Mechanisms Supporting Long-Lived Episodes of Propagating Nocturnal Convection within a 7-Day WRF Model Simulation

S. B. Trier, C. A. Davis, D. A. Ahijevych, M. L. Weisman, and G. H. Bryan

National Center for Atmospheric Research,* Boulder, Colorado

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ABSTRACT

A large-domain explicit convection simulation is used to investigate the life cycle of nocturnal convection for a one-week period of successive zonally propagating heavy precipitation episodes occurring over the central United States. Similar to climatological studies of phase-coherent warm-season convection, the longest-lived precipitation episodes initiate during the late afternoon over the western Great Plains (105°–100°W), reach their greatest intensity at night over the central Great Plains (100°–95°W), and typically weaken around or slightly after sunrise over the Midwest (95°–85°W). The longest-lived episodes exhibit average zonal phase speeds of \(\sim 20 \text{ m s}^{-1}\), consistent with radar observations during the period. Composite analysis of the life cycle of five long-lived nocturnal precipitation episodes indicates that convection both develops and then propagates eastward along an east–west-oriented lower-tropospheric frontal zone. An elevated \(\sim 2\text{-km-deep layer of high-}\theta_e\text{ air helps sustain convection during its period of greatest organization overnight. Trajectory analysis for individual episodes reveals that the high-}\theta_e\text{ air originates both from within the frontal zone and to its south where, in this latter case, it is transported northward by the nocturnal low-level jet (LLJ). The mature (nocturnal) stage composite evinces a thermally direct cross-frontal circulation, within which the trajectories ascend 0.5–2 km to produce the elevated conditionally unstable layer. This transverse vertical circulation is forced by deformation frontogenesis, which itself is supported by the intensification of the nocturnal LLJ. The frontal zone also provides an environment of strong vertical shear, which helps organize the zonally propagating component of convection. Overnight the convection exhibits squall-line characteristics, where its phase speed is typically consistent with that which arises from deep convectively induced buoyancy perturbations combined with the opposing environmental surface flow. In a large majority of cases convection weakens as it reaches the Midwest around sunrise, where environmental thermodynamic stability is greater, and environmental vertical shear, frontogenesis, and vertical motions are weaker than those located farther west overnight.

1. Introduction

Carbone et al. (2002) documented the common occurrence of successive time–space coherent episodes of heavy rainfall that travel across the eastern \(\frac{2}{3}\) of the United States during the warm season (May–August). The lifetime of individual episodes ranged from \(\sim 12\) h, which is consistent with that of a single long-lived mesoscale convective system (MCS), to up to 60 h, with the longest episodes typically comprising several sequential MCSs. These precipitation episodes propagate in the sense that their zonal speed is significantly greater than the environmental winds through most of the troposphere in which they are embedded. The episodes also exhibit phase coherence with respect to the diurnal cycle. Heavy precipitation typically occurs overnight in the central Great Plains (100°–95°W), reaches the Midwest (95°–85°W) around or slightly after sunrise, and typically either terminates or weakens temporarily from sunrise through early afternoon. The objective of the current study is to use a numerical simulation of a weeklong period of successive coherent propagating rainfall episodes to examine mechanisms supporting nocturnal convection and its subsequent decay after sunrise.

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A detailed theory of organized nocturnal convection does not currently exist, however its association with several common environmental factors has been widely documented. The Great Plains nocturnal southerly low-level jet (LLJ), which studies have linked to differential cooling of the sloped terrain (e.g., Holton 1967; Bonner and Paegle 1970) and the cessation of turbulent mixing within the planetary boundary layer (PBL) during the evening (Blackadar 1957), provides a source of conditionally unstable air for MCS development and sustenance (e.g., Maddox 1983; Cotton et al. 1989). Poleward nocturnal jets of thermodynamically unstable air in the lee of north–south (N–S)-oriented mountain ranges are a common environmental feature in the global distribution of MCSs (Laing and Fritsch 2000).

Maddox et al. (1979) found that a particular type of MCS that has a distinct nocturnal bias initiates on the northern or “cool” side of a quasi-stationary surface front downstream from the axis of maximum low-level (e.g., 900–850 hPa) flow. In a composite study of a 105 MCSs, Kane et al. (1987) found that roughly half of their cases occurred under the “frontal” conditions described by Maddox et al. In addition to transporting moist conditionally unstable air, Trier and Parsons (1993) found that the LLJ contributed to mesoscale horizontal convergence within the frontal zone, which helped further reduce convective inhibition. Augustine and Caracena (1994) noted that the frontogenetic character of the late afternoon lower-tropospheric flow, itself, is an important predictor of whether long-lived, intense nocturnal MCS activity will occur.

Tuttle and Davis (2006) established that extended (e.g., weeklong) regimes of eastward traveling convection over the central United States are often confined to narrow corridors (<5° latitude) where nocturnal LLJs impinge on quasi-stationary surface fronts oriented approximately parallel to moderate-to-strong midtropospheric westerlies. Such midtropospheric flows often contain embedded low-amplitude shortwave disturbances that may help trigger convection. Maddox (1983), however, recognized that lower-tropospheric conditions (most notably warm advection) were crucial in allowing organized convection to persist overnight.

Propagating nocturnal convection over the central United States is poorly forecasted by current operational numerical weather prediction (NWP) models (Davis et al. 2003). This is partly attributed to the inability of such models to explicitly resolve convection. On the other hand, research cloud models that permit explicit deep convection have historically relied on idealized horizontally homogeneous initial conditions. Such models have had difficulty simulating the life cycle of elevated convection (i.e., convection that is locally rooted above the PBL), which can occur above preexisting surface outflows (e.g., Schmidt and Cotton 1989) or fronts (e.g., Smull and Augustine 1993). Elevated convection is particularly common at night over the central United States due to both the surface stabilizing effect of radiative cooling and the frequent interaction of low-level jets and surface fronts, which can produce significant vertical displacements of conditionally unstable air.

Numerical models capable of explicitly resolving deep convection over large heterogeneous domains have recently become available as research tools. For example, Coniglio and Stensrud (2001) examined the life cycle of a single long-lived MCS simulated in a composite environment representative of that in which derechos (Johns and Hirt 1987; Johns 1993) form, while Jewett and Wilhelmson (2006) found that frontal forcing played a significant role in the organization of convective cells within developing MCSs.

Our approach is to use a numerical model with explicit deep convection over a heterogeneous domain of subcontinental scale to produce a weeklong simulation that captures successive episodes of long-lived propagating convection for a period during which similar heavy precipitation episodes were observed. The emphasis is on understanding the response of organized nocturnal convection to its environment, which itself varies both spatially and temporally (in part due to diurnally driven processes). This is addressed using composites of the structure and environment of five simulated long-lived nocturnal MCSs together with model-based trajectories, which collectively allow us to clarify the roles of environmental features such as surface fronts, the LLJ and their interaction on various stages of the convective life cycle.

