The Structure, Evolution, and Dynamics of a Nocturnal Convective System Simulated Using the WRF-ARW Model

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ABSTRACT

Previous studies have documented a nocturnal maximum in thunderstorm frequency during the summer across the central United States. Forecast skill for these systems remains relatively low and the explanation for this nocturnal maximum is still an area of active debate. This study utilized the WRF-ARW Model to simulate a nocturnal mesoscale convective system that occurred over the southern Great Plains on 3–4 June 2013. A low-level jet transported a narrow corridor of air above the nocturnal boundary layer with convective instability that exceeded what was observed in the daytime boundary layer. The storm was elevated and associated with bores that assisted in the maintenance of the system. Three-dimensional variations in the system’s structure were found along the cold pool, which were examined using convective system dynamics and wave theory. Shallow lifting occurred on the southern flank of the storm. Conversely, the southeastern flank had deep lifting, with favorable integrated vertical shear over the layer of maximum CAPE. The bore assisted in transporting high-CAPE air toward its LFC, and the additional lifting by the density current allowed for deep convection to occur. The bore was not coupled to the convective system and it slowly pulled away, while the convection remained in phase with the density current. These results provide a possible explanation for how convection is maintained at night in the presence of a low-level jet and a stable boundary layer, and emphasize the importance of the three-dimensionality of these systems.

1. Introduction

This work is motivated by the need to improve forecasts of summertime, nocturnal convective systems that occur over the Great Plains of North America. The problem is challenging, as studies (e.g., Heideman and Fritsch 1988; Uccellini et al. 1999; Weckwerth et al. 2004) have argued that the accuracy of quantitative precipitation forecasts is consistently lower during the warm season as a result of the increased role of deep convection. Researchers have demonstrated that there is a pronounced nocturnal maximum in rainfall across the central United States (e.g., Kincer 1916; Wallace 1975) and that nocturnal convection has been difficult to represent in numerical weather prediction (NWP) modeling systems, particularly those with parameterized convection (e.g., Surcel et al. 2010). Climate models have also had difficulty in accurately representing these nocturnal convective systems (Bukovsky and Karoly 2009; Pritchard et al. 2011). Although convection-allowing models have demonstrated some skill (Davis et al. 2003; Clark et al. 2007), significant challenges still exist even with convection-permitting ensembles (Clark et al. 2009). Our study focuses on a high-resolution, convection-allowing simulation of a nocturnal mesoscale convective system (MCS) over the southern Great Plains.

Fritsch et al. (1986) asserted that MCSs account for 30%–70% of the warm season precipitation over the region between the Rocky Mountains and the Mississippi River, with larger contributions during the months of June–August. The rainfall that is generated from these MCSs is essential to mitigating drought and enhancing midsummer crop growth, which increases overall agricultural production. However, MCSs also herald the potential for severe weather including hail, damaging winds, tornadoes, and flash floods (Maddox 1980; Jirak et al. 2003).

The nocturnal maximum over the Great Plains can be attributed in part to convective systems that originate over the higher plains and the mountains to the west (Cotton et al. 1983; Wetzel et al. 1983), which then become associated with an eastward-moving envelope of deep convection (e.g., Carbone et al. 2002; Ahijevych et al. 2004; Parker and Ahijevych 2007; Carbone and
The nocturnal environment over the Great Plains associated with these storms is often characterized by the presence of a stable boundary layer (SBL) and a low-level jet (LLJ) (e.g., Blackadar 1957; Holton 1967; Shapiro et al. 2016). The LLJ is often associated with an elevated maximum in equivalent potential temperature $\theta_e$ and the advection of warm, moist air into the region (Maddox 1983; Wetzel et al. 1983; Fernando and Weil 2010). In addition, the LLJ is frequently present during nocturnal convective events (Bonner 1968; Maddox 1983) and is associated with heavy rainfall (Arritt et al. 1997; Tuttle and Davis 2006).

The combination of an SBL and the most unstable air aloft in association with the LLJ results in a greater occurrence of elevated convection at night, which presents an additional challenge for accurately predicting nocturnal storms (Wilson and Roberts 2006). Here, we define convection to be elevated if it occurs in an environment with the greatest instability located above Earth’s surface (e.g., Corfidi et al. 2008). A few previous authors have used different definitions for elevated convection, such as Parker (2008), whose definition was that a convective system should ingest no parcels from the surface layer. Elevated storms, often taking the form of large, nocturnal MCSs, can be initiated from air located above the boundary layer or they can originate from surface-based systems that become elevated over time (Corfidi et al. 2008; Parker 2008).

Several previous observational studies have described a preferential location for the formation of nocturnal MCSs. Maddox et al. (1979) observed a distinct nocturnal preference of storm initiation on the cool side of a quasi-stationary surface front that was located downstream from the axis of maximum low-level flow. Colman (1990) developed a climatology of elevated thunderstorms across the United States and found the typical elevated thunderstorm to occur north of a surface warm front in a region with northeasterly surface winds. Trier and Parsons (1993) also found the favored position for the development of large, organized, nocturnal convective events over the central United States to be north of a quasi-stationary frontal boundary due to the LLJ overrunning the front. However, nocturnal MCSs can occur well to the south of a frontal zone (e.g., Koch et al. 2008a).

In contrast to theory that guides the upscale growth and structure of surface-based convection (e.g., Rotunno et al. 1988), a comprehensive theoretical framework of nocturnal convection is lacking (Trier et al. 2006). Some recent studies have demonstrated that many nocturnal MCSs produce surface cold pools and respond to them as daytime MCSs would (Trier et al. 2006; Parker 2008; Peters and Schumacher 2015a, b, 2016). However, another possibility is that weak and transient cold pools can generate gravity waves in the stable air that cause persistent lifting of conditionally unstable air (Crook and Moncrieff 1988; Parker 2008; Schumacher 2009; Marsham et al. 2010). A bore is a gravity wave response that can be generated as a density current intrudes a low-level stable layer (e.g., Crook 1988; Rottman and Simpson 1989). Surface observations of density currents and bores both include hydrostatic pressure jumps during the passage of their frontal boundaries; however, the surface temperature rises or does not change as a bore passes by, and the pressure change is permanent (Koch et al. 1991). Recent observational studies (Weckwerth et al. 2004; Knupp 2006; Wilson and Roberts 2006; Koch et al. 2008a, b; Tanamachi et al. 2008; Martin and Johnson 2008; Hartung et al. 2010; Marsham et al. 2011) have revealed that atmospheric bores are generated by outflows from nocturnal convection and can, in turn, trigger strong vertical motions, resulting in intense upward displacements of air parcels (~0.5–1.5 km) in the lowest ~3 km and a sustained elevation of the wave duct. Whether the mechanism is a bore or a gravity wave, a wave duct can trap the waves by preventing the vertical propagation of significant wave energy out of the ducting layer (Lindzen and Tung 1976).

