Mesoscale Vertical Motions near Nocturnal Convection Initiation in PECAN

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(Manuscript received 8 January 2017, in final form 29 March 2017)

ABSTRACT

Radiosonde measurements from the Plains Elevated Convection At Night (PECAN) 2015 field campaign are used to diagnose mesoscale vertical motions near nocturnal convection initiation (CI). These CI events occur in distinctly different environments including ones with 1) strong forcing for ascent associated with a synoptic cold front and midtropospheric short wave, 2) nocturnal low-level jets interacting with weaker quasi-stationary fronts, or 3) the absence of a surface front or boundary altogether. Radiosonde-derived vertical motion profiles in each of these CI environments are characterized by low- to midtropospheric ascent. The representativeness of these vertical motion profiles is supported by distributions of corresponding mesoscale averages from model-produced 0–6-h ensemble forecasts. Thermodynamic data from radiosondes are then analyzed along with selected model ensemble members to elucidate the role of the vertical motions on subsequent CI. In a case with strong forcing for mesoscale ascent, vertical motions facilitated CI by reducing convection inhibition (CIN). However, in the majority of cases, weaker but persistent vertical motions contributed to the development of elevated, approximately saturated layers with lapse rates greater than moist adiabatic. Such layers have negligible CIN and, thereby, the capacity to support CI even without strong finescale triggering mechanisms in the environment. This aspect may distinguish much central U.S. nocturnal CI from typical daytime CI. The elevated unstable layers occur in disparate large-scale environments, but a common aspect of their development is mesoscale ascent in the presence of warm advection, which results in upward transports of moisture (contributing to local increases of moist static energy) with adiabatic cooling above.

1. Introduction

Central U.S. warm-season precipitation frequencies have a pronounced diurnal maximum at night (e.g., Wallace 1975). Most of this nocturnal precipitation occurs within mesoscale convective systems (MCSs) including mesoscale convective complexes (MCCs; Maddox 1980; Fritsch et al. 1986), which are supported by the nocturnal low-level jet (LLJ; Blackadar 1957; Holton 1967; Du and Rotunno 2014; Shapiro et al. 2016), and warm advection with mesoscale ascent (Maddox 1983) in lower-tropospheric baroclinic zones. Using data from the Oklahoma–Kansas Preliminary Regional Experiment for STORM-Central (PRE-STORM; Cunning 1986), several studies (e.g., Blanchard 1990; Stumpf et al. 1991; Fortune et al. 1992; Smull and Augustine 1993) examined mature stage MCSs whose overall convective structure is more complex (e.g., three-dimensional) than that in typical squall lines. Such systems often include elevated convection (e.g., Corfidi et al. 2008), where the source of air participating in deep convective updrafts originates from above the PBL.

The Plains Elevated Convection At Night (PECAN) field campaign conducted from 1 June to 16 July 2015 (Geerts et al. 2017) took advantage of recent innovations in observing technologies and design, along with advances in numerical modeling systems, to revisit the central U.S. nocturnal convection problem. PECAN concentrated on a wide range of topics influencing elevated nocturnal convection, including convection initiation (CI). Forecasting elevated nocturnal CI is particularly challenging since it frequently occurs in the absence of identifiable surface boundaries (Wilson and Roberts 2006), which provide useful visual cues that often aid forecasting of daytime CI.

Carbone et al. (2002) illustrated that nocturnal convection in the central United States often originates as eastward-propagating thunderstorms in the lee of the
Rocky Mountains much earlier in the diurnal cycle. However, central U.S. nocturnal convection may also initiate locally (Carbone and Tuttle 2008), which has been documented in numerous case studies. The circumstances and organization of this local CI over the central United States vary widely among cases. Local CI can begin as isolated thunderstorm clusters north of quasi-stationary surface fronts before eventually organizing upscale into MCSs or MCCs (e.g., Trier and Parsons 1993). While in other cases, mesoscale regions of new convection occur in proximity to mature eastward-propagating MCSs. For example, new bands of convection often initiate in broad zones of warm advection ahead of, and perpendicular to, preexisting MCSs (e.g., Fortune et al. 1992; Smull and Augustine 1993; Coniglio et al. 2012). In other cases new CI may manifest as mesoscale bands perpendicular to, but rearward of, preexisting MCSs (e.g., Trier et al. 2010; Keene and Schumacher 2013; Peters and Schumacher 2015, 2016). In such cases the nocturnal LLJ interacts with the lower-tropospheric baroclinic zone, which is intensified by MCS outflow.

