal inversion ($875$–$900$ mb), south of the surface front (Fig. 20), more closely matched the magnitude of the ascent expected from assuming upglide of parcels along the instantaneous isentrope pattern.

If possible diabatic effects are neglected, the $1.5^\circ C$ $850$–$670$-mb cooling during the 3-h period from $2100$ 3 June and $0000$ 4 June north of the surface front at PTT (Fig. 17a) is consistent with the local cooling expected in a statically stable environment where the inclination of parcel trajectories is greater than that of the instantaneous isentrope pattern. This observed cooling was mesoscale in nature, as comparable $1.0$–$2.0^\circ C$ 3-h temperature falls were also observed in similar layers over the north-central Oklahoma and south-

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**Fig. 19.** Vertical profiles of average vertical $p$ velocity $\omega$ (dashed) and horizontal divergence (solid) for the mesoscale region bounded by the rawinsonde stations of Enid, Oklahoma (END), Pratt, Kansas (PTT), and Wichita, Kansas (IAB), denoted by dashed triangle (centroid marked by a cross) in the inset at (a) 2100 CST 3 June, (b) 2230 CST 3 June, and (c) 0000 CST 4 June 1985. Also within the inset, the position of the quasi-stationary surface front is indicated by the bold barbed line, and the locations of radar echo are indicated by the light shading.
central Kansas region at END and IAB (not shown). This situation contrasts with the steady case in which the cooling associated with upglide along a stationary pattern of isentropes is exactly balanced by horizontal advection such that no local temperature change results.

The thermodynamic energy equation expressed in isobaric coordinates

\[
\frac{\partial T}{\partial t} = -V \frac{\partial T}{\partial s} + \omega \left( \frac{kT}{p} - \frac{\partial T}{\partial p} \right) + \frac{Q}{c_p} \tag{1}
\]

may be used to quantify the relationship of the pronounced mesoscale ascent north of the surface front to the observed 850–670-mb cooling. In this equation, \( V \) is the horizontal wind component along the streamline \( s \); \( T \) is the temperature, \( k = R/c_p \), where \( c_p \) is the specific heat at constant pressure \( p \); \( R \) is the gas constant; and \( Q \) is the diabatic heating rate. In (1), local temperature changes \( (\partial T/\partial t) \) at constant pressure may be associated with horizontal temperature advection \( -V(\partial T/\partial s) \), adiabatic vertical motion \( \omega(kT/p - \partial T/\partial p) \), and diabatic effects \( (Q/c_p) \).

The range of 775 mb \( \omega \) near PTT of \(-4 \) to \(-6 \) \( \text{mb s}^{-1} \) (Fig. 21), taken with the lapse rate of \( \partial T/\partial p = 7.6^\circ \text{C} \) \( (100 \text{ mb})^{-1} \) present within the 850–700-mb layer of the 2230 3 June PTT sounding, yields a range of possible adiabatic cooling rates of \( 1.3^\circ \text{C} - 2.0^\circ \text{C} \) \( (3 \text{ h})^{-1} \). Since the deep cloudiness and precipitation associated with MCC 4 did not arrive or develop locally at PTT until after 0000 4 June (Fig. 5), the occurrence of evaporative cooling was not likely. Thus, the contribution to the diabatic term was likely dominated by longwave radiative cooling. Climatological studies (e.g., Dopplick 1972) have indicated that away from the earth’s surface at 775 mb, the radiative cooling rate is generally small, approximately \( 0.2^\circ \text{C} \) \( (3 \text{ h})^{-1} \). Hence, the diabatic contribution may be neglected when compared to the adiabatic cooling in this particular case. The 850- and 700-mb analyses for 2230 3 June (Figs. 18a,b) suggest that the warm advection within the layer of observed cooling was quite weak in the vicinity of PTT. The fact that the zone of strong warm advection at 850 mb was confined to the north of PTT (Fig. 18a) is due to the lingering presence of anomalously warm temperatures in the wake of MCC 2.

Our consideration of the terms in (1) has demonstrated that the observed \( 1.5^\circ \text{C} \) cooling at 775 mb occurred primarily in association with mesoscale adiabatic ascent. The cooling throughout the 150-mb-deep layer above the jet axis was significant because it further reduced the convective inhibition for the most conditionally unstable air, situated slightly below the jet axis (Fig. 17a, dashed curves), which was nearly saturated at 0000 4 June as a result of lifting during its northward transport above the shallow frontal surface. This erosion of convective inhibition near the northern terminus of the low-level jet facilitated the rapid development of the convection shortly after 0000 4 June without the necessity of strong surface forcing (e.g., local heating of the surface, localized lifting along a gust front).