Recent highly idealized modeling studies have demonstrated the potential role of bores in the evolution of squall lines in the presence of an SBL (Parker 2008; French and Parker 2010). Parker (2008) found that as the stability increases and the convection becomes elevated, the mechanism responsible for lifting the parcels flowing into the system evolves from a surface cold pool to a feature Parker characterized as a bore with some gravity wave characteristics. Eventually, the interaction between the bore lifting drives the squall line, which then propagates at the speed of the bore (French and Parker 2010). In a related mechanism for the maintenance of convection, Fovell et al. (2006) proposed that lifting from high- and low-frequency gravity waves propagating ahead of the storm promotes new cell development by prepping the advance environment.

The idealized two-dimensional framework for the maintenance of nocturnal convection outlined by French and Parker (2010) does not address the three-dimensional variations of these nocturnal storms. The goal of our investigation is to advance our knowledge of the dynamics, structure, and evolution of nocturnal convective systems through examination of a simulated convective system from the perspectives of theory as applied to atmospheric bores (e.g., Scorer 1949; Rottman and Simpson 1989; Baines 1995) and frameworks often
utilized to explain the dynamics of convective systems (e.g., Rotunno et al. 1988; Bryan and Rotunno 2014; Peters and Schumacher 2016). Our approach allows for a more in-depth investigation of convective dynamics and the linkages to wave theory than the previously described idealized studies. Specifically, we investigated a nocturnal, elevated MCS event that occurred over the southern Great Plains during the night of 3–4 June 2013. The MCS initiated in the afternoon as surface-based storms associated with a well-defined surface cold pool. As the boundary layer stabilized at night and the LLJ intensified, the system transitioned from being surface-based to elevated and it produced waves, including bores. The MCS was simulated with the Weather Research and Forecasting (WRF) Model, allowing us to test the ability of a numerical model to recreate this observed transition and associated wave structure.

The paper is organized as follows. A description of the chosen case is given in section 2. Section 3 outlines the model configuration used in this study. Section 4 provides an overview of the simulation. Convective system dynamics and wave theory are applied in section 5 to explain the results presented in section 4, followed by conclusions in section 6.

2. Description of the observed case

A detailed synoptic analysis of this 3–4 June 2013 event can be found in Blake (2015). The storm was associated with a short-wave trough that moved through northern Oklahoma with a low pressure center over southeastern Colorado and western Kansas. A quasi-stationary, warm frontal boundary was observed across central Kansas with a well-defined LLJ developing over parts of Oklahoma and Texas. The nocturnal MCS occurred well to the south of this frontal boundary, which allows for insight into the mechanisms responsible for the initiation and maintenance of nocturnal convection apart from frontal ascent.

The convection in this study lasted for approximately 22 h, from the early afternoon of 3 June until the late morning of 4 June. The system became organized (Fig. 1a) with storms initiating in southwestern Kansas and the Oklahoma Panhandle around 1700 local standard time (LST)\(^1\) on 3 June near the intersection of the dryline and an outflow boundary from earlier convection in northwestern Kansas. Data from the Oklahoma Mesonet (Brock et al. 1995; McPherson et al. 2007) allowed us to document the evolution of surface quantities such as temperature, pressure, rainfall, wind speed, and wind direction and to determine whether the MCS was associated with a density current and/or a bore. Data from the Buffalo, Oklahoma, mesonet site (Fig. 2a) show that a well-defined surface cold pool was present early in the system’s life cycle as convection strengthened, moved to the east-southeast, and then turned more toward the south-southeast with time (Figs. 1b,c).

Within the NOAA Rapid Refresh (RAP) model, a strong gradient in convective available potential energy (CAPE) was evident in the southerly flow at 1 km AGL, the layer with the maximum CAPE (Fig. 3). The region of maximum CAPE at 1 km AGL also corresponded to reduced convective inhibition (CIN) (Fig. 3). Between 2300 and 0100 LST, the initial system dissipated as it moved out of this zone of high CAPE into less favorable conditions in central Oklahoma. Data from the Chickasha, Oklahoma, mesonet site, located in the central part of the state (Fig. 2b), show that from ~0000 to 0030 LST a pressure jump was accompanied by a minimal change in surface temperature, indicating the

\(^1\)This study is concerned with the diurnal cycle; hence, we used LST, where LST = UTC – 6 h.
feature that moved through was likely a bore. The surface observations of the bore correspond to the radar “fine line” depicted in Fig. 1c, which passed over Chickasha shortly after 0000 LST. Following the passage of the bore, the temperature decreased between 0030 and 0130 LST, consistent with the passage of the weak convective cold pool (Fig. 2b). There was no rainfall recorded at the Chickasha mesonet site because the earlier convection had dissipated.

As the initial convective system died out, the bore continued to propagate toward the southeast. Further exploration of this bore was undertaken through the utilization of retrieved temperature and humidity profiles obtained by the MP-3000A microwave radiometer located on the roof of the National Weather Center in Norman, Oklahoma. The general lack of cloud cover for this event increased the reliability of the microwave retrievals (Castleberry 2014). The retrieved profiles show pronounced cooling through a deep layer extending from 1 to 4 km above ground level (AGL) (Fig. 4a). The water vapor mixing ratio (g kg\(^{-1}\)) field reveals moistening of 6–7 g kg\(^{-1}\) of water vapor at heights between 500 m and 2 km AGL associated with the first two wave fronts of the bore (Fig. 4b). The joint presence of moistening and cooling suggests lifting in a stable environment where moisture decreases with height. Vertical displacements were also calculated using the observed temperature changes and assuming that any lifted parcels would follow the dry-adiabatic lapse rate (Fig. 4c). The maximum vertical displacements with the bore approached 900 m between 1 and 3 km with a net upward displacement of ~400 m with the passage of the bore. Figure 4d depicts the evolution of the potential temperature field with height and shows how the colder air in the SBL was indeed displaced upward by the bore. The magnitude of the lifting was consistent with the previously mentioned studies, although the lifting associated with the bore extended well beyond the SBL into the lower troposphere. Such deep lifting is likely to bring parcels closer to their level of free convection (LFC), which was ~4 km AGL over this region for parcels lifted from 1 km AGL, and reduce any CIN present in the

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**Fig. 2.** Time series data of surface temperature (°C; red curve), surface pressure (mb; dark yellow curve), accumulated rainfall (mm; shaded green), wind speed (m s\(^{-1}\); cyan curve), and wind direction (°; magenta curve) for (a) the Buffalo mesonet site from 1700 to 0400 LST and (b) the Chickasha mesonet site from 2300 to 0500 LST. The light blue arrows in (a) and (b) mark the passage of a cold pool and the orange arrows in (b) mark the passage of a bore.
environment. This lifting and its relationship to CAPE and CIN will be explored later within the context of our simulations.