2. Numerical model and experimental design

a. Case selection

Of interest in this study is the convection that occurred over the approximate eastern 7⁄8 of the United States from 0000 UTC 3 July to 0000 UTC 10 July 2003. This one-week period comprises two intensive observing periods (IOPs) of the Bow Echo and MCV Experiment (BAMEX; Davis et al. 2004). BAMEX was a mobile experiment that used both aircraft and ground-based observing systems to study bow-echoes associated with long-lived MCSs, and mesoscale convective vortices (MCVs) and their subsequent convection. Bow echoes most often occur in synoptic regimes of large CAPE and strong vertical shear (e.g., Weisman 1993; Przybylinski 1995) and frequently produce severe sur-
face winds. Several bow-echo-producing MCSs of long duration occurred during this one-week period. BAMEX IOP 18 (5–6 July) was devoted to the detailed study of a single bow echo described by Wakimoto et al. (2006a, b).

b. Numerical model and simulation methodology

A mass coordinate version (version 1.3) of the Weather Research and Forecasting Model (WRF), similar to that described by Skamarock et al. (2005), is integrated for the 168-h period over the single subcontinental-scale domain shown in Fig. 1. The model contains $625 \times 445$ horizontal grid points with 4-km grid spacing. Thirty-five vertical levels are used in a stretched vertical grid with spacing ranging from $\sim 100$ m near the surface, to $\sim 500$ m at 2.5 km AGL and $\sim 1$ km above 14 km. The modeling system employs the Yonsei University (YSU) PBL scheme (Noh et al. 2001), the Oregon State University (OSU) land surface model (Chen and Dudhia 2001), a longwave and shortwave radiation parameterization (Dudhia 1989), and a bulk microphysics parameterization based on Lin et al. (1983). No cumulus parameterization is used.

The initial condition and lateral boundary conditions (updated every 3 h) are obtained from Eta operational model analyses. We emphasize that our extended-duration simulation is not an attempt at deterministic prediction of any single convective system. Instead, our goal is to accurately simulate the statistical characteristics of mesoscale convection that occurs within a particular weeklong synoptic regime type commonly associated with regular long-lived episodes of zonally propagating convection. Inasmuch as the statistics of the modeled convection resemble those of the observed convection (section 3), the model dataset, with its internal dynamical consistency, may be used as a surrogate to study mechanisms responsible for evolution of these long-lived precipitation episodes and their constituent MCSs.

Of particular relevance in our simulation design is the selection of horizontal domain and grid spacing. Here, our choices represent a compromise that allows us to capture the initiation and most or all of the subsequent life cycle of precipitation episodes, while having sufficient horizontal resolution to accurately portray their overall mesoscale structure.

Weisman et al. (1997) noted that mesoscale aspects of organized convection, including overall structure, and mass and momentum fluxes, may be at least qualitatively reproduced using 4-km horizontal grid spacing. These findings were confirmed by Bryan et al. (2003), with the caveat that adequately resolving deep convective cells, which are the building blocks of MCSs, requires horizontal grid spacings of 1 km or less. With 4-km horizontal grid spacing, the WRF has an effective resolution of $\sim 7\Delta x = 28$ km (Skamarock 2004). Thus, the current simulation is best suited to examine mesoscale components of convective systems and their interaction with the larger-scale environment. A similar configuration of WRF was used for 36-h forecasts for field operations in BAMEX, where it was found to successfully forecast the degree of mesoconvective organization, and, in a qualitative sense, the dominant mode (e.g., supercells, bow echo, MCV) of organization (Done et al. 2004).

3. Overview of the simulated rainfall and comparison with observations

Following Carbone et al. (2002) we simplify the statistical analysis of simulated and observed rainfall by averaging hourly rain rates in meridional and zonal directions and plotting results on time versus longitude and time versus latitude (Hovmøller) diagrams. Accumulated rainfall is a model output variable saved every 30 min in our simulation. For observations we use the WSI Corporation NOWrad national composite of the Weather Surveillance Radar-1988 Doppler (WSR-88D) network at 15-min frequency and apply a single reflectivity–rain-rate relationship ($Z = 300R^{1.5}$), where $Z$ is reflectivity factor (mm$^6$ m$^{-3}$) and $R$ is rainfall rate (mm h$^{-1}$). These radar data, which are on a $\sim 2$-km horizontal grid, have significant limitations for quantitative precipitation estimation (Ahijevych et al. 2004). However, they are adequate for our purposes of analyzing geographical patterns and general propagation characteristics of the rainfall.

Time versus longitude diagrams of rainfall rate for
both the simulation (Fig. 2a) and the observations (Fig. 2b) reveal characteristic downward-sloping streaks indicative of eastward-moving heavy rainfall episodes. In some cases latitudinally averaged heavy rainfall is continuous along a streak, while in others it is intermittent with later redevelopment at a more eastern longitude as described by Carbone et al. (2002). The number of long-lived (t ≥ 12 h) rain streaks that travel a substantial distance eastward (≥10° longitude) in the simulation and observations are similar with a frequency of ~1 day⁻¹ during the one-week period. Moncrieff and Liu (2006) obtain similar results in an explicit simulation using a different model [the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5)] with different physical parameterizations, which suggests the importance of large-scale forcing on the simulated behavior of convection.

To objectively quantify coherence between any two locations on the time versus longitude diagrams, a cosine-rectangular weighting function (Ahijevych et al. 2001; Carbone et al. 2002) is rotated until maximum correlations are obtained. The length of the autocorrelation rectangle is ~7° longitude, which falls between the 3° and 12° longitude range used by Carbone et al. (2002). The results of these autocorrelation fits, with a minimum correlation of 0.35, and additional longevity (t ≥ 9 h), and propagation distance (zonal span ≥500 km) criteria required to define a streak, are superimposed on Fig. 2. A similar number of coherent rain streaks are defined by this method in the simulation (12) and the observations (14). Histograms of rain streak zonal speed indicate a similar frequency distribution in the simulation (Fig. 3a) and observations (Fig. 3b) with the majority of rain streaks exhibiting rapid zonal speeds of around 20 m s⁻¹.

The simulated rain streaks during this period are confined to a northwest–southeast (NW–SE)-oriented corridor of total (168 h) rainfall exceeding 81 mm extending from South Dakota into southern Michigan and Ohio (Fig. 4a). The overall location and magnitude of the heaviest rainfall is consistent with radar-observed values (Fig. 4b). Within this simulated corridor of heavy rainfall, maximum amounts extend from the eastern Minnesota/Iowa border eastward into Ohio, while in the observations maximum amounts (>243 mm, ~10 in) are somewhat larger and more concentrated over northern Indiana. These extreme amounts are associated with repeated heavy rains occurring over the same general location associated with both the weakening phase of traveling nocturnal convection around sunrise and afternoon redevelopment of convection. The precipitation corridor occurs at the southern edge of the strong large-scale 500-hPa geopotential height gradient, where moderately strong (15–20 m s⁻¹) time-averaged westerly flow and broad north–south baroclinity coexist (Fig. 1).

Embedded within the time-averaged midtropospheric westerlies are frequent short wave disturbances. Figure 5 presents an example of the association of midtropospheric short waves with the diurnal cycle of convection. At 2000 UTC 4 July (Fig. 5a), convection initiates over western South Dakota immediately down-
stream from a well-defined short wave. The midtropospheric trough overnight (Figs. 5b,c) is collocated with the convection that has organized upscale. Convection reorganizes the following afternoon in the vicinity of the remnant short wave located at 1500 UTC 5 July in western Michigan (Fig. 5d). Meanwhile, the next in a series of short waves is situated over eastern Wyoming (Fig. 5d). This short wave, which is weaker and located farther east (relative to the diurnal cycle) than that of the previous day (Fig. 5a) is associated with convection initiation over South Dakota at 0000 UTC 6 July (not shown).