In an analysis of the internal structure of MCC 4, Fortune et al. (1992) remarked that the discrete convective clusters that were present during the system's intensive growth stage were relatively unaffected by convective outflows. This point is evident in Fig. 6b, which shows the occurrence of the 300-km-wide zone of convective clusters in essentially undisturbed easterly surface flow north of the surface front. Note that the zone of convective clusters was nearly collocated with the region of 775-mb mesoscale ascent situated above the frontal surface 2.5 h earlier (cf. Figs. 6b and 21). In contrast, convective inhibition increased overnight in the region south of the front, where strong and deep mesoscale ascent was absent. In this region, localized surface forcing was clearly necessary for the maintenance of deep convection.

c. The influence of the midtropospheric temperature gradient

It is clear from the evolution of the PTT soundings (Figs. 16 and 17a) that the northward transport of moist, unstable air by the low-level jet led to dramatic
local increases of CAPE north of the surface front. The persistent south-to-north midtropospheric temperature gradient was an additional factor that led to the abundant conditional instability that characterized the preconvective environment over this region. The 500-mb conditions that were discussed earlier in section 2 are shown in greater detail in Fig. 18c. Prior to the development of nocturnal convection, the 500-mb temperature was 4°C cooler at PTT than at SUL, which is located approximately 200 km south of the frontal boundary. Thus, as the moist air was transported northward (Fig. 18a), it became situated under colder air aloft (Fig. 18c).

The unstable conditions generated north of the quasi-stationary front in the preconvective environment, therefore, resulted not only from overrunning of the shallow quasi-stationary front by the warm, moist air transported northward by the low-level jet but also from the jet underrunning (Means 1952) the warmer midtropospheric air that was present south of the front. At midnight, average CAPE for air parcels lifted from within the 50-mb-deep layer situated just above the frontal surface at PTT and IAB ranged from 2260 to 2530 J kg⁻¹ with lifted indices of −7.0° to −8.5°C. Calculations made with surface-based air parcels averaged through the lowest 50 mb from the FSI, HET, and SUL late-afternoon soundings (not shown), located 100–200 km south of the surface front, revealed similar layer averaged CAPEs of 2000 to 2600 J kg⁻¹ but less buoyancy at midlevels with lifted indices of −5.5° to −6.5°C. This difference was due in part to the warmer midtropospheric conditions present south of the surface front. This comparison demonstrates that the northward transport of high-θₑ air at night by the low-level jet into regions that are significantly colder aloft fosters an environment, north of surface quasi-stationary or warm fronts, that may be characterized by even greater buoyancy above the frontal surface than the buoyancy present south of the surface front during the afternoon, where strong boundary-layer heating often occurs.

5. Summary and discussion

In this study we have described the evolution of environmental conditions that facilitated the development of a nocturnal MCC. The particular case studied was the last in a series of four MCCs that occurred over the southern Great Plains of the United States during the 3–4 June 1985 period of the PRE-STORM field experiment. Characteristics of the environment in advance of the incipient MCC were similar to those found in previous composite studies (Maddox, 1983; Cotton et al., 1989). The salient features included a low-level jet that intensified and veered during the night in the vicinity of a quasi-stationary surface frontal boundary and lower-tropospheric horizontal convergence and ascent within a mesoscale region that possessed substantial convective instability for air parcels lifted from near the jet level. This region also exhibited a broad north-south horizontal temperature gradient in the middle troposphere.

The fortuitous location of convective initiation at the western edge of the PRE-STORM domain has enabled us to utilize an enhanced resolution rawinsonde network and surface mesonet to identify details of the pre-MCC environment that are not readily apparent from composite studies and clearly document how the process of mesoscale thermodynamic destabilization can occur at night over the southern Great Plains of the United States in advance of a developing MCC. Our analysis has indicated that:

- diurnal boundary-layer processes (e.g., surface radiative cooling) most likely initiated the transformation of the strong southerly flow south of a quasi-stationary frontal boundary during the afternoon into an elevated low-level jet during the evening;
- this low-level jet was in turn responsible for transporting high-θₑ air over a considerable distance north-
ward as it sloped gradually upward above the shallow wedgelike frontal surface;

- the shallow ascent above the frontal surface (near the low-level jet height) was replaced by stronger and deeper lifting near the jet’s terminus, 150–300 km north of the surface frontal boundary, where the MCC later experienced its most explosive growth;

- the rapid development of deep convection over this region was facilitated by a dramatic vertical destabilization above the frontal surface. Strong moisture advection by the low-level jet was responsible for an increase from almost no CAPE to maximum CAPE of approximately 3000 J kg$^{-1}$ or greater in a period of 3 h or less. Midtropospheric buoyancy within this region actually exceeded daytime values south of the surface front, since the air within the core of the low-level jet originated in the heated boundary layer south of the front and was transported northward into a region possessing significantly colder midtropospheric temperatures. The extreme local destabilization rate was partly a consequence of the very stable preexisting conditions occurring in the wake of the second MCC in this series.

Several of these features are illustrated in a schematic 400–km cross section (Fig. 22) that represents conditions across the frontal zone just prior to the eruption of the convective clusters over the mesoscale region displayed in Fig. 6b.