During the dissipation of the initial system, new convection started to develop behind the leading system around 2200–2300 LST (Fig. 1c), a phenomenon known as rearward off-boundary development (Peters and Schumacher 2014). This new convection subsequently developed into a nocturnal, elevated MCS, which will be the primary focus of our study. Between 0100 and 0200 LST the system organized itself into two convective bands (Fig. 1d). The strongest storms occurred when these two convective bands merged between 0200 and 0400 LST. In this region, a second bore formed along the leading edge of the system. The data from the Chickasha mesonet site indicate that sharp increases in surface pressure were accompanied by changes in wind direction, an increase in wind speed, and an increase in surface temperature immediately followed by a sharp decrease and the onset of rainfall (Fig. 2b). Based on the expected structure of these systems, the surface data again most likely depict a bore propagating just ahead of the density current. The radar reflectivity illustrates the development of a broad region of trailing stratiform precipitation behind the leading edge between 0400 and 0600 LST (Figs. 1e,f). The MCS gradually dissipated between 0800 and 1000 LST. Unfortunately, the close proximity of this second bore to the MCS and its precipitation meant the event was associated with significant water vapor and high liquid water path values, preventing an accurate determination of the profiles of temperature and humidity from the microwave radiometer (Castleberry 2014).

3. Design of the simulation

The WRF is an NWP model that was developed as a collaborative effort among several institutions (e.g., National Center for Atmospheric Research, National Oceanic and Atmospheric Administration, Naval Research Laboratory, Federal Aviation Administration) and various university scientists (Skamarock et al. 2008). Our investigation used version 3.6.1 of the WRF with the advanced research core (WRF-ARW). The model employs terrain-following hydrostatic pressure coordinates on an Arakawa C grid. Our simulations were conducted for 22 h, beginning at 1200 LST 3 June and ending at 1000 LST 4 June, spanning the three domains shown in Fig. 5. Three domains were used in order to resolve convective processes for the feature of interest across the inner domain while maintaining computational efficiency. We utilized a two-way nest, in which the outer domain provided boundary values for the inner grid, and the inner grid could then influence the coarse outer domain. The outer domain contained 121 × 121 horizontal grid points with 9-km grid spacing, the middle domain contained 301 × 301 grid points with 3-km grid spacing, and the inner domain contained 499 × 529 grid points with 1-km grid spacing. The location of the 1-km domain was chosen to capture as much of the MCS evolution as possible given computing restraints. The 3-km domain was also crucial, as many of the wave features produced by the convection propagated out of the 1-km domain by 0800 LST. Rayleigh gravity wave damping was utilized aloft, beginning around 11 km AGL. We used 100 vertical levels for all three domains. The vertical grid spacing ranged from 65 m near the surface to 200 m beyond 1.5 km AGL with the first model level at 32 m AGL.

The lateral boundary conditions for the simulation were updated every hour utilizing RAP output obtained from the National Operational Model Archive and Distribution System (NOMADS) server. The RAP data did not contain soil information. Thus, we employed the Noah land surface model (Ek et al. 2003), taken from the North American Land Data Assimilation System, for these simulations. The physical parameterizations were chosen based on the
The Mellor–Yamada–Nakanishi–Niino (MYNN) model was chosen for the PBL scheme because the Mellor–Yamada–Janjić (MYJ) version did not accurately recreate the storm and Yonsei University (YSU) approach produced a warm surface temperature bias. The reader is referred to Blake (2015) for further results of the sensitivity tests.

**4. Overview of the simulation**

The WRF simulation generally performed well when compared to observations, as can be seen from comparisons of Figs. 1 and 6. In the observations and in the simulation, small, unorganized cells form around 1330 LST 3 June with the system organizing into an MCS around 1700 LST (Fig. 6a). The observations and the

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**Fig. 4.** Time–height cross sections of (a) retrieved perturbation temperature (K), (b) perturbation water vapor mixing ratio (g kg⁻¹), (c) vertical displacements (km), and (d) total potential temperature (K) derived from the MP-300A microwave radiometer retrievals [(a)–(c) taken from Castleberry (2014)]. The small symbol at ~0030 LST indicates a measurement and subsequent retrieval that may be impacted by an optically thick cloud with a high liquid water path (~500 g m⁻²). Vertical displacements in (c) were estimated using the perturbation temperatures and the dry adiabatic lapse rate. The displacements are calculated relative to the temperature at the initial times [the first ~10 min of T(z) data] in this cross section.
model have the convection moving to the east and southeast with the initial line dying out between 2300 and 0100 LST, and a second line of storms forming as the first line is diminishing, with the nocturnal MCS dissipating between 0900 and 1000 LST. One subtle difference in storm structure occurs near 0400 LST (Figs. 1e and 6e) when the model produces a relatively broad area of convection with reflectivities greater than 50 dB. In the observations, the region with reflectivities of 50 dB or greater was confined to a narrower west-southwest–east-northeast-oriented band at this time. This difference is a known problem with microphysical parameterizations (Morrison and Milbrandt 2015).

As expected, given the similarities discussed above, the model and the observations both depict similar regional environments, including a south-to-north corridor of high-CAPE values located over western Oklahoma and the Texas Panhandle that is in place from the start of the simulation until the passage of the nocturnal MCS (Figs. 3 and 7b). Since the observations suggest a transition from convection with surface-based convergence associated with a cold pool to lifting aloft partly associated with a bore, we investigated the evolution of the CAPE and CIN of the inflow air both at the surface and at 1 km AGL. We utilized the 3-km domain and focused on inflow conditions ahead of the system (i.e., Fig. 7) over a time span from 1700 LST 3 June to 0630 LST 4 June. The various storm inflow regions were determined subjectively. The inflow region for the surface flow is located southeast of the MCS with CAPE initially at 1200 J kg\(^{-1}\) with a CIN of \(-175\) J kg\(^{-1}\) (Fig. 8). As the night progressed, the surface flow stabilization and CAPE decreases to 400 J kg\(^{-1}\), while the magnitude of CIN increases to \(-500\) J kg\(^{-1}\) until 2100 LST, when the surface inflow lacks significant CAPE. The stabilization of the surface conditions is expected from nocturnal radiational cooling and the formation of an SBL.