As in the above example, the onset of the precipitation episodes occurs within several hundred km of the western cordillera and roughly in phase with the maximum cumulative diurnal heating. This is consistent with the general behavior of long-lived precipitation episodes in warm-season climatological studies (e.g., Carbone et al. 2002). However, for the current one-week period, there are moderate spatial and temporal differences of several degrees of longitude and hours, respectively, in the onset of the precipitation episodes (Fig. 2b) that appear related to differences in the strength and timing of the midtropospheric short waves. Eta Model analyses (not shown) reveal that the coherent, mobile short waves (with frequency of $\sim 1$ day$^{-1}$) originate west of the Pacific coast and up to a day in advance of the onset of each precipitation episode, and are thus not directly generated by the elevated heat source that the cordillera provides.

The simulated convection (Figs. 6a–e) at its stage of greatest overall organization (referred to hereafter as the mature stage) is compared with concurrent composite radar observations (Figs. 6f–j) for five cases. These selected precipitation episodes represent the subset of phase-coherent rain streaks (labeled C1–C5 in Fig. 2a) that initiate during the afternoon or early evening over the western or central Great Plains ($105^\circ$–$100^\circ$W) and persist until around sunrise with either some weakening or termination occurring shortly thereafter, as has been similarly documented in previous climatological studies (e.g., Carbone et al. 2002, their Fig. 12; Ahijevych et al. 2004).

A common aspect of these five episodes of simulated convection is the tendency for mature stage nocturnal
convection to occur north of a quasi-stationary east-west (E–W)-oriented surface front and LLJ, which themselves vary in strength among the cases (Figs. 6a–e). Although the simulated convection contains less trailing stratiform precipitation than observed, N–S-oriented and E–W-oriented convective bands occur in both the simulations and observations and sometimes constitute components of the same precipitation episode, consistent with previous MCS case studies (e.g., Smull and Augustine 1993). The length of the N–S-oriented convective lines is sometimes greater in the observations. This difference is most pronounced in case 4 (Figs. 6d,i) and could be a manifestation of finescale triggering mechanisms (such as a dryline feature extending southwestward from the E–W-oriented frontal zone in that particular case) not always being adequately resolved with 4-km horizontal grid spacing.

In general, the N–S-oriented convection has a substantial component of vertical shear perpendicular to its major axis, consistent with its orientation approximately normal to the baroclinic zone. The E–W-oriented bands are oriented primarily parallel to the vertical shear and exhibit structure similar to the “training-line-adjoining stratiform” archetype (Schumacher and Johnson 2005) in observations. The E–W-oriented bands do not initially propagate in the zonal direction and may be analogous to the locally forced nocturnal component in climatological time-longitude diagrams.

**Fig. 5.** Simulated 500-hPa horizontal winds, geopotential height (contour interval = 30 dam), and the vertical component of relative vorticity ($10^{-5}$ s$^{-1}$) at (a) $t = 44$ h, (b) $t = 50$ h, (c) $t = 56$ h, and (d) $t = 63$ h. Dashed lines and arrows indicate the axes of short wave troughs and convection initiation sites discussed in the text.
of warm-season rainfall (Tuttle and Davis 2006, their Fig. 4c).

4. Composite mesoconvective structure and dynamics

a. Methodology

In this section we focus on the simulated mesoscale structure of the approximately N–S-oriented convective bands (Figs. 6a–e) since they contribute most substantially to the overall zonal propagation of the five simulated heavy precipitation episodes. The most basic elements of their structure are conveyed using 200-km-long composite vertical cross sections at three evolutionary stages. These consist of a growth stage that marks the earliest linear organization, which occurs from 2200 to 0300 UTC for the individual cases; a mature stage that marks their greatest horizontal scale, organization, and intensity of mesoscale circulation, which ranges from 0400 to 0800 UTC; and a weakening stage, where MCSs are in varying stages of decay from 1100 to 1300 UTC. The composites comprise vertical cross sections for individual cases constructed at single times most representative of the above stages.

The composite for the weakening stage requires more cautious interpretation than those of the other two stages, since the variations in organization and strength among the five cases is greatest during this stage. Three cases weaken substantially (cases C1, C3, and C4), one persists in a moderately weakened state for a significant time (case C2), while another (case C5) weakens only slightly. The chosen compositing strategy at these later times reflects an additional objective to examine the average structure at differing longitudes and stages of the diurnal cycle. In this way the growth phase represents late afternoon/early evening organization over the high plains, the mature phase represents overnight organization over the central Great Plains, while the weakening phase represents the organization around or slightly after sunrise over the Midwest. Although our primary emphasis is on robust features of the composites, individual cases are also examined to help understand causes of variability.

The vertical cross sections (locations shown in Figs. 6a–e) are taken through the approximate leading-edge centroid of convection where the MCSs are typically most intense and two-dimensional in organization. The simulation contains cellular convection near the leading edge of MCSs that is most likely underresolved (Bryan et al. 2003). To both avoid unrepresentative cellular structure and to emphasize mesoscale aspects of the thermodynamic and flow structure, vertical cross sections for each case are averaged for 40 km normal to their axes (i.e., centered across the section lines in Figs. 6a–e). Vertical cross-section fields are then interpolated to a height above ground level (AGL) vertical coordinate prior to compositing.

Vertical profiles of environmental quantities that influence MCS evolution including CAPE and convective inhibition (CIN; Colby 1984) are calculated using virtual temperature (e.g., Doswell and Rasmussen 1994) and also composited. To ensure that these quantities are not strongly influenced by local convectively induced circulations, vertical profiles are obtained at \( x = 190 \text{ km} \) in the vertical cross sections, 30 km ahead of the leading edge of the MCS-induced cold pool.

b. MCS vertical structure and dynamics

Strong PBL-based convection occurs during the growth stage (Fig. 7a) with inflow of high-equivalent potential temperature, \( \theta_e \), air originating from in front (east) of the system that ascends steeply above the convectively produced cold pool. The high-\( \theta_e \) air contributes to a composite CAPE of 1800 J kg\(^{-1}\) for PBL air parcels (Fig. 8a). Average CIN of \( \sim 60 \text{ J kg}^{-1} \) exists for PBL-based convection (Fig. 8a), which suggests that the convection requires localized forcing (i.e., lifting at the leading edge of the convective system) at this stage, since updrafts of order \( w = \sqrt{2 \text{CIN}} \sim 10 \text{ m s}^{-1} \) are required to displace PBL air to its level of free convection.

Significant evolution of the mesoscale flow features commonly associated with well-developed squall lines occurs by the mature stage (Fig. 7b). These include the rearward expansion and intensification of front-to-rear flow behind the leading edge convective zone and the descending rear-inflow jet of low-\( \theta_e \) air, which now penetrates to the leading edge of convection, beneath the ascending front-to-rear flow.