In the current study we use PECAN radiosonde data and a 50-member model ensemble to examine effects of mesoscale vertical motions on nocturnal elevated CI. Model-based budget studies (Trier et al. 2014a,b) have illustrated the critical role of mesoscale vertical motions in reducing negative buoyancy in environments of both initiating and mature elevated MCSs. Both upward transports of moisture that raise the relative humidity of convecting air parcels and adiabatic cooling from vertical displacements in layers above these parcels can be important in reducing the negative buoyancy. These effects of vertical motion can promote CI by saturating inflow layers or by simply reducing the negative buoyancy so that finescale triggering mechanisms can more easily initiate convection. Such triggering mechanisms are often related to antecedent convection, and include density-current outflows and bores (e.g., Carbone et al. 1990) or trapped gravity waves (Fovell et al. 2006).

Moist absolutely unstable layers (MAULs) may be created when a layer becomes saturated and has a lapse rate that is greater than moist adiabatic (Bryan and Fritsch 2000). By definition, such layers contain no convective inhibition (CIN) and, within the approximations of parcel theory, require no further lifting mechanism for CI to ensue. MAULs have only relatively recently gained attention, perhaps because they are not often observed outside of deep and intense lifting along MCS outflows (e.g., Bryan and Fritsch 2000) or near hurricane eyewalls (e.g., Barnes 2008). Bryan and Fritsch argue that this is because such lifting needs to generate instability faster than it can be removed by convective overturning as a result of buoyancy. However, there may be situations where persistent, but weaker, vertical motions may contribute to the development of MAULs. For example, Schumacher and Johnson (2009) show a MAUL in a composite sounding of extreme rainfall environments (see their Fig. 14) where mesoscale convective vortices contribute to gradual mesoscale ascent.

Though recent model-based studies have quantified possible mechanisms contributing to nocturnal elevated CI, the relevant physical processes suggested by these modeling studies have not been widely confirmed using field observations. In the current study we present evidence obtained from analysis of PECAN radiosonde data to suggest that, unlike for PBL-based daytime CI, a significant fraction of nocturnal CI over the central United States could be associated with deep elevated MAULs or nearly saturated layers. We further document that the production of such layers coincides with mesoscale vertical motions in the preconvective environment.

This paper is organized as follows. In section 2 we overview the nocturnal CI and diverse large-scale environments in which it occurs for five different cases during PECAN. We discuss the methodologies used for the diagnosis of mesoscale vertical motions and results from this analysis in section 3. Section 4 examines the influence of the diagnosed mesoscale vertical motions on the evolution of the environmental thermodynamic vertical structure and how this can support CI.

2. Overview of convection initiation cases

Our focus is on CI that occurs within the core PECAN domain (Geerts et al. 2017) indicated by the dashed rectangles in Fig. 1. The synoptic conditions for the five CI cases examined in this paper fit broadly into three different categories: 1) strong synoptic forcing associated with a cold front and a midtropospheric short wave (Fig. 1b), 2) interaction of a nocturnal LLJ with a quasi-stationary lower-tropospheric front (Figs. 1a,d,e), and 3) a nocturnal LLJ located immediately downstream from a midtropospheric ridge axis in the absence of a well-defined surface front (Fig. 1c).

a. Data

Mesoscale horizontal convergence and vertical velocity are estimated (section 3a) for each case using approximately simultaneous PECAN radiosonde observations from three different locations. These locations compose triangles whose positions relative to both ongoing convection and CI are shown in Figs. 2–6. The radiosonde data are obtained primarily from the fixed profiling (FP) sites (Clark 2016; Holdridge and Turner 2015; Vermeesch 2015; UCAR/NCAR—Earth Observing Laboratory 2015, 2016a,b) within the core PECAN domain (Geerts et al. 2017, their Fig. 2) but, in some cases, are supplemented by mobile
profiling (MP) launches (Klein et al. 2016; Wagner et al. 2016; UCAR/NCAR—Earth Observing Laboratory 2016c) and mobile GPS (MG) launches (Ziegler et al. 2016) from vehicles dispatched during PECAN IOPs (Geerts et al. 2017). Surface temperature analyses in Figs. 2–6 are constructed using the General Meteorological Package (GEMPAK) utility (Koch et al. 1983) with a Barnes objective analysis chosen to emphasize large mesoscale ($L \geq 100$ km) features including frontal zones. Sharper boundaries including the leading edge of synoptic fronts and convectively induced outflows are, thus, not well resolved.

Our case selection criteria require there be no significant deep convection occurring within the radiosonde triangle at the time of the observations, but that CI (defined by the first appearance of radar reflectivity $\geq 40$ dBZ)
subsequently occurs somewhere inside the triangle within 30–150 min of the most recent observations. In addition, we require that the CI occurs at least 50 km from any ongoing convection and that it later attains mesoscale (e.g., $L \geq 100$ km) dimensions.