The conditions aloft at 1 km AGL are quite different (Fig. 8). The inflow region at this height is located nearly due south of the MCS, with CAPE and CIN initially quite close to the surface values at 1700 LST, suggesting that the 1-km values were within the convective boundary layer. Subsequently, CAPE decreases and the magnitude of CIN only increases to \(-200\) J kg\(^{-1}\) so that by 2130 LST, there is no longer any significant CAPE within the 1-km inflow. This pronounced decrease in the CAPE precedes the dissipation of the first system. After convective cells develop into the second MCS, CAPE increases to over 2000 J kg\(^{-1}\) and the magnitude of CIN decreases to \(-80\) J kg\(^{-1}\) by 0230 LST. For a given amount of lifting, it would be easier to initiate and sustain deep convection within these nocturnal conditions than those observed in the PBL during the day. By 0630 LST there is no longer any inflow with significant CAPE or CIN into the MCS at 1 km.

The high values of CAPE at 1 km are primarily due to large values of the water vapor mixing ratio within the strong wind speeds associated with the LLJ (Fig. 9). The time of peak wind speeds, \(\approx 0230\) LST, corresponds to the time at which CAPE in the 1-km inflow is the greatest. As discussed earlier, the LLJ is known to provide an elevated source of moist, unstable air. Thus, advection associated with the nocturnal LLJ brought moist, unstable air into the storm, which played a crucial role in creating a high-CAPE inflow and likely impacted the longevity of the MCS.

![Fig. 5. Domain configuration used by the WRF Preprocessing System (WPS) for the model runs.](image)

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Fig. 6. Composite reflectivity (dBZ) generated from the WRF Model at (a) 1700, (b) 2000, (c) 2330, (d) 0130, (e) 0400, and (f) 0700 LST in the 3-km domain. Compare to the observed reflectivities in Fig. 1. The black triangle in (d) represents the location of Chickasha.
To examine the structure of the storm relative to the advection of CAPE with the LLJ, we investigated the potential temperature difference between the cold pool and the ambient air ahead of the storm at the surface and 1 km AGL. The potential temperature difference was calculated within the 3-km domain in a framework moving with the MCS (Fig. 10). From 1700 LST 3 June to 2030 LST 4 June, the difference is stronger at the

![Horizontal cross section of reflectivity (dBZ), wind vectors, and CAPE (J kg⁻¹) in the 3-km domain at 1 km AGL valid at (a) 1930 and (b) 0130 LST. Reflectivity contours are drawn from 0 to 70 dBZ with a contour interval of 10 dBZ. The thick black line represents the line used to calculate CAPE and CIN in the 1-km inflow region.](image)

**Fig. 7.** Horizontal cross section of reflectivity (dBZ), wind vectors, and CAPE (J kg⁻¹) in the 3-km domain at 1 km AGL valid at (a) 1930 and (b) 0130 LST. Reflectivity contours are drawn from 0 to 70 dBZ with a contour interval of 10 dBZ. The thick black line represents the line used to calculate CAPE and CIN in the 1-km inflow region.

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![Three curves that illustrate how CAPE and CIN evolve in three distinct inflow regions with time at the surface and at 1 km AGL. An inflow region is defined as the region containing the air flowing into the storm, and its location changes with height. A 130–150-km line was manually drawn across the various inflow regions every 30 min from 1700 LST 3 Jun to 0630 LST 4 Jun. CAPE and CIN were calculated by averaging their values at nine points along each line, roughly every 15 km. The red curve is for the surface inflow region for the first system, the blue curve is for the 1-km inflow region for the first system, and the cyan curve is for the 1-km inflow region for the second system.](image)

**Fig. 8.** Three curves that illustrate how CAPE and CIN evolve in three distinct inflow regions with time at the surface and at 1 km AGL. An inflow region is defined as the region containing the air flowing into the storm, and its location changes with height. A 130–150-km line was manually drawn across the various inflow regions every 30 min from 1700 LST 3 Jun to 0630 LST 4 Jun. CAPE and CIN were calculated by averaging their values at nine points along each line, roughly every 15 km. The red curve is for the surface inflow region for the first system, the blue curve is for the 1-km inflow region for the first system, and the cyan curve is for the 1-km inflow region for the second system.
surface than at 1 km AGL (Fig. 11a). As the night progresses, the situation changes. Beginning at 2100 LST, the potential temperature difference is greater at 1 km AGL than at the surface. The difference increases to as large a magnitude as 6°C between 0300 and 0400 LST during the passage of the nocturnal, elevated MCS. Between 0000 and 0030 LST, the surface cold pool from the first convective system dies out and the cold pool from the second system does not form until 0100 LST. This gap in the occurrence of the cold pool warranted the exclusion of difference values at 0000 and 0030 LST. At 1 km AGL, the cold pool from the first system also diminishes around 0000 LST, but the cold pool from the second system has formed by that time. The second convective system is the result of elevated convection initiation, which explains why it possesses a stronger potential temperature difference at 1 km AGL and later develops a surface-based cold pool. A comparison between the surface potential temperature difference observed in the RAP BUFR data and simulated by the WRF (Fig. 11b) suggests extremely close agreement, lending credence that the simulated MCSs have characteristics similar to the observed convection.

To explore the generation of the cold air aloft, we examined the evolution of the reflectivity and vertical velocity fields at 1 km AGL. The horizontal cross sections reveal a wave pattern in the vertical motions that forms during the mature stage of the nocturnal, elevated MCS (Fig. 12). Data at 5-min intervals from WRF were used to observe how variables such as temperature, pressure, accumulated rainfall, wind speed, and wind direction evolved with time at the first model level at Chickasha. The model output (Fig. 13) indicates a 1.6-mb (1 mb = 1 hPa) pressure jump between 0345 and 0350 LST that is associated with no appreciable change in temperature, wind speeds of ~15 m s⁻¹, and an abrupt shift in wind direction. This feature is followed by a 2.5°C increase in temperature from 0350 to 0405 LST, which is then followed by a 3.7°C drop in temperature and a 1.3°C increase in dewpoint from 0405 and 0410 LST with the onset of rainfall a few minutes later. According to past observations of bores (Koch et al. 1991), these data suggest that the feature is most likely a bore just ahead of the density current and the convection, which is also in agreement with the Chickasha surface observations (Fig. 2b). From past observational studies, the warming with and following the passage of the wave is likely due to the downward mixing of stable air. Vertical cross sections of buoyancy were created in two directions using the 1-km domain to get a better sense of