In addition to the pronounced evolution of internal mesoscale flow features of the MCS, substantial changes in the thermodynamic character of the MCS inflow occur by the mature stage (2300 to 0300 local time). Maximum \( \theta_v \) has increased by \( \sim 2\text{–}3 \text{ K} \) (Figs. 7a,b) resulting in maximum composite CAPE of 2100 J kg\(^{-1}\), which unlike for the growth stage, is now centered nearly 1.5 km AGL (Fig. 8b). The elevated layer of composite CAPE > 1000 J kg\(^{-1}\) is nearly 2 km deep, with negligible CIN through the upper \( \frac{3}{4} \) of this layer (Fig. 8b). While average conditions have become more thermodynamically favorable for convection based above the PBL, they are now far less favorable in the lowest 750 m where CIN has increased substantially since the growth stage. This results from both the intensity (i.e., static stability) of the frontal zone within
Fig. 6a-f live 4/C
which the mature stage convection occurs and surface-based diurnal cooling.

The maximum \( 	heta_e \) inflow during the weakening stage is also elevated (Fig. 7c), but is \( \sim 5 \) K less than during the mature stage, resulting in substantial decreases in maximum CAPE (Fig. 8c). These thermodynamic changes are consistent with the weakening of convection and its associated mesoscale vertical circulation (Fig. 7c).

The composite vertical cross sections reveal a mesoscale structure similar to previously documented squall lines, particularly during the mature stage. A widely discussed paradigm for squall-line maintenance and propagation invokes regeneration of deep convection

\[ \text{Fig. 6. (Continued)} \]

\[ \text{Fig. 6. (left) Simulated maximum radar reflectivity in a vertical column, 900-hPa horizontal winds and potential temperature (contour interval = 2.5 K) for the mature stage of long-lived nocturnal precipitation episode cases (a) C1 at \( t = 31 \) h, (b) C2 at \( t = 52 \) h, (c) C3 at \( t = 79 \) h, (d) C4 at \( t = 125 \) h, and (e) C5 at \( t = 152 \) h. Bold lines in (a)–(e) indicate the locations of 200-km-long vertical cross sections used to construct the mature stage composite vertical cross section in Fig. 7b. (right) NOWrad composite radar images at times corresponding to cases (f) C1, (g) C2, (h) C3, (i) C4, and (j) C5.} \]
at the leading edge of a diabatically cooled density current as it advances in a vertically sheared environment (Rotunno et al. 1988; Fovell and Ogura 1989; Weisman and Rotunno 2004; Bryan et al. 2006). Rotunno et al. (1988) argue that the important physical process is the interaction of the environmental shear and the horizontal vorticity generated at the leading edge of the cold pool. The latter can be quantified by $c$, which is also the theoretical propagation speed of a cold pool (density current) in a fluid of infinite depth, which is given by

$$
c^2 = -2 \int_{z=0}^{z=H} B \, dz,
$$

where $H$ is the depth of the cold pool, and $B = g(\theta_v/\theta_e - q_c)$ is the total cold pool buoyancy including contributions from virtual potential temperature perturbations, $\theta_v$, and the mixing ratio of all condensate, $q_c$. Ignoring horizontal vorticity sources within the cold pool, an "optimal" condition for deep lifting at the leading edge of the cold pool occurs when $c / \Delta u \sim 1$, where $\Delta u$ is the environmental vertical shear through the depth of the cold pool oriented normal to its leading edge. For $c / \Delta u < 1$, updrafts are predicted to tilt downshear, which inhibits upscale growth and organization, while $c / \Delta u > 1$ predicts the upshear tilt commonly observed in the mature-to-dissipating stage of squall lines.

Limitations in horizontal resolution preclude a rigorous dynamical analysis of the convection in the current simulation. Nevertheless, it is instructive to examine its overall behavior in light of recent theories relating squall-line evolution and motion (next subsection) to cold pools, vertical shear, CAPE, and other factors. Here, environmental density stratification through a significant portion of the cold pool depth in nocturnal convection (Carbone et al. 1990) could result in differences from previous idealized simulations, in which the PBL is often approximately neutral and thus more representative of daytime conditions.

We illustrate the capacity for horizontal vorticity generation near the cold pool leading edge with 40-km
along-line averaged vertical profiles of the virtual potential temperature perturbation \( \theta'_v = \bar{\theta}'_v - \bar{\theta}'_v \), where \( \bar{\theta}'_v \) and \( \bar{\theta}'_v \) are, respectively, average values over the 20-km regions centered at \( x = a \) and \( x = b \) in Fig. 9. Environmental wind profiles are obtained at \( x = 260 \) km in Fig. 9, a full 100 km ahead of the surface horizontal \( \theta_v \) gradient, in order to mitigate the possible influence of anvil circulations on upper tropospheric winds, since deep shear has also been found to be influential in squall-line evolution (e.g., Coniglio and Stensrud 2001).

The simulated mature stage cold pools have substantial maximum \( \theta_v \) perturbations of \( -5 \) to \( -9 \) K and extend upward to near the melting level around \( -4 \) km AGL (Fig. 10a; Table 1). Despite fairly uniform depth \( H \) (Table 1), differences in vertical structure occur among the cases. The strongest frontal cases (cases C2 and C5; cf. Figs. 6a–e) are coldest relative to the environment at 1.5–2 km AGL (Fig. 10a). Large \( \theta'_v > 0 \) occurs above the cold pool and is associated with strong latent heat release in the upshear-tilted updraft branch (Fig. 7b). The simulated deep regions of large cooling in the lower troposphere and warming in the middle troposphere are consistent with BAMEX dropsonde analyses in deep convection (Bryan et al. 2005).

Vertical shear oriented normal to the cold pool through the cold pool depth ranges from 13 to 28 m s\(^{-1}\) with a mean value of 20 m s\(^{-1}\) (Fig. 11a and Table 1). Weaker vertical shear is present through the remainder of the middle troposphere (Fig. 11a). The “suboptimal” \( c/\Delta u = 2.7 \) for case 1 is similar to that found by Coniglio and Stensrud (2001) for a long-lived simulated derecho. This suboptimal condition results from both anomalously large CAPE of 2850 J kg\(^{-1}\) (Table 1),
which promotes a strong cold pool, and the lack of a significant east–west-oriented surface front (Fig. 6a), which is consistent with the absence of strong low-level shear. Other cases more closely satisfy the horizontal vorticity balance criterion, with $c/\Delta u = 1.5$ for the strong vertical shear cases (C2, C4, and C5) associated with the strongest surface fronts (Figs. 6b,d,e).

As noted earlier, the Rotunno et al. (1988) vorticity balance criterion excludes consideration of vorticity within the cold pool itself. Weisman (1992) argued that convectively generated rear-inflow jets that remain elevated to near the top of the cold pool modify the horizontal vorticity balance in such a way that aids lifting beneath the jet height at the leading edge of squall lines beyond what the $c/\Delta u$ criterion might otherwise imply. The strongest elevated rear-inflow jets occur in cases C2 and C5 (Fig. 12), where environmental vertical shear is particularly strong.