Since for mesoscale vertical motions it is the vertical displacements occurring over several hours that influence changes to thermodynamic structure favoring CI, successive horizontal divergence and vertical motion
calculations separated by 3 h are performed at the same triangle location in cases 2 (Figs. 3a,b) and 3 (Figs. 4a,b). This facilitated the analysis of the effect of the vertical motions on sounding evolution (section 4). Because of a
In addition to the requirements for triangle location relative to CI, the scale dependence of horizontal divergence and vertical velocity necessitated the establishment of triangle size and shape criteria. These criteria are loosely based on those of Trier and Davis (2007), who examined vertical motions in large mesoscale convective vortices during the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX; Davis et al. 2004). However, in the current study available observations resulted in our triangles typically being larger, ranging from 10,969 to 27,047 km² and having less variation in their obliqueness, with the smallest angles ranging from 33° to 52°.

b. Cases

1) STRONG COLD FRONT AND MIDLEVEL SHORT WAVE

For case 2 (IOP 16) vertical motion is calculated at both 0000 and 0300 UTC 26 June using a triangle composed of fixed profiling sites located at Ellis, Kansas (FP3); Brewster, Kansas (FP5); and Minden, Nebraska (FP4), which are collectively situated within a strong surface baroclinic zone (Figs. 3a,b) and between multiple ongoing convective systems. The convective systems included an MCS (system A) (Figs. 3a,b), which developed during the daytime north of the core PECAN domain and moved southeastward with the second of two midtropospheric short waves indicated in Fig. 1b. Farther south, a southwest (SW)–northeast (NE)-oriented zone of convection (system B) formed during the late afternoon and early evening near the leading edge of the surface cold front (Figs. 3a,b). Two adjacent intense lines of convection (system C) formed during the evening east of the triangle prior to the CI of interest (Fig. 3c), which initiated near the eastern portion of the triangle around 0410 UTC. The CI of interest rapidly became a ~100-km-long intense line with large maximum radar reflectivity values >50 dBZ along the eastern edge of the triangle (Fig. 3c) before dissipating during the next 2 h (not shown). The coexistence of multiple organized convective systems during a large portion of the diurnal cycle is consistent with the strong synoptic forcing during this period (Fig. 1b).

2) INTERACTION OF THE NOCTURNAL LLJ WITH A QUASI-STATIONARY FRONT

Cases 1 (Fig. 2), 4 (Fig. 5), and 5 (Fig. 6) also featured CI occurring within a north (N)–south (S)-oriented lower-tropospheric temperature gradient. However, in each of these cases the N–S temperature gradient is weaker than for case 2 (Fig. 3) and the CI is associated with the interaction of the nocturnal LLJ with the quasi-stationary surface front (Figs. 1a,d,e).
In case 1 (IOP 5) CI occurs at 0330 UTC 6 June ahead of a preexisting MCS propagating eastward along the cold (north) side of the surface frontal zone (Figs. 2a,b). The coverage of convection increases dramatically over the next 2 h and redefines the forward flank of the MCS (Fig. 2c). For this case, vertical motion is calculated 30 min prior to the first CI for a triangle (Fig. 2a) composed of fixed profiling sites located at Minden (FP4) and Hesston, Kansas (FP6), and a mobile profiling site MP3 operated by the University of Wisconsin–Madison.

In case 4 (IOP 23) CI also occurs north of the leading edge of the surface baroclinic zone and ahead of a preexisting MCS (Fig. 5). Here, the new convection organizes as a band of cells approximately perpendicular to the main MCS (Fig. 5c) and the surface temperature gradient (Fig. 5a). However, unlike in case 1 (Figs. 2b,c), this new region of convection does not evolve to become the dominant component of the MCS (not shown). Horizontal divergence and vertical motion in this case are calculated over a triangle composed of the fixed profiling sites at Lamont, Oklahoma (FP1); Greensburg, Kansas (FP2); and Hesston (FP6), at 0600 UTC 10 July (Fig. 5b) approximately 60 min prior to CI.

In case 5 (IOP 31) CI occurs in the late evening and overnight within the mesoscale region between multiple small- to medium-sized MCSs (Fig. 6). By 0600 UTC [0000 local standard time (LST)] 16 July, the dominant convective organization comprises a long approximate east (E)–west (W)-oriented band, which is situated within the surface baroclinic zone along the Kansas–Nebraska border region (Fig. 6c). Horizontal divergence and vertical motions are calculated at 0200 UTC over the triangle composed of the fixed profiling sites located in Brewster (FP5) and Minden (FP4), and the mobile GPS site MG2 operated by North Carolina State University (Fig. 6a). New convection emanates from multiple locations over the next 2 h (Fig. 6b) and, ultimately, merges within the triangle to compose the western part of the much longer E–W-oriented band by 0600 UTC (Fig. 6c).