Fig. 9. As in Fig. 7b, but for the mixing ratio (g kg⁻¹).
how the system structure varied from south to southeast. The orientations selected for the vertical cross sections are north–south (N–S) and northwest–southeast (NW–SE) (Fig. 12a). We defined the buoyancy field as follows:

\[ B = g \left[ \frac{\theta' + 0.61(q_v - \bar{q}_v) - q_c - q_r - q_i}{\theta} \right], \]  

where \( \theta \) and \( \bar{\theta} \) represent the potential temperature and water vapor mixing ratio of the ambient air ahead of the density current, \( \theta' \) is the difference between \( \theta \) and \( \bar{\theta} \), \( q_v \) is the cloud water mixing ratio, \( q_r \) is the rainwater mixing ratio, and \( q_i \) is the mixing ratio of frozen condensate. The terms in the buoyancy equation were calculated as a function of height relative to an ambient air mass well...
ahead of the system. The ambient air terms were calculated in the 1-km domain at each model level using an area average over a $10\text{km} \times 10\text{km}$ box centered in the ambient air, which was defined to be at least $10\text{km}$ ahead of the leading edge of the lifting in the cross section. Note that the vertical acceleration due to the effective buoyancy also includes the vertical gradient in the buoyancy perturbation pressure gradient (e.g., Doswell and Markowski 2004). The calculated distribution of buoyancy would, along with the dynamic terms, contribute to pressure perturbations that will further impact the organization and structure of the system. The perturbation pressure field was not diagnosed in this study.

The relatively deep layers of negatively buoyant air (Fig. 14) ahead of the density current at the surface in both vertical cross sections are consistent with the net upward displacement of a stable air mass, as would be expected from a borelike disturbance. The boundary layer air in Fig. 14 is indeed stable, as evidenced by the sharp increase of potential temperature with height, but the stability varies between the two cross sections with a deeper nocturnal stable layer evident along the N–S vertical cross section. The conclusion of a wave response with the negative buoyancy associated with the lifting of stable air is supported by the displacement of the isentropes in these figures. Significantly more positive buoyancy is found above $4\text{km}$ in the NW–SE cross section, with the location of the buoyant plume collocated with the leading edge of the density current, corresponding to a stronger updraft in this region.

The vertical cross sections of vertical velocity in Fig. 15 that correspond to the two vertical cross sections in Fig. 14 provide insight into differences found in the buoyancy field. In both vertical cross sections, the lifting extends well above the nocturnal SBL, which was generally less than $500\text{m}$ (refer to Fig. 53 in Blake (2015) and the potential temperature fields in Figs. 14 and 15). Along the N–S cross section (Fig. 15a), the strongest ascent at the leading edge of the system occurs below $2\text{km}$, with additional lifting extending over the lowest $5\text{km}$. This lifting is paired with strong downward motion followed by an alternating ascent and descent pattern above the density current in the lowest $3\text{km}$. The pattern of descent and ascent near the leading edge of the system extends well into the troposphere. The vertical motion pattern is more complex aloft within the layers of weak positive buoyancy. In contrast, the convectively active, NW–SE vertical cross section (Fig. 15b) reveals that the height of the strongest ascent increases with rearward extent, with the largest vertical motions near the leading edge of the positive buoyancy aloft (see Fig. 14b). A more in-depth analysis of the evolution and dynamics of the wave field will be undertaken in the future. We will continue to refer to this wave as a bore, since the wave is associated with net displacements of stable air in the lower troposphere, forms along and moves ahead of the leading edge of the convection, and produces variations in the surface fields consistent with bores.

Recent observational (Bryan and Parker 2010) and modeling (Bryan and Morrison 2012) studies have
shown that cold pools with a depth of 3–4 km do occur in surface-based storms, so the deep layers of negative buoyancy in Fig. 14 do not necessarily indicate the system is elevated. Previous investigations (e.g., Colman 1990) have somewhat arbitrarily selected a few layers aloft to determine if convective instability is present for elevated convection rather than calculating the full vertical profile of CAPE and CIN. Such approaches would often underestimate or even miss the presence of other elevated unstable layers. In this study, vertical cross sections of CAPE and CIN for both orientations were calculated from the model data to demonstrate that the simulated MCS is elevated (Fig. 16). In the N–S orientation (Figs. 16a,c), the ambient environment depicts high CIN confined to the lowest ~300 m, with the highest CAPE located in the layer between ~300 m and ~1 km. As the bore passes by, the high-CAPE layer is displaced upward to ~1.5–2 km and the CIN decreases as is expected from this ascent. The lifting is not sufficient for the CAPE to be fully realized. For the NW–SE orientation (Figs. 16b,d), the ambient environment shows large CIN below ~1 km associated with the nocturnal PBL topped by large values of elevated CAPE above ~1.5 km. While the NW–SE orientation has more CIN in the SBL, the magnitude of CIN within the layer of maximum CAPE is actually smaller than in the N–S cross section. The lifting by the bore again reduces the CIN, as parcels are brought closer to their LFC. In this NW–SE cross section, the net displacement of the high-CAPE layer leads to buoyant convection.

Previous studies (e.g., Koch et al. 2008a; Coleman and Knupp 2011) have suggested that lifting by bores creates favorable conditions by reducing CIN. However, these studies utilized values of CAPE and CIN either at the surface or within the stable boundary layer, which does not detect the important processes and highly favorable conditions aloft in association with the LLJ. A related study (Peters and Schumacher 2016) showed that flow with an SBL interacting with a cold pool led to an increase in CIN and a decrease in CAPE as the flow encountered and passed over the outflow boundary. In our study, all of the CAPE is ingested by the system on its southeastern flank, and the CAPE and CIN go to 0 (Figs. 16b,d). On the southern flank, CAPE decreases as the flow encounters and passes over the cold pool, while CIN remains about the same (Figs. 16a,c). We will now further explore the observed variation in structure around the cold pool through the application of convective system dynamics and wave theory in an effort to better understand the structure and maintenance of nocturnal convective systems.