On average, both cold pool strength (Fig. 10b) and environmental vertical shear (Fig. 11b) decrease by a factor of ~2 during the near sunrise weakening stage. A notable exception occurs in case C5, where the cold pool and environmental vertical shear strengths decrease only slightly and the MCS experiences only minimal weakening.

Despite their occurrence above a stable PBL, the overall strength and organization of the simulated nocturnal MCSs appear to be influenced by the interaction of the MCSs-induced cold pool and the environmental vertical shear. Some aspects of MCS evolution also appear related to this interaction. In particular, strong, long-lived cases are associated with elevated rear-inflow jets and continuing strong environmental vertical shear. However, other aspects of the evolution are influenced by significant evolution in the environment itself, including decreases in CAPE prior to weakening (Fig. 8c), as also found by Coniglio and Stensrud (2001). Additional important changes to the environment of the convection are examined in subsequent sections.

c. MCS propagation

As emphasized in Carbone et al. (2002), the zonal speed of the simulated rain streaks, $c_{mod}$, cannot be
explained by simple advection. Figure 13a compares the zonal speed of the precipitation episodes with the density-weighted mean environmental zonal flow $\bar{U}$ through the depth of the cloud-bearing layer (1–12 km AGL) for the five cases during their mature stage. Although there is a clear correspondence between the phase speed of precipitation episodes and the mean flow, with the most rapidly moving rain streaks occurring in the strongest environmental flow, $\bar{U}$, systematically underestimates $c_{\text{mod}}$. 

We also evaluate the applicability of the analytic theory for density current speed (e.g., Benjamin 1968) as an explanation for the movement of the rain streaks. For this calculation the right-hand side of (1) is evaluated at the edges of a control volume wherein the pressure is approximately hydrostatic. In the current cases this corresponds to a location that is typically 40–60 km behind the leading edge of the cold pool (i.e., within the darker shading centered at $x = b'$ in Fig. 9). Following Rotunno et al. (1988, p. 480), we calculate ground-relative cold pool speed, $c_{gr} = c + u_0$, by adding an environmental ground-relative retarding wind $u_0$.

For cases such as the current ones, where stagnation occurs at the leading edge of the density current ($c/\Delta u \geq 1$), setting $u_0$ equal to the surface wind is justifiable on theoretical grounds (e.g., Rotunno et al. 1988). However, surface frictional effects were not represented in Rotunno et al. (1988). In our simulations, friction results in more complicated near-surface wind profiles and it is thus unclear what is the most appropriate choice for $u_0$. For this reason, and because of the general lack of clear theoretical guidance, several different estimates of $u_0$ are used.

One estimate uses the $\sim200$ m AGL environmental zonal flow, $u_0 = u_{200m}$, which resides above the friction layer at night and roughly corresponds to the greatest possible retarding flow (i.e., strongest easterlies). Another estimate, $u_0 = 0.62\pi$, is based on the mean environmental flow through the depth of the cold pool. This is an empirically derived estimate used in previous studies of atmospheric density currents (e.g., Simpson and Britter 1980; Seitter 1986). In those studies density current speed was determined using $c_{gr} = k(\Delta p/\rho_0)^{0.5} + 0.62\pi$, where $\Delta p_0$ is the surface pressure perturbation, $\rho_0$ is a base-state surface density, and $k$ is an empirical constant. Under hydrostatic conditions, when the theoretical value of $k = \sqrt{2}$ is used, the first term on the right-hand side of the above relation yields values of $c$ within a few percent of those obtained from integration of (1) through depth $z = D$ (where the model pressure perturbation $\Delta p \approx 0$) in Fig. 9. However, these previous studies typically reduce $k$ from its theoretical value to account for frictional effects, which we do not. Therefore, our estimates that use $u_0 = u_{200m}$ and $u_0 = 0.62\pi$ as retarding motions, should be regarded, respectively, as lower and upper bounds on possible zonal speeds of rain streaks resulting from density current propagation. A third intermediate estimate uses the retarding motion, $u_0 = u_{10m}$, which is based on the model-derived environmental zonal wind near the surface at 10 m AGL.

We first consider the estimates of $c_{gr}$ that result from the buoyancy anomaly through the cold pool depth $H$ alone (Fig. 13b). This is obtained by adding the estimates of the retarding flow to the value of $c$ obtained from the vertical integration of (1) from $z = a'$ to $z = H$, where $z = a'$ denotes the lowest height of a level surface common to the averaging regions centered on $x = a$ and $x = b'$ in Fig. 9. In contrast to the estimates of rain streak movement obtained using the density-weighted mean zonal wind (Fig. 13a), these estimates based on the cold pool negative buoyancy systematically overpredict the speed of rain streaks (Fig. 13b).
surprising given the far greater complexity of an eastward-moving squall system compared to an isolated density current (e.g., Moncrieff and So 1989). The composite simulated vertical structure of the MCS during its mature stage (Fig. 7b) resembles the triple branch archetype of Moncrieff (1992), which includes overturning and upshear-tilted updraft branches in addition to the deep, cold descending branch behind the leading edge.

In addition to these kinematic differences, the vertical profiles of buoyancy (Fig. 14) exhibit a compensating region of positive buoyancy in the middle troposphere and an additional region of negative buoyancy.

**Fig. 14.** Buoyancy profiles obtained by subtracting $20 \times 40 \, \text{km}^2$ environmental averages (centered at $x = a$ in Fig. 9) from identically sized averages from behind the cold pool edge (centered at $x = b'$ in Fig. 9). These profiles were used to estimate the zonal propagation of precipitation episodes in Figs. 13b,c.

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**Fig. 13.** Zonal speed of model rain streaks (Fig. 2) for the five composited cases of simulated long-lived nocturnal precipitation episodes (solid circles) and various estimates of the zonal speed based on (a) the density-weighted mean environmental zonal winds in Fig. 11 (open square symbols) and (b) and (c) ground-relative density current motions, $c_{gr} = c + u_0$, obtained by adding estimates of environmental retarding flow, $u_0$ (see text), to $c$ values obtained from vertical integrations of Eq. (1) up through the depths (b) $z = H$ and (c) $z = D$ in Fig. 9. In (b) and (c) the ranges represent bounds of uncertainty on $c_{gr}$ based on three different estimates of $u_0$. The median estimate of $c_{gr}$ is indicated for each case by the open circles.
near the tropopause. The hydrostatic surface pressure perturbations that contribute to cold pool motion are the net effect of all three regions of significant buoyancy perturbations in the vertical profile. Thus, rain streak zonal speeds are better explained when we consider the deep buoyancy anomaly (Fig. 13c) that exists from \( z = z \) to \( z = D \) in Fig. 9. The range of estimates appear somewhat slow in two of the five cases (C4 and C5). However, when our median estimate of retarding flow is employed the estimates are within 10% of \( c_{\text{mod}} \) for three of the five cases and within 25% for all but one case (C5).

Case C5 is also unique among the cases since it is the final long-lived precipitation episode during this week-long active period of nocturnal convection and is associated with a large-amplitude trough and eastward-moving cold front (not shown). Note that the zonal motion estimate based on the mean tropospheric flow (obtained 100 km east of the precipitation system) also constitutes a particularly large underestimate of \( c_{\text{mod}} \) relative to other cases (Fig. 13a). This case featured substantial large-scale gradients across the precipitation system, with much stronger tropospheric westerlies upstream, which could have resulted in a low bias of our advection estimate. The strong winds aloft upstream also suggest that large-scale effects could be contributing more to the movement of this precipitation system than in other cases.