3) NO SURFACE FRONT

CI in case 3 (IOP 19) occurs in a relatively warm region (Fig. 4) southwest of a cooler air mass centered in the upper Midwest. The lack of a clear association of this CI with a surface front distinguishes this case from the other four cases. This aspect of the current case was previously noted by Reif and Bluestein (2017), who also documented that approximately 25% of central Great Plains warm-season nocturnal CI occurs in the absence of well-defined surface boundaries. Horizontal divergence and vertical motions are calculated at both 0300 and 0600 UTC 5 July for a triangle composed of fixed profiling sites (Figs. 4a,b) located at Brewster (FP5), Ellis (FP3), and Greensburg (FP2). The southern end of the CI of interest is located within this triangle at 0700 UTC (Fig. 4c), 60 min after the end of the 3-h period during which horizontal divergence and vertical motion are estimated. According to Reif and Bluestein (2017), this relatively late occurrence of CI is characteristic of convective systems that initiate away from surface boundaries.

3. Mesoscale vertical motions

a. Analysis methods

Pressure and horizontal wind data from the PECAN radiosondes1 that compose the vertices of the triangles are vertically interpolated to isobaric surfaces separated by 5 hPa. Mesoscale horizontal divergence,

\[
\mathbf{\nabla} \cdot \mathbf{V} = \frac{1}{A} \frac{DA}{Dt},
\]

valid at the centroids of these triangles is then approximated on each of these isobaric surfaces using the approach of Bellamy (1949), where the instantaneous rate of change in area \(A\) of the triangle is estimated using the horizontal winds at the triangle vertices. In this procedure, the triangle is divided into two nonoverlapping right triangles, each containing two vertices comprising radiosonde wind observations. The total area of the triangle \(A\) is obtained from \(\sum_{i=1}^{2} A_i = 0.5 \sum_{i=1}^{2} b_i h_i\), where \(b_i\) and \(h_i\) are the base and height dimensions of the right triangles, respectively. The locations of the triangle vertices are then perturbed on each interpolated pressure level using the instantaneous wind observations acting over an arbitrary time interval \(\Delta t\). The resulting triangle is subdivided into two right triangles as before and the total area is recalculated and compared to the original triangle area to obtain \(\Delta A\) during \(\Delta t\), as well as the horizontal divergence estimate from (1).

Vertical profiles of the pressure vertical velocity, \(\omega = Dp/Dt \sim -\rho gw\), are obtained using mass conservation:

\[
\omega(p_2) = \omega(p_1) - \int_{p_1}^{p_2} \nabla \cdot \mathbf{V} \, dp,
\]

where \(p_2\) is the highest common pressure of the interpolated radiosonde data for the three locations that constitute the triangle vertices and \(p_1\) is the pressure at the approximate level of the tropopause. This pressure is

1 The soundings used in this study are from Vaisala model RS92- SGP and RS41 radiosondes, which had either 1- or 2-s data transmissions during their vertical ascent. Additional information may be found online (http://catalog.eol.ucar.edu/pecan).
taken as either 150 hPa or the lowest common interpolated pressure among the three locations. In (2) we use a kinematic lower boundary condition,

\[ \omega(p_1) = -\bar{\rho} g (\nabla H \cdot \nabla Z_0), \]

where \( \nabla H \), \( \nabla Z_0 \), and \( \bar{\rho} \) are the average surface wind, terrain height gradient, and density within the triangle, to estimate the effects of terrain-following flow at and below \( p_1 \) on vertical motion at \( p_1 \).

Though there is uncertainty in this estimate, the diagnosed \( \omega(p_1) \) values of \(-0.5 \) to \(-1.3 \mu_b \cdot s^{-1} \), which reflect upslope flow, are relatively small. However, much larger errors in \( \omega \) can accumulate when there are systematic horizontal divergence errors, as (2) is vertically integrated. In most cases, large magnitudes of mesoscale \( \omega \) at or above the tropopause imply excessively large surface pressure tendencies. Thus, triangles where \( |\omega(p_2)| \geq 10 \mu_b \cdot s^{-1} \) are judged unreliable and discarded. However, in other cases, an adjustment is applied to the horizontal divergence (e.g., Fankhauser 1969; O’Brien 1970), which ensures no net (i.e., total vertically integrated) horizontal divergence in the vertical column. The amount of horizontal divergence subtracted from the raw (i.e., original) divergence at each of the \( N \) interpolated pressure levels is given by

\[ 2D(j - 1) \]

\[ N(N - 1) \]

where \( D \) is the vertically integrated horizontal divergence at the top of the column and \( j \) is the index of the pressure level, with \( j = 1 \) at the bottom of the column and \( j = N \) at the top. In summary, this adjustment to the horizontal divergence is largest (2D/N) at the top of column \( p = p_2 \) and decreases linearly to zero at the bottom \( p = p_1 \). A vertical profile of adjusted \( \omega \) is then obtained from (2) using the adjusted horizontal divergence.