5. Dynamical inferences

a. Convective system dynamics

Rotunno et al. (1988), often referred to as RKW (for Rotunno, Klemp, and Weisman), proposed a theory for the optimal conditions for a vertically oriented updraft along outflow boundaries. The optimal state for convection, consisting of vertically oriented updrafts, occurs when the negative horizontal vorticity produced baroclinically by the cold pool is balanced by the positive horizontal vorticity associated with the ambient vertical wind shear, resulting in a steady-state momentum force balance (Bryan and Rotunno 2014). This balance is indeed favorable for convection, as it produces deeper and stronger ascent of warm air at the leading edge of the cold air (Parsons 1992). While the interaction between the vertical shear and the cold pool has a pronounced impact on the vertical motions along the cold pool, a number of factors can influence the intensity and
duration of convective systems (e.g., magnitude of the CAPE and CIN, midlevel humidity, distribution of diabatic sources within the convective system, deep vertical shear) so that the use of RKW to determine the intensity or longevity of convection has been questioned (e.g., Stensrud et al. 2005; Coniglio et al. 2012).

The analysis of contrasting wind shear scenarios that follows is similar to that of Trier et al. (2010) and Peters and Schumacher (2015a,b). In order for there to be positive horizontal vorticity, $dU/dz$ must be greater than 0, where $U$ represents the component of the wind speed perpendicular to the cold pool in the cross section. Within this framework, positive $dU/dz$ means that it is oriented away from the cold air. Figure 17 from Bryan and Rotunno (2014), illustrates that for any positive $c/\Delta U$ ratio, where $c$ is the cold pool intensity and $\Delta U$ is the wind speed difference across the shear layer, there will be an upward parcel displacement. Negative $c/\Delta U$ ratios would produce much smaller parcel displacements (Moncrieff and Liu 1999). French and Parker (2010) discuss how the shear in the effective inflow layer containing a maximum in CAPE and minimum in CIN is more important for elevated systems than the shear in the SBL. Therefore, we will focus on the integrated vertical wind shear over the layer of maximum CAPE. A positive integrated $dU/dz$ will be classified as favorable, while a negative integrated $dU/dz$ will be classified as unfavorable. This favorable and unfavorable classification is similar to that of Peters and Schumacher (2015a,b), who focused on the orientation of $\Delta U$ with respect to the outflow boundary. These studies found that $\Delta U$ was oriented toward the cold pool on the southwestern flank.

FIG. 14. Vertical cross sections of the buoyancy field (m s$^{-2}$) in the 1-km domain at 0430 LST for the (a) N–S and (b) NW–SE orientations. The contours represent isentropes, and the ground-relative wind vectors projected onto the path of the cross section are also drawn. The horizontal black lines denote the density current head and the vertical cyan lines mark the leading edge of the convection. The arrows indicate the density current height $d_o$ and the inversion height $h_o$. The path of each cross section is shown in Fig. 12a.

FIG. 15. As in Fig. 14, but for vertical velocity.
of the MCS (unfavorable), but $\Delta U$ pointed toward the warm air on the southeastern flank (favorable).

Vertical profiles of the wind speed and vertical wind shear in the ambient environment along the N–S and NW–SE orientations are shown in Fig. 17 for 0400 LST. Both orientations depict the expected vertical shear profile for an elevated system with negative vertical wind shear associated with the LLJ beneath positive vertical wind shear. However, the integrated vertical wind shear over the layer of maximum CAPE is quite different for each orientation. On the south flank of the system, most of the CAPE is located between 0 and 1.5 km AGL (Fig. 16a), and the integrated $dU/dz$ is $-0.217\text{s}^{-1}$ over this layer. Conversely, the strongest CAPE along the system’s southeast flank is concentrated between 1.5 and 3 km AGL (Fig. 16b), corresponding to an integrated vertical wind shear of $0.077\text{s}^{-1}$. The NW–SE orientation has a positive value for integrated $dU/dz$, which is quite favorable for strong lifting along the outflow/bore, while the N–S orientation has a negative value for integrated $dU/dz$, which is unfavorable.

To supplement our vertical wind shear analysis, $\Delta Z_{LFC}$, the distance a parcel must be vertically displaced to reach its LFC, was computed for both orientations (Peters and Schumacher 2016). First, $\Delta Z_{LFC}$ was calculated for the southeastern flank to clarify why the bore eventually moves out ahead of the convection and why the convection stays coupled to the elevated density current. The vertical displacements of the isentropes were utilized to obtain an estimate for how far upward the parcels with the highest CAPE are displaced as they ascend over the bore. If the upward displacement of parcels along the bore is smaller than their $\Delta Z_{LFC}$ values ahead of the outflow, then the bore is insufficient to trigger convection, and parcels instead continue to travel rearward of the bore until the density current lifts them to their LFCs. The parcels with the highest CAPE ahead of the system are located ~2 km AGL for the NW–SE orientation (Fig. 16b), and they are displaced approximately 500 m by the bore. The LFC ahead of the outflow in the NW–SE orientation is approximately 3 km, yielding a $\Delta Z_{LFC}$ of 1 km for the parcels originating at 2 km AGL. The lifting associated with the bore occurs primarily below 2 km AGL on the system’s southeastern flank, and since most of the CAPE is above this layer, the bore does not trigger convection and it slowly pulls away from the system. Destabilization by the bore drops the LFC from 3 to 2.75 km. The parcels at 2.5 km AGL, which now have a $\Delta Z_{LFC}$ of 0.25 km, are then displaced 1 km by the elevated density current, as
indicated by the isentropes. This displacement is greater than $\Delta Z_{LFC}$, and convection is observed as a result (Fig. 18a).

Here $\Delta Z_{LFC}$ was also calculated for the southern flank to illustrate why convection does not continue to the south (Fig. 18b). Ahead of the outflow on the N–S orientation, the LFC is approximately 3 km. Parcels with the highest CAPE are located ~0.5 km AGL (Fig. 16a), yielding a $\Delta Z_{LFC}$ of 2.5 km. The parcels are displaced 1 km by the bore, which is a larger displacement than on the NW–SE orientation yet still not sufficient to reach the LFC, which drops from 3 to 2.75 km after the passage of the bore. The parcels at 1.5 km, which now have a $\Delta Z_{LFC}$ of 1.25 km, are then displaced 0.5 km by the density current. Therefore, convection is not favored to move to the south, which is correct as it primarily continues to the southeast.