Overall, the agreement between modeled rain streak phase speed obtained using deep (~15 km) buoyancy perturbations and the near-surface retarding flow is quite reasonable considering several potentially severe assumptions inherent in this comparison. Among these are steadiness in both the derivation of the theoretical propagation speed in (1) and that implicit in the model precipitation data fits (Fig. 2), and the neglect of surface friction in the derivation of (1).

5. Composite lower-tropospheric analysis

To gain understanding of the mechanisms accounting for the environmental changes that influence the convective life cycle discussed in the previous section, we now consider the evolution of the lower troposphere over a \( 800 \times 720 \text{ km}^2 \) region surrounding the frontal zone in which the zonally propagating convection occurs. Horizontal cross sections (Fig. 15) of various meteorological fields on constant pressure surfaces or through layers of a specified pressure depth are created for each case. Individual fields from the five different cases are then composited relative to the N–S centroid locations of the leading-edge convection. In addition, composite vertical cross sections are constructed across the frontal zone in advance of propagating convection. These vertical cross sections (Figs. 16 and 18) are created similarly to those described in section 4a, including 40-km averaging in the section-normal direction.

a. Horizontal cross sections

During the growth stage, maximum CAPE is greatest in the vicinity of the composite lower-tropospheric frontal zone delineated by the 925–675-hPa thickness contours (Fig. 15a). The maximum CAPE increases to over 2000 J kg\(^{-1}\) and becomes more localized within the frontal zone at night during the mature stage (Fig. 15b). Maximum CAPE weakens, and in contrast to the growth and mature stages, now extends from the southern edge to several hundred kilometers south of the baroclinic zone during the weakening stage (Fig. 15c).

Maximum CAPE (Figs. 15a–c) is spatially correlated with the maximum 850-hPa water vapor mixing ratio (Figs. 15d–f), particularly during the mature and weakening stages when convection is elevated (section 4). The frontal zone is subject to strong horizontal deformation during the growth (Fig. 15d) and mature (Fig. 15e) stages. Warm advection occurs south of the frontal zone in the growth stage (Fig. 15d) and becomes stronger during the mature stage as the nocturnal LLJ intensifies (Fig. 15e). Although warm advection continues along the southern part of the front during the weakening stage, strong deformation is absent (Fig. 15f). The absence of deformation during this stage is due to both the diurnal veering of the LLJ south of the front (Figs. 15e,f) and changes in flow north of the front from easterly (Fig. 15d) to westerly (Fig. 15f) as one moves east.

Lower-tropospheric (925–675 hPa) vertical shear varies across the frontal zone (Figs. 15a–c), with differences most pronounced during the mature stage (Fig. 15b). The northerly vertical shear south of the frontal zone at this time is a manifestation of relatively weak northwesterly flow aloft overlying the southwesterly LLJ. Within the frontal zone, the vertical shear is approximately westerly and oriented primarily parallel to the thickness contours during the growth (Fig. 15a) and mature (Fig. 15b) stages, consistent with thermal wind considerations. A significant component of the vertical shear is, however, oriented across the thickness contours. The cross-frontal component of the vertical shear suggests a thermally direct transverse frontal circulation (e.g., Keyser and Shapiro 1986) forced by horizontal deformation acting on the thermal gradient. During the weakening stage, the vertical shear in the vicinity of the front is weaker and oriented more closely parallel to the thickness contours, consistent with a weaker
FIG. 15. Composite horizontal cross sections of maximum CAPE based on the air parcel with largest equivalent potential temperature in the vertical column (500 J kg\(^{-1}\) shading interval), 925–675-hPa thickness (contour interval = 10 dam), and 925–675-hPa vertical shear (vectors) for the (a) growth, (b) mature, and (c) weakening stages of the long-lived nocturnal precipitation episodes denoted in Fig. 2a and discussed in the text. (d)–(f) Same as in (a)–(c), respectively, but with fields of 850-hPa water vapor mixing ratio (shading interval = 1.5 g kg\(^{-1}\)), potential temperature (contour interval = 2 K), and ground relative winds (vectors). Bold arrows in (d)–(f) are schematic streamlines emphasizing horizontal deformation characteristics. Lines AB in (e) and CD in (f) denote locations of vertical cross sections shown in Figs. 16a,b, respectively. In each panel the + symbol denotes the location of the leading edge convection relative to which the five cases of long-lived nocturnal precipitation episodes were composited.
thickness gradient (Fig. 15c) and lack of strong deformation (Fig. 15f).

b. Vertical cross sections

A time-averaged composite vertical cross section (location shown in Fig. 15e) from the 6-h preceding the mature stage (Fig. 16a) shows the LLJ in the plane of the cross section ascending the northeastward sloping frontal surface. Maximum mesoscale ascent in this composite is \( \sim 5 \text{ cm s}^{-1} \) and occurs at \( x = 300 \) km. This location is slightly south of a region of strongest deformation frontogenesis, \( -\left( \frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) > 0 \), which is coincident with the strongest surface potential temperature gradient. A clearly defined thermally direct transverse circulation is evident about the frontal zone. The inflow begins to ascend as it enters the deformation zone, with weak subsidence occurring a few hundred kilometers farther northeast and an instantaneous maximum in mature stage \( q_v \) extends upward from the time-averaged surface deformation zone.

The structure in this vertical cross section contrasts markedly with that in an identically constructed vertical cross section (location shown in Fig. 15f) for the 6 h preceding the weakening stage. In the weakening stage composite (Fig. 16b), the frontal slope is shallower and the southwest horizontal wind component in the plane of the cross section is much weaker. Together, these conditions imply weaker isentropic ascent within the frontal zone, which is consistent with the weaker spatially and temporally averaged maximum upward motion of 2–3 cm s\(^{-1}\). Although maximum \( q_v \) during the weakening stage is comparable to that during the mature stage, the local maximum does not extend as far north of the front, nor does it exhibit as prominent an upward bulge emanating from the surface (cf. Figs. 16a,b).

c. Moisture source regions

Since it contributes importantly to the conditional instability, it is of interest to investigate the origin of the deep moisture within the frontal zone during the mature stage. Numerous studies have discussed the importance of the horizontal moisture transport by the LLJ on the warm-season hydrological cycle of the central United States (e.g., Hellland and Schubert 1995; Higgins et al. 1997). Although horizontal transport may be important in establishing the large-scale lower tropospheric condition of the regime in which this 7-day simulation occurs, it is interesting to note that the horizontal advection of moisture toward the frontal zone by the southerly flow is approximately neutral to strongly negative from the growth (Fig. 15d) to the mature (Fig. 15e) stages of propagating convection, respectively.