b. Results

Figure 7 presents both the original and adjusted vertical profiles of horizontal divergence and \( \omega \) within the triangles for the five cases discussed in section 2b (Figs. 2–6). Adjustments to the horizontal divergence are evident from the large differences between the raw and adjusted \( \omega \) at the top of the profiles in some of the cases (Fig. 7), which are likely indicative of systematic errors in the raw horizontal divergence. Both wind estimate errors and horizontal displacements of the radiosondes from the position of the launches can contribute to such errors. The wind estimate errors are likely to be a more substantial contributor to \( \omega \) errors near \( p = p_2 \) than elsewhere because of the upward vertical integrations used to diagnose \( \omega \) from (2). Moreover, horizontal drift of the radiosondes ranges from about 0 to 13 km at 500 hPa, which is only a small fraction of the triangle leg lengths (typically <5%) and much less than at 150 hPa (Table 1). Thus, we anticipate the \( \omega \) vertical profiles in Fig. 7 to be much more accurate in the low to midtroposphere (i.e., beneath 500 hPa) and we restrict our analysis to these layers.

Each profile has a 150–300-hPa-deep layer of horizontal convergence in the lower to middle troposphere, with upward motion maximized at the top of these layers (Fig. 7). The vertical profiles for cases 1 (Fig. 7a), 4 (Fig. 7d), and 5 (Fig. 7e), which correspond to pre-CI conditions within and north of a quasi-stationary surface frontal zone [section 2b(2)], are each diagnosed from measurements occurring at a single time and have similar maximum mesoscale upward motion of \(-4 \) to \(-5 \mu_b \cdot s^{-1} \) located from 750 to 800 hPa. Such vertical velocities acting for several hours would result in upward vertical displacements of about 50 hPa (\(-0.5 \) km), which can substantially moisten these levels (section 4b). In the case 4 profiles, which are diagnosed from measurements several hours later than in cases 1 and 5, there is a shallow layer of horizontal divergence near the surface (Fig. 7d).

The vertical profiles for cases 2 (Fig. 7b) and 3 (Fig. 7c) are each averages from two nominal times separated by \(-3 \) h. In case 3, which corresponds to pre-CI conditions in the absence of a front or significant surface boundary [section 2b(3)], there is also a layer of near-surface divergence (Fig. 7c). Similar to case 4, the CI in this case occurred after 0600 UTC, which is later in the diurnal cycle than for the other three cases. The deepest horizontal convergence and strongest mesoscale vertical motion occur in case 2 (Fig. 7b), which corresponds to the environment of a strong cold front and midtropospheric short wave [Fig. 1b; section 2b(1)]. In this case, while strongest at 0300 UTC (Fig. 7b) approximately 1 h prior to CI in the triangle (Fig. 3c), the shape and amplitude of the average low to midtropospheric vertical motion profile suggest similar conditions occurring for several hours before.

c. Comparison of results with short-range ensemble forecasts

Apart from wind estimate errors and errors related to radiosonde drift, there are possible representativeness issues related to diagnosing mesoscale vertical motions from triangles, since information from only three locations at each pressure level is used. To assess representativeness, adjusted \( \omega \) profiles in Fig. 7 are compared with corresponding area-averaged \( \omega \) calculated over the same mesoscale triangles from an ensemble of 0–6-h forecasts.

For each case, a 50-member ensemble forecast is produced using initial conditions (ICs) provided by a
continuously cycled mesoscale, ensemble adjustment Kalman filter data assimilation system (Anderson 2001, 2003), as described in Schwartz et al. (2015). This forecast system assimilates a variety of mesoscale and synoptic observations\(^2\) with the Data Assimilation Research Testbed (DART; Anderson et al. 2009) software to produce a 50-member ensemble of mesoscale analyses with 15-km horizontal grid spacing every 6 h. Using these ICs, ensemble forecasts are integrated from 0000 to 0600 UTC using version 3.6.1 of the Advanced Research core of the Weather Research and Forecasting (WRF) Model (ARW; Skamarock and Klemp 2008). Boundary conditions are provided by Global Forecast System (GFS) analyses. The forecast domain includes the entire CONUS using 15-km horizontal grid spacing (415 × 325 grid points) and 39 vertical levels. Cumulus and microphysical parameterizations are provided by the Tiedtke cumulus scheme (Tiedtke 1989; Zhang et al. 2011) and the Thompson et al. (2008) bulk microphysics scheme, respectively. Other parameterizations include the Mellor–Yamada–Janjic’ (MYJ; Mellor and Yamada 1982; Janjić 1994, 2002) PBL scheme, the Rapid Radiative Transfer

\(^2\)The data assimilation does not include field observations from PECAN.
Table 1. Launch times and sizes for radiosonde triangles, and the horizontal displacement of their vertices at different pressure levels for triangles shown in Figs. 2–6, which are used to calculate horizontal divergence and kinematic vertical motions in Fig. 7. The letter “M” indicates missing data due to radiosonde termination prior to reaching the indicated pressure level (150 hPa).