Analyses of the second MCC of this series during 3–4 June (e.g., Stumpf et al. 1991; Fortune et al. 1992; Smull and Augustine 1993) have demonstrated that similar developments of large MCCs to the north of quasi-stationary surface boundaries need not be restricted to the evening and overnight hours if strong southerly geostrophic flow is present. However, a distinct nocturnal bias of frontal events is apparent from the composite studies of Maddox et al. (1979) and Kane et al. (1987). The additional differential advection and horizontal convergence afforded by enhancement of the low-level southerlies at night provide a nocturnal mechanism for destabilizing mesoscale regions to deep convection (e.g., Means 1952; Bonner 1966). It is also possible that other factors such as cloud-top radiative cooling in the upper troposphere (e.g., Chen and Cotton 1988) might aid in the persistence of such systems at night.

In this particular case we also observed convection south of the surface frontal boundary. Here, the local onset of precipitation was accompanied by a sharp surface wind shift and a pronounced drop in $\theta_v$, characteristic of squall lines sustained by lifting along their outflow boundary or gust front. The “squall-like” convection persisted through the evening hours but then rapidly dissipated overnight. Unlike the region north of the front, deep and strong mesoscale ascent was not present in this environment. As surface cooling continued overnight, localized forcing at the gust front was evidently no longer able to lift boundary-layer parcels to their LFC. North of the surface front, the cooling above the axis of the low-level jet that arose from mesoscale ascent, helped reduce convective inhibition, which aided the development of deep convection in the apparent absence of significant surface forcing. This

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**FIG. 22.** An idealized north–south schematic cross section depicting the thermodynamic and flow structure immediately prior to the development of deep convection above the wedge-shaped cold air mass (lightly shaded). The vectors represent the flow in the $x$–$z$ plane (vertical component is greatly exaggerated). The dot inside the circle represents easterly flow (out of the page) within the cool, moist air mass below the frontal surface. The dashed line represents the boundary of the warm, moist air mass transported northward above the frontal surface by the southerly low-level jet (LLJ) whose axis is denoted by the bold streamline.
comparison between the differing characteristics and evolution of convection on the northern and southern side of the surface frontal boundary emphasizes the importance of mesoscale ascent along with moisture advection in preconditioning mesoscale regions for the sudden development of deep convection and provides insight into why the zone north of quasi-stationary or surface warm fronts is a favored location for nocturnal MCC development.

Additional work is needed to fully understand the reason(s) for the 100-mb deep layer of horizontal convergence situated above the broad vertically sloping frontal zone that was associated with the enhanced mesoscale ascent. This horizontal convergence was associated with a marked decrease in wind speed from south to north along the axis of the jet (Fig. 1b). Horizontal convergence along the jet axis near its terminus is expected since the diurnal mechanism to locally generate supergeostrophic southerly momentum is absent north of the front, where a daytime boundary layer possessing southerly flow does not exist. However, studies of the low-level jet and its relationship to convective development (e.g., Pitchford and London 1962) have indicated that horizontal convergence along its axis can also occur in the absence of shallow frontal boundaries, suggesting that other factors may contribute to the convergence.

In this paper, we have examined a mechanism of in situ mesoscale forcing that is believed to be a substantial contributor to the warm-season nocturnal precipitation maximum over the central United States. Other modes of organized mesoscale convective development that occur over the nocturnal environment of the central United States include 1) convective systems that form near or over the elevated terrain in the vicinity of the Rocky Mountains that grow and persist overnight as they move into the unstable environment of the central Great Plains (e.g., Bonner 1966; Cotton et al. 1983; Wetzel et al. 1983), 2) "secondary" convective events described by Carbone et al. (1990) in which shallow disturbances initiated by prior upstream convective activity readily propagate in the statically stable nocturnal boundary layer and subsequently initiate deep convection upon encountering more favorable environmental conditions (e.g., increased low-level moisture, stronger vertical wind shear), and 3) both frontal (e.g., Menard and Fritsch 1989) and prefrontal (e.g., Johnson and Hamilton 1988; Rutledge et al. 1988) squall lines that often reach their peak convective intensities in the late afternoon or early-evening hours but may persist overnight in a slowly decaying state characterized by weakening convection at their leading edge followed by large regions of stratiform precipitation.

Although detailed observations have in general not been available, climatologies from other regions of the world (e.g., Velasco and Fritsch 1987; Miller and Fritsch 1991) have indicated that the majority of MCCs 1) are nocturnal, 2) form downwind of major mountain ranges, and 3) are associated with low-level jets of high-$\theta_v$ air. The common large-scale physiographic setting and kinematic and thermodynamic characteristics of these MCC environments suggest that mechanisms similar to the one described in this paper may operate to help promote MCC development in regions other than the central United States.

While some questions regarding the dynamics of this and other MCC environments remain, we are encouraged by the ability of a spatially and temporally enhanced resolution rawinsonde network to clearly resolve the rapid (1–3 h) mesoscale kinematic and thermodynamic changes preceding the development of a nocturnal MCC. The deployment of new high-resolution instrumentation over the central United States for operational purposes (e.g., Stensrud et al. 1990) will undoubtedly lead to the improved ability to monitor the rapid evolution of the mesoscale environment that often occurs prior to the development of widespread deep convection and, therefore, aid in the short-range forecasting of organized mesoscale convective systems.

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