Deep moisture within the frontal zone is, however, consistent with the time-averaged LLJ ascending isotropic surfaces, which are approximately parallel to \( q_v \) surfaces both within the leading edge of the frontal zone and to its south (Fig. 16a). Another possibility is that the moisture originates locally within the frontal zone PBL. A concentrated zone of maximum near-surface \( q_v \) occurs along and immediately north of the 0000 UTC surface wind shift (Fig. 17) for a horizontal composite (in this case relative to surface frontal posi-
tion) prior to mature stage convection. So-called pooling of moisture near east–west-oriented surface fronts has been reported in previous studies of derecho environments (Johns 1993; Coniglio and Stensrud 2001) and climatological studies of episodes of traveling nocturnal convection confined to narrow latitudinal corridors (Tuttle and Davis 2006).

Regardless of the source of moisture (local or LLJ), upward motion within the secondary circulation forced by deformation frontogenesis appears crucial in establishing the large moisture values over a deep layer within the frontal zone at night. Figure 18 shows 6-h $q_v$ increases of $\sim 1$ g kg$^{-1}$ above 1 km AGL both within and immediately downstream of the vertical gradient of moisture in the frontal zone (cf. Fig. 16a) prior to the mature stage.

6. Trajectory analysis

Trajectory analyses for the individual cases quantify the local contribution to deep moisture originating from within the frontal zone and the LLJ-influenced contribution originating to its south. These analyses also explain differences in the horizontal moisture distributions from the mature to weakening stages (Figs. 15e,f). Both forward and backward trajectories are calculated using 5-min time steps with model output data interpolated from their original 30-min frequency.

a. Forward trajectories for a strong frontal case (case 2)

The second nocturnal precipitation episode (Fig. 6b) possessed the strongest surface front and LLJ in the series of five cases (Figs. 6a–e). This particular episode also featured the most pronounced afternoon surface $q_v$ maximum within the frontal zone (not shown).

Forward trajectories released within the PBL at 925 hPa ($\sim 500$ m AGL) during early afternoon both from within and considerably south of the frontal zone compose the elevated high-$\theta_e$ zone ($\sim 344$ K) in advance of deep convection 8.5 h later during the midevening (Fig. 19a). The frontal zone air, which approaches from the east, undergoes average vertical displacements of $\sim 600$ m (Fig. 19b). The air originating in the daytime PBL south of the front experiences much greater total average vertical displacements of $\sim 1500$ m, with some individual trajectories rising 2200 m (Fig. 19c). This trajectory branch also traverses a greater horizontal distance (Fig. 19a) since it constitutes part of the LLJ during the evening. Average CAPE (CIN) along these LLJ trajectories increases (decreases) 750 J kg$^{-1}$ (275 J kg$^{-1}$) as a result of the air parcels ascending while moving northward away from a region with a significant daytime PBL capping inversion into the region above the frontal zone, which is colder aloft (cf. Fig. 1).
b. Back trajectory analysis for the mature stage

The forward trajectory analysis for the strong frontal case conforms well to the conceptual picture in the time-averaged composite (Fig. 16a), in which ascending LLJ air contributes to the majority of the depth of the high-$\theta_e$ layer supporting propagating nocturnal convection while air of frontal origin undergoes more modest vertical displacements and constitutes the lower part of this layer. These multiple source regions of conditionally unstable air are now confirmed for other cases using 6-h back trajectories released at 10-hPa intervals, where $\theta_e > 342$ K, in advance of the approximate leading-edge centroid of mature-stage propagating convection. As in the analysis of mesoconvective structure in section 4, we choose a horizontal location in advance of the central portion of the convective line as being most representative of the environment of the convective system as a whole, but recognize that meridional variation occurs (section 5).

Average trajectories for the frontal and LLJ airstreams in each of the five cases are presented in Fig. 20a. The corresponding layers that comprise these averaged trajectories are indicated on accompanying soundings at the trajectory release points for each case (Fig. 21). Air of both frontal and LLJ origin has comparable CAPE at the trajectory release points (i.e., in advance of the propagating convection). The frontal air, however, has a more stable lapse rate and thus significant CIN. Excluding the anomalously dry case C3 (Fig. 21c), CIN ranges from 50 to 125 J kg$^{-1}$ and may be overcome by strong localized lifting at the leading edge of the mature N–S-oriented zonally propagating convection. By contrast, LLJ air has negligible CIN (Fig. 21) when it reaches the elevated large CAPE layer (cf. Fig. 8b) and supports both the zonally propagating convection that occurs in all cases, and locally developing convection oriented along the front, which is widespread in case C2 (Fig. 6b) and more isolated in case C4 (Fig. 6d).

For each case, geopotential height, relative humidity (RH), $q_v$ and $\theta$ are averaged for the individual sets of trajectories corresponding to the two different layers of origin (i.e., frontal and LLJ). The average 6-h vertical displacement of LLJ air (Fig. 22a) is 900 m, with significant vertical displacements occurring in all four cases. While the LLJ air eventually becomes moist relative to surroundings (Fig. 16a), this does not occur until after it undergoes the isentropic ascent. Here, the air is potentially warm with average $\theta = 311$ K (Fig. 22c) and initially highly subsaturated with average RH of $\sim 55\%$ (Fig. 22b) that increases to $\sim 90\%$ along trajectories.

Air of frontal origin, which is potentially cooler (Fig. 22g) and typically moister (Fig. 22h) than LLJ air (Fig. 22d), undergoes a more modest average 6-h vertical displacement of 300 m (Fig. 22e) compared to the 900 m for LLJ air (Fig. 22a). The character of the vertical displacements also varies more among families of frontal trajectories than for LLJ air, with only the strong frontal cases (C2, C4, and C5) experiencing significant upward displacements.
c. Back trajectory analysis for the weakening stage

Unlike for the mature stage of eastward propagating convection, during the weakening stage the thermodynamically unstable air ($\theta_e > 342$ K) that sustains convection comes entirely from southwest of the front (Fig. 20b). These similarly averaged trajectories have a significantly greater westerly component than do the mature stage LLJ trajectories (Fig. 20a). This is consistent with the inertial oscillation, which produces an anticyclonic rotation of the PBL winds (e.g., Blackadar 1957).

Smaller average 6-h upward displacements (Fig. 23a) than during the mature stage (Fig. 22a) are consistent with both a shallower frontal slope (cf. Figs. 16a, b), and the flow direction becoming more parallel to the front. Average upward displacements for LLJ air during the
respective weakening (400 m) and mature (900 m) stages are consistent with that expected from isentropic upglide of the frontal surface in the 6-h time-averaged composite cross sections (Fig. 16), where maximum mesoscale vertical velocities are \(\frac{1}{10}\) and \(\frac{5}{2}\) ms \(^{-1}\).

Although weaker, the mesoscale lifting during the weakening stage is still sufficient to approximately saturate lower-tropospheric layers in the strongest frontal cases (C2, C4, and C5) due to the larger initial relative humidities (Fig. 23b) than those along LLJ trajectories during mature stage (Figs. 22b). The larger initial relative humidity during the weakening stage results from smaller \(\theta\) (Figs. 22c, 23c) and slightly larger \(q_v\) (Figs. 22d, 23d).