<table>
<thead>
<tr>
<th>Triangle time and date</th>
<th>Triangle leg lengths (km)</th>
<th>Vertex x drifts (km) at 500 hPa</th>
<th>Vertex y drifts (km) at 500 hPa</th>
<th>Vertex x drifts (km) at 150 hPa</th>
<th>Vertex y drifts (km) at 150 hPa</th>
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</table>


Figure 8 compares vertical profiles of adjusted ω diagnosed from observations with corresponding area-averaged ω from the model ensemble. Since an important criterion used for triangle selection is the absence of significant precipitation within the triangle at the time of the radiosonde observations, ensemble members having concurrent area-averaged rain rates exceeding 0.5 mm h⁻¹ in the triangles are not included in the comparison. This eliminates 0–10 of the 50 ensemble members, depending on the case (Fig. 8). There is no explicit smoothing of ω in either the ensemble members or the observations. However, the averaging of O(100) grid points over the mesoscale triangle area in the model results in damping of small scales that could still be retained in radiosonde measurements used to diagnose ω.

Though there is substantial variation in the amplitude of the simulated ω among ensemble members in individual cases (Fig. 8), low- to midtropospheric ω minima occur for both the ensemble (gray shading) and observations (red curves). The observations have greater amplitude than the profile composed of the ensemble median at each level (black curves), and are more strongly peaked in the vertical, which is not surprising considering the substantial variations in the heights of vertical motion extrema expected in an ensemble.

The biggest disagreements between much of the ensemble and observations are for case 2 (Fig. 8b). In this case, both the timing of a synoptic cold front (Fig. 1b) and possible influences from convection located close to the observations taken from the southern vertex (FP3) of the triangle at 0300 UTC (Fig. 3b) are factors that could affect this comparison. In case 3 (Fig. 8c), the ω minimum typically occurs at lower altitudes in the ensemble members than in the observations. However, for all five cases there are individual members (blue curves) whose vertical profiles of ω are similar to the observed ones (red curves).

4. Influences of mesoscale vertical motions on nocturnal convection initiation

Ensemble members with ω profiles similar to observations constitute temporally continuous datasets, which provide spatial information in individual cases that is not available from field observations alone. In this section we use ensemble members with the smallest vertically averaged (p₁ to 450 hPa) root-mean-square (RMS) error of ω (Fig. 8, blue curves) compared with observations, together with these PECAN radiosonde observations, to illustrate how these mesoscale vertical motions can influence the thermodynamic environment for the different situations (section 2) in which elevated nocturnal CI subsequently occurs.

a. Cold front and midlevel short wave (case 2)

The effects of lifting in case 2 are evident in the evolution of the FP3 soundings (Fig. 9) located at the south end of the radiosonde triangle within which CI subsequently occurs (Fig. 3). In particular, an elevated approximately dry-adiabatic layer located between 830 and 775 hPa at 0000 UTC 26 June deepens and becomes situated between 800 and 725 hPa by 0300 UTC (Fig. 9). The mean mesoscale ascent estimated by the average of the 0000 and 0300 UTC adjusted ω profiles (Fig. 7b) is 5 μbs⁻¹ at 815 hPa, which is the average pressure of the base of this layer. The mean ascent at 750 hPa, which is the average pressure at the top of this layer, is 8 μbs⁻¹. The resulting vertical displacements from these 3-h average vertical motions are 54 and 86 hPa, respectively, which are qualitatively consistent with but greater than the corresponding 30- and 50-hPa upward displacements of the bottom and top of this dry-adiabatic layer at FP3 (Fig. 9).

These quantitative differences may be influenced by horizontal ω gradients located between FP3 and the FP3–FP4–FP5 triangle center, where the ω diagnosis is
valid. This assertion is supported by the 750-hPa $\omega$ from the smallest-RMS $\omega$ ensemble member 46 (Fig. 10a), which indicates weaker vertical motion at the southeastern vertex (FP3) of the triangle than near its center. The strong horizontal $\omega$ gradients and their variability among ensemble members also appear to influence the relatively small magnitudes of the median values of simulated $\omega$ in Fig. 8b. The band of strong mesoscale upward motion in member 46 near the triangle is collocated with a region of reduced magnitudes of maximum negative buoyancy ($B_{\text{min}}$; Trier et al. 2014a,b) of the 50-hPa averaged highest-$\theta_e$ parcel (Fig. 10b), which favors CI. Here, $B_{\text{min}}$ (in units of virtual temperature) is the largest virtual temperature deficit [$T_{\text{v(parcel)}} - T_{\text{v(sounding)}}$] a pseudoadiabatically displaced parcel attains beneath its level of free convection (LFC), or below 500 hPa if no LFC occurs. The convective available potential energy of the most unstable parcels (MUCAPE) is large and these parcels are elevated (0.4–1.6 km AGL) within the triangle at 0300 UTC (Fig. 10c), ~1 h before observed CI (cf. Fig. 3c).