b. Hydraulic theory

Hydraulic theory is utilized to show that bores are an expected response from the interaction between the cold pool and the ambient environment. We will approximate our environment as a two-layer, inviscid flow over an obstruction. Following Koch et al. (1991), we will assume that the density current acts in an analogous manner to an obstruction. Specifically, four flow regimes are possible when a density current encounters a surface-based stable layer: supercritical, subcritical, partially blocked, and completely blocked flow (Rottman and Simpson 1989; Koch et al. 1991). The reader is referred to Fig. 16 in Koch et al. (1991) for a more detailed description of the flow types. Bores will be initiated in the partially or completely blocked flow regimes. It has been shown that the possible time-independent solutions of the shallow water equations are determined by two nondimensional parameters, $D_o$ and $F$ (Houghton and Kasahara 1968):

$$D_o = \frac{d_o}{h_o}$$

$$F = \frac{(U - C_{dc})}{C_{gw}} = \frac{(U - C_{dc})}{\sqrt{g \left(\frac{\Delta \theta_{vm}}{\theta_{vm}}\right) h_o}}$$

where $D_o$ is the nondimensional height, defined as the ratio of the density current height $d_o$ to the height of the inversion of the ambient air $h_o$. The value of $d_o$ was set equal to the height above the density current at which $\theta_e$ became greater than the surface $\theta_e$ of the ambient air. If there were no clouds present, then the virtual potential temperature would have been sufficient. The level at which $\frac{d\theta}{dz}$ first drops below 0.005 K m$^{-1}$ is represented by $h_o$. The lapse rate must also be less than 0.005 K m$^{-1}$ over a layer at least 200 m thick. This criterion eliminates noise in the vertical profile of $\frac{d\theta}{dz}$ from qualifying as the inversion height. In addition, $F$ is the Froude number, which is the ratio of the density current speed $C_{dc}$ to the phase speed of a long, internal gravity wave ($C_{gw}$) confined within the stable layer, and $C_{dc}$ was

![Fig. 17. Vertical profiles of the horizontal wind speed (m s$^{-1}$) of the ambient air in the 1-km domain at 0400 LST for the (a) N–S and (b) NW–SE orientations. The vertical wind shear (s$^{-1}$) is color contoured in the background, and the integrated vertical wind shear is shown over the layer of maximum CAPE.](image-url)
estimated for both orientations using vertical cross sections of potential temperature to track how far the interface between the warmer, ambient air and the colder, density current air moved over time. The virtual potential temperature at the surface for the ambient air is $u_{vw}$, and $D_{u_{vw}}$ is the inversion strength, which is the difference in $u_{x}$ from the surface to the inversion top of the ambient air.

Table 2 provides a list of the flow regimes, estimated $C_{dc}$, $d_{o}$, and mean values of $h_{o}$, $\Delta u_{vw}$, $D_{o}$, and $F$ for both orientations. According to these calculations, a partially blocked flow regime is occurring in both the N–S and NW–SE orientations at 0415 LST (Fig. 19), which means that a bore should be generated. In addition to $d_{o}$, $D_{o}$, $F$, and $C_{dc}$, the bore strength $b_{str}$, bore depth $h_{1}$, and phase-corrected bore speed $C_{bore}$ are also calculated. Refer to Blake (2015) for values of these quantities.

c. Linear wave theory

According to linear wave theory, in order to maintain a long-lived wave, some mechanism must exist in order to trap wave energy and prevent the vertical propagation of wave energy out of the layer (Crook 1988; Koch and Clark 1999). One mechanism by which waves can horizontally propagate is through internal reflection. The equation for the vertical wavenumber $m^{2}$ is given by

$$m^{2} = \frac{N^{2}}{(U - C_{bore})^{2}} - \frac{\partial^{2}U}{\partial z^{2}} - k^{2}, \quad (4)$$

where $N^{2}$ is the square of the Brunt–Väisälä frequency and the other terms have been previously defined (Scorer 1949; Crook 1988). The Scorer parameter $l^{2}$, given by the first two terms on the right-hand side of the following equation, is often utilized to diagnose the probability of a wave duct. Given a packet of waves encompassing a spectrum of wave modes (each with its own specific horizontal and vertical wavelength), as $l^{2}$ decreases with height, some of the wave modes are trapped under an exponentially decaying layer, while the rest of the wave modes continue to vertically propagate.

The Scorer parameter was plotted with height at 0415 LST in order to determine if an appropriate wave duct is
in place. The following phase-corrected bore speeds were utilized: 13.33 m s\(^{-1}\) for the N–S orientation and 15.71 m s\(^{-1}\) for the NW–SE orientation (Blake 2015). Both the N–S and NW–SE orientations contain a layer of positive Scorer parameters in the lowest kilometer, topped by a layer with a negative Scorer parameter centered around 1 km (Fig. 20). If we estimate the mean positive and negative Scorer layers to be \(l_1 = 5 \times 10^{-5}\) and \(l_2 = -0.12 \times 10^{-5}\) for the N–S orientation and \(l_1 = 0.6 \times 10^{-5}\) and \(l_2 = -0.09 \times 10^{-5}\) for the NW–SE orientation, both with \(z_1\), the height of the wave duct, \(\sim 800\) m, and the square of the horizontal wavenumber of \(k_x^2\), \(\sim 0.4 \times 10^{-7}\), then according to (4) a wave with a wavelength of 10 km should be trapped, which is roughly consistent with the simulations (Figs. 14 and 15). However, Baines (1995) assumed that the second negative Scorer layer is infinite; thus, the vertical wave motion in the infinite layer should approach 0. The negative Scorer layer is not infinite in the simulations, and, in accordance with theory, the waves associated with the bore do not appear to be completely trapped within the wave duct for the NW–SE orientation, as the vertical motions extend up to \(\sim 8\) km with the wave ducts located at \(\sim 0.75–1.25\) and \(\sim 2.25–2.75\) km AGL. This finding is also supported by the observations provided by the microwave radiometer retrievals (Fig. 4), which also suggest lifting throughout the lower troposphere. We do note, however, that the retrievals become less dependable with increasing elevation.

One noticeable difference exists between the two orientations. The N–S orientation contains seven layers of positive Scorer below negative Scorer extending up to 8 km, while the NW–SE orientation contains three layers of positive Scorer below negative Scorer located below 4.5 km. In addition, the layer of negative Scorer at \(\sim 4.5\) km AGL for the NW–SE orientation is very weak. Our observations of complex, layered wave motion particularly in the N–S cross section, which we described in section 4, appear consistent with multiple positive–negative Scorer layers. Further analysis based on linear wave theory to explain these multiple layers and how they affect vertical motions and wave structures will be conducted in the future. Given that our previous analysis revealed the southern flank of the storm is less favorable for lifting, these results confirm that the southern edge of the MCS is more likely to be characterized by a lack of convection, but potentially more pronounced wave activity.