Recall that the mature stage LLJ trajectories originate within the strongly heated PBL of the central Great Plains. This contrasts with the weakening stage trajectories, which both originate farther east, where conditions are generally cooler in this particular large-scale pattern, and later in the diurnal cycle after the PBL has cooled substantially. In addition, an average gradual cooling of 1.5 K occurs along the 6-h nocturnal trajectory paths during the weakening stage (Fig. 23c). The smaller \(\theta\) explains the significantly smaller maxi-

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**Fig. 22.** Average values of (a) geopotential height (m MSL), (b) relative humidity (%), (c) potential temperature (K), and (d) water vapor mixing ratio (g kg \(^{-1}\)) for the low-level jet family of 6-h back trajectories for the mature stage of long-lived nocturnal precipitation episodes C2 (medium dashed), C3 (single dot dashed), C4 (thin solid), and C5 (long dashed) shown in Fig. 20a. The bold solid lines in (a)–(d) are the average of these four cases. (e)–(h) Same as in (a)–(d), but for the frontal family of 6-h back trajectories for the long-lived nocturnal precipitation episodes C1 (short dashed), C2, C3, C4, and C5 shown in Fig. 20a.
maximum CAPE for weakening stage convection than for mature stage convection discussed in sections 4 and 5 (Figs. 8b,c and Figs. 15b,c). The shift of maximum $q_v$ from north to south of the surface front from the mature (Fig. 15e) to weakening stage (Fig. 15f) results from the absence of the daytime frontal zone source of large $q_v$ for the weakening stage and the evolution of the LLJ trajectory paths to a direction more parallel to the front during the weakening stage. This shift influences the horizontal distribution of maximum CAPE (Figs. 15b,c) from north to south of the front with time.

The environment, on average, is far less favorable for intense convection during this stage than during the mature stage. The minority cases, where significant convection persists beyond sunrise (case C2 and most notably case C5), are ones where a combination of significant lifting (>400 m) along trajectories (Fig. 23a) and moderate-to-strong environmental vertical shear (Fig. 11b) persists.

7. Summary and discussion

An explicit (convection-permitting) version of the WRF is used to simulate eastward traveling convection for a one-week period during which regular long-lived ($t > 9$ h) episodes of deep convection attained maximum intensity and organization at night over the central United States. Rather than attempting deterministic prediction of individual precipitation episodes as is traditionally done, the current study examines statistics of an ensemble of MCSs during a particularly active period of organized convection. Multiscale composite analysis of the environment and mesoconvective structure of the evolutionary stages of multiple episodes of convection is then combined with trajectory analysis of individual cases to diagnose mechanisms responsible for the convective life cycle.

Both the observed and simulated precipitation episodes are confined to a narrow corridor (<5° latitude) for the entire week, as has been more generally observed in climatological studies (e.g., Carbone et al. 2002; Tuttle and Davis 2006) of warm-season convection over the central United States. This localization of convection is aided by the synoptic condition of moderate-to-strong midtropospheric flow directed approximately parallel to an east–west-oriented quasi-stationary surface front. Zonal phase speeds of the simulated convection compare favorably with radar observations from the one-week period.

The simulation indicates that initial afternoon/early evening convection is based in the PBL, but is later aided by a 2-km-deep conditionally unstable layer beginning ~750 m above the surface during its most organized phase overnight and during its weakening stage around sunrise. Significant conditional instability is restricted to a region of several hundred kilometers immediately north of the surface front and results from locally deep moisture. The trajectories that comprise the lowest $\frac{2}{3}$ or more of the deep conditionally unstable layer is composed of nocturnal low-level jet (LLJ) air that originates in the PBL south of the front.

Although the importance of the LLJ to organized nocturnal convection has been known for decades, our results suggest a less widely appreciated mechanism by which it supports nocturnal convection. The LLJ is likely the ultimate source of moisture for the active
convection regime. However, on the shorter time scale of the nocturnal convection its more critical role appears to be in enhancing frontogenesis and its associated secondary circulation, within which large vertical displacements of moisture occur. This process ensures that within the frontal zone both deep moisture and strong vertical shear (implied by thermal wind considerations) coexist. The deep moisture provides enhanced conditional instability with negligible CIN to fuel the propagating convection, which remains organized by virtue of the low-to-midtropospheric vertical shear.

Given the favorable environmental conditions, the mesoscale organization and longevity of the precipitation episodes result from continuous regeneration of deep convection along the leading-edge cold pool, as reported in previous studies of squall lines (e.g., Rotunno et al. 1988). The overall zonal phase speed of heavy precipitation during the mature stage is consistent with that arising from the vertically integrated MCS-induced buoyancy (equivalent to that associated with the hydrostatic part of the MCS-induced surface pressure perturbation), modulated by the opposing environmental surface flow. Significant contributions to the vertically integrated buoyancy arise not only from the MCS cold pool but also from convectively induced buoyancy that extends upward to above the tropopause.

Although aspects of the mature-stage convective organization and intensity appear to be well explained by conventional theories of squall-line dynamics, the evolution of convection is strongly influenced by both geographic and temporal changes to its environment, which have not been accounted for in most previous idealized simulations. Our study provides a hypothesis for why propagating convection is climatologically observed (Carbone et al. 2002, their Fig. 12) to terminate, or at least weaken temporarily, around and slightly after sunrise. In the simulation it appears that the weakening of convection has little to do with the sunrise itself, but is more closely tied to the inertial period and the cumulative effects of nocturnal cooling. In particular, LLJ winds develop a more westerly component due to the inertial oscillation and are thereby oriented more parallel to the surface front during the weakening stage of convection. This contributes to weaker deformation and frontogenesis, and is consistent with both the weaker environmental upward vertical motions and weaker vertical shear during this stage. Having originated later and farther east, the weakening stage trajectories are potentially colder than LLJ trajectories from the mature stage, which additionally results in lower $\theta_v$ and less CAPE.

This study has addressed the nocturnal portion of the convective life cycle that occurs in a large-scale pattern commonly associated with multiple episodes of phase-coherent zonally propagating convection confined to narrow latitudinal corridors in the central United States. Climatological observations of phase-coherent traveling convection routinely persisting through the nocturnal hours have recently emerged from other continental regions including East Asia (Wang et al. 2004) and sub-Saharan Africa (Laing et al. 2006). A common aspect of these convective episodes and nocturnal convection on additional continents including South America (e.g., Velasco and Fritsch 1987) is upscale organization of convection downstream from significant topography, as found in North America (Carbone et al. 2002; Ahijevych et al. 2004). Composite studies (e.g., Laing and Fritsch 2000) further indicate similar features of the large-scale environment including significant baroclinity with strong low-level wind shear and low-level jets of high-$\theta_v$ air impinging on these baroclinic zones, most notably in midlatitude-to-subtropical South America and East Asia. This suggests that the mechanisms that support long-lived nocturnal convection described in this paper could be operative elsewhere.

Our focus has been on only a limited part of the life cycle of long-lived convective episodes. However, Carbone et al. (2002) found that convection often regenerates over the Midwest or eastern United States during the afternoon along the approximate phase line as the nocturnal convection that originated farther west during the previous heating cycle. This is evident from Fig. 2 in both the observations and simulation (cases C3 and C4 in particular) of this one-week period and was discussed briefly in section 3. Similar intermittency and regeneration characteristics within multiday precipitation episodes have also been documented over other continents (e.g., Wang et al. 2004; Laing et al. 2006). The linkage between these separate diurnal cycles of phase-coherent convection is unclear and constitutes an important topic for future research.

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