Substantial local cooling and moistening between 775 and 725 hPa at FP3 from 0000 to 0300 UTC coincides with the vertical displacement of the elevated dry-adiabatic layer (Fig. 9) making conditions more
favorable for elevated CI by 0300 UTC. This contrasts with the strong cooling and drying below 825 hPa associated with horizontal temperature and moisture advection behind the surface cold front (Fig. 9), which led to large CIN of \(-187 \text{ J kg}^{-1}\) for surface-based parcels at FP3. At 0300 UTC the most unstable 50-hPa-deep parcel for FP3 is located from 790 to 740 hPa (Table 2) and has CAPE of 2410 J kg\(^{-1}\). Despite improving conditions for elevated CI, there remains CIN of \(-24.1 \text{ J kg}^{-1}\) at 0300 UTC, necessitating finescale vertical velocities of \(w = \sqrt{2 \text{CIN}} \approx 7 \text{ m s}^{-1}\) for CI to ensue from the 790–740-hPa layer.

**b. Interaction of LLJ with a stationary front near ongoing convection (cases 1, 4, and 5)**

Common attributes of cases 1, 4, and 5 are the interaction of a nocturnal LLJ and a quasi-stationary surface front, with CI occurring near preexisting MCSs. One difference between these cases and the previously discussed case (case 2; section 4a) is evidence of elevated nearly saturated layers having lapse rates greater than moist adiabatic (red curves in Fig. 11). Water vapor saturation results in shallower 30–50-hPa-deep MAULs embedded within these deeper elevated conditionally unstable layers. The MAULs are based at 780 hPa for case 1 at 0259 UTC 6 June in FP4 (Fig. 11a), and at 830 hPa for case 4 at 0600 UTC 10 July in MP4 (Fig. 11b). By definition MAULs have negligible CIN and \(B_{\text{min}}\) and, in these particular cases, significant CAPE values of 1800 J kg\(^{-1}\) (case 1) and 1570 J kg\(^{-1}\) (case 4) for 50-hPa-deep averaged air parcels beginning at their base (Table 2). Since MAULs are unlikely to persist for long outside of deep convection (Bryan and Fritsch 2000), their presence here may perhaps signify the earliest stages of CI (cf. Figs. 2b and 5c), which could be slowed by dry, stable layers beginning \(\sim 1 \text{ km}\) above at 670 hPa in case 1 (Fig. 11a) and 700 hPa in case 4 (Fig. 11b).

In case 4 there is strong horizontal flow deformation near an 850-hPa temperature gradient (Fig. 12a). Augustine and Caracena (1994) discuss a similar frontogenesis lower-tropospheric pattern, with the axis of contraction parallel to \(\nabla_H T\), in precursor environments for strong nocturnal MCSs. Frontogenesis is also common in more general elevated thunderstorm environments (Colman 1990; Moore et al. 2003). Mesoscale vertical motion profiles from triangles in case 4 (Fig. 12a) are consistent with the vertical component of a transverse frontal circulation pattern with upward (red) and downward (cyan) branches separated by \(\sim 200 \text{ km}\) (Fig. 12b).

The frontogenetic effect of the horizontal wind field is explicitly illustrated with the smallest-RMS \(\omega\) ensemble member for case 4, which has similar 850-hPa flow and temperature patterns (Fig. 13a) to those observed (Fig. 12a) near the FP1–FP2–FP6 triangle. The change in the magnitude of the horizontal temperature gradient following the flow that arises from the horizontal flow acting on the temperature gradient is

\[
F = \frac{D}{Dt} [\nabla_H T] = \frac{[\nabla_H T]}{2} \text{ [DEF cos(2}\beta - \text{DIV}]},
\]

where

\[
\text{DEF} = \left[\left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}\right)^2 + \left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}\right)\right]^{0.5}
\]

and

\[
\text{DIV} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}
\]

are the total horizontal deformation and divergence, respectively, and \(\beta\) is the angle between the isotherms and the axis of dilatation.

The two components of \(F\) on the rhs of (4) are added, and their sum is smoothed using a sinusoidal weighting function emphasizing mesoscale horizontal wavelengths of \(L \approx 75 \text{ km}\) in Fig. 13a. Frontogenesis occurs in the 850-hPa frontal zone with \(F \approx 0.3–0.4 \text{ K (100 km)}^{-1} \text{ h}^{-1}\) near the FP1–FP2–FP6 triangle (Fig. 13a), where mesoscale ascent is both simulated and observed (Fig. 8d, cf. blue and red curves). In the simulation, there is negligible negative buoyancy for the most unstable parcel (Fig. 13a) located at \(\sim 1 \text{ km AGL}\) (Fig. 13b) within and near this triangle. The 0600 UTC MP4 sounding located \(\sim 50 \text{ km}\) southwest of the triangle centroid (Fig. 13) also has its most unstable parcel at \(\sim 1 \text{ km AGL}, \sim 200 \text{ m}\) beneath the base of the MAUL (Fig. 11b).