### Table 2

<table>
<thead>
<tr>
<th>Orientation</th>
<th>N–S</th>
<th>NW–SE</th>
</tr>
</thead>
<tbody>
<tr>
<td>(h_0) (m)</td>
<td>979.65</td>
<td>1285.81</td>
</tr>
<tr>
<td>(\Delta \theta_{av}) (K)</td>
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<td>11.28</td>
</tr>
<tr>
<td>Estimated (C_{av}) (m s(^{-1}))</td>
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<td>15.71</td>
</tr>
<tr>
<td>Estimated (d_0) (m)</td>
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<td>1977.32</td>
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<tr>
<td>(D_0)</td>
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<td>1.54</td>
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<tr>
<td>(F)</td>
<td>1.67</td>
<td>1.75</td>
</tr>
<tr>
<td>Flow regime (SP, PB, CB, SB)</td>
<td>PB</td>
<td>PB</td>
</tr>
</tbody>
</table>

**FIG. 19.** Plot of the nondimensional height (\(D_0\)) and the Froude number (\(F\)) at 0415 LST for the N–S (red circle) and NW–SE (blue circle) orientations. The \(D_0, F\) parameter space indicates the flow regime that is occurring: a, supercritical; b, partially blocked; c, completely blocked; and d, subcritical.
6. Summary and conclusions

Forecast skill for nocturnal convection remains relatively low, and one reason is the difficulty in accurately simulating storm systems associated with the greater occurrence of elevated convection at night. This study utilized version 3.6.1 of the WRF-ARW Model to simulate a nocturnal MCS event that occurred over the southern Great Plains on 3–4 June 2013. The system was well simulated, as evidenced by the surface data, radar, and microwave radiometer retrievals being in relatively good agreement with the simulation.

The MCS in this study began as a surface-based system driven by the lifting of high-CAPE and low-CIN boundary layer air. As the night progressed, nocturnal radiational cooling stabilized the PBL, diminishing the CAPE and CIN in the storm’s surface inflow region. However, the advection of moist, unstable air associated with the nocturnal LLJ created an inflow region of high CAPE and low CIN at the 1-km level, and conditions aloft became increasingly favorable for convection. This finding is in agreement with previous observations of nocturnal convection and the idealized work (e.g., French and Parker 2010), which found the source of parcels with high CAPE to be 1 km AGL or higher for elevated convection. From the cold pool analysis, the potential temperature difference became significantly stronger aloft than at the surface as the MCS transitioned into its nocturnal, elevated stage. This observation is also consistent with previous studies that claim the SBL inhibits deep penetrative downdrafts from reaching the surface, preventing the formation of a surface cold pool and indicating the lesser importance of a surface cold pool in maintaining convection at night (e.g., Trier and Parsons 1993; Parker 2008).

The storm’s outflow varied in three dimensions and our application of theories of convection and wave dynamics provided some insight into these variations. Shallower ascent was likely along the southern flank of the MCS, as expected from the vertical shear and $\Delta Z_{LFC}$ analysis. The convectively active southeastern flank had a layer of elevated CAPE that interacted with a deep (4–5 km) zone of negative buoyancy to produce lifting. While the bore on the southeastern flank did not trigger convection by itself, it assisted in transporting high-CAPE air toward its LFC, and the $\Delta Z_{LFC}$ analysis revealed that the additional lifting by the elevated density current allowed for deep convection to occur. As a result, the bore was not coupled to the convective system and it slowly pulled away, while the convection remained in phase with the elevated density current.

![FIG. 20. Two lines that illustrate how the Scorer parameter ($\times 10^{-5} \text{ m}^{-2}$) changes with height at 0415 LST for the N–S (red line) and NW–SE (blue line) orientations. The vertical black line represents a Scorer parameter of 0 m$^{-2}$.](image)
Several differences exist between our results and previous idealized simulations. For example, in the Parker (2008) and French and Parker (2010) studies, the convection became elevated and associated with a borelike feature only after nocturnal cooling eliminated CAPE within the SBL. In our case, a bore response was generated in the presence of low-level CAPE. In addition, the soundings used in the idealized Parker simulations were for severe convection, while our case was more consistent with the class of less severe nocturnal MCSs that were investigated in the Haghi et al. (2017, hereafter HPS) systematic study of IHOP 2002 (Weckwerth et al. 2004). The MCS in this study was associated with a well-defined maximum in CAPE and low CIN above the SBL, with greater CIN in the SBL compared to the Parker simulations, favoring elevated convection in our case and surface-based convection in the Parker simulations.

Another key difference is that the Parker simulations had convection staying “locked on” to the singular wave response, which was not generated by the cold pool until quite late in the simulations. In our result, the presence of partially blocked flow generated a bore response much earlier on in the simulation that moved slowly away from the density current. This difference suggests that the Parker simulations do not fall in the partially blocked flow regime that is frequently observed in the nocturnal environment. In unpublished work by the second author and colleagues (Parsons et al. 2017, manuscript submitted to *J. Atmos. Sci.*), it is shown that the idealized Parker simulations generally fall in the supercritical flow regime, which is represented by region A in Fig. 19 and is located outside the distribution of observed flow regimes in the HPS systematic study. Forward propagation speeds for MCSs differ substantially, depending on whether they are coupled with a bore or a density current. The density current speeds in our study ranged from 13 to 16 m s\(^{-1}\), in contrast to between 20 and 30 m s\(^{-1}\) in the Parker simulations. Determining whether convection phases with a bore or a density current in an environment where both are present is therefore of practical importance to weather forecasting. These results are a step toward improving our understanding of convective regimes where bores and cold pools are both present.

Other studies of nocturnal convection have noted that low-frequency waves contribute to lifting and moistening the ambient air ahead of the system (Fovell et al. 2006; Parker 2008; Adams-Selin and Johnson 2013). While some low-frequency waves were evident that did move ahead of the system, the convection in this study did not display a clear discrete propagation as described by Fovell et al. (2006). Although these low-frequency waves appear to be worth investigating, they were not addressed in this paper.

Follow-on studies to this work include a systematic examination of conditions over the Great Plains showing that partially blocked flow regimes with favorable wave trapping and bores are common over the region (HPS) and a more general exploration of the characteristics of lifting in association with bores and its relationship to deep convection (Parsons et al. 2017, manuscript submitted to *J. Atmos. Sci.*). Also of interest is repeating the simulations using other model parameterizations (i.e., the NSSL two-moment microphysics scheme, which also produced an accurate representation of the event). The recent Plains Elevated Convective At Night (PECAN) field campaign (Geerts et al. 2017) will allow us to apply this analysis to a greater number of cases.

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