In case 5, there is a similar cooling and moistening above the surface in the FP5 soundings (Fig. 11c) with
diagnosed low- to midtropospheric mesoscale ascent in the adjacent triangle (Fig. 7e) prior to CI (Fig. 6). These soundings indicate that the midtropospheric cooling and moistening occur gradually for the 2 h prior to CI.

The influences of mesoscale vertical motion on the development of the elevated moist conditionally unstable layers that support CI are estimated using triangle values and thermodynamic evolution in soundings from Fig. 11. The adiabatic equations governing local potential temperature and water vapor changes are

\[
\frac{\partial \theta}{\partial t} = -\mathbf{V} \cdot \nabla \theta - \omega \frac{\partial \theta}{\partial p} \tag{5}
\]

and

\[
\frac{\partial q_y}{\partial t} = -\mathbf{V} \cdot \nabla q_y - \omega \frac{\partial q_y}{\partial p}, \tag{6}
\]

respectively. Assuming that the \( \omega \) profiles diagnosed from the mesoscale triangles are both representative of the vertical motions for the \( \Delta t = 2 \) h period, and uniform near and within the triangle, the related temperature and water vapor changes at the sounding locations are approximated by \(-\mathbf{\omega} \cdot \partial \theta / \partial p \Delta t\) and \(-\mathbf{\omega} \cdot \partial q_y / \partial p \Delta t\), respectively, where the overbars denote layer averages and brackets denote 2-h time means. Using \( \mathbf{\omega} = -2.2 \) \( \mu \)s\(^{-1}\) from the 800–750-hPa layer of the 0600 UTC 10 July FP1–FP2–FP6 triangle (Fig. 7d) and the \( \theta \) and \( q_y \) profiles from the 0400, 0500, and 0600 UTC MP4 soundings (Fig. 11b; see Fig. 13 for location) yields \( \Delta \theta = -0.7 \) K and \( \Delta q_y = +1.1 \) g kg\(^{-1}\), which is most of the potential temperature (\( \Delta \theta = -0.9 \) K) and more than half the actual moisture (\( \Delta q_y = +2.0 \) g kg\(^{-1}\)) changes, respectively. The 0.9 g kg\(^{-1}\) departure from the observed moisture change suggests that horizontal moisture advection may be augmenting the vertical moisture advection in influencing the 2-h \( \Delta q_y \) in case 4 (Fig. 11b). For case 5, using \( \mathbf{\omega} = -4.4 \) \( \mu \)s\(^{-1}\) from the 720–670-hPa layer of the 0200 UTC 16 July MG2–FP5–FP4 triangle (Fig. 7e), the \( \theta \) and \( q_y \) profiles from the 0200, 0300, and 0400 UTC
FP5 soundings (Fig. 11c), yields $\Delta T = -1.2$ K and $\Delta q = +2.3$ g kg$^{-1}$, which represent most of the actual potential temperature change ($\Delta T = -1.6$ K), and all of the actual moisture ($\Delta q = +2.2$ g kg$^{-1}$) change, respectively.

Unlike for cases 4 and 5, the development of the deep approximately saturated conditionally unstable layer at FP4 (Fig. 11a) situated near CI in case 1 (Fig. 2b) is not dominated by local cooling and moistening from upward vertical transports. For case 1, using $\omega = -2.9 \mu b$ s$^{-1}$ for the 800–710-hPa layer of the 0300 UTC 6 June MP3–FP4–FP5 triangle (Fig. 7a), along with the 0000 and 0300 UTC FP4 soundings, yields $\Delta T = -1.5$ K and $\Delta q = 0.8$ g kg$^{-1}$ from the effect of vertical motions alone. However, the actual changes during this 3-h period are $\Delta T = -0.1$ K and $\Delta q = 3.8$ g kg$^{-1}$ (Fig. 11a), which differ substantially from these estimates.

Here in case 1, horizontal advections [first terms on the rhs of (5) and (6)] likely contribute much more significantly to the local thermodynamic evolution (Fig. 11a) than they do in cases 4 (Fig. 11b) and 5 (Fig. 11c). The horizontal advection is more difficult to estimate than mesoscale vertical motions using the PECAN dataset. However, if we assume geostrophic flow, the layer-averaged horizontal temperature advection may be approximated from a single sounding using

$$\nabla \cdot \mathbf{V} \frac{T}{p} = \frac{f}{R \ln(p_b/p_t)} |\mathbf{V}_{gb}||\mathbf{V}_{gs}| \sin \alpha, \tag{7}
$$

where $|\mathbf{V}_{gb}|$ and $|\mathbf{V}_{gs}|$ are the geostrophic wind magnitudes at the bottom and top of a thin layer defined by pressures $p_b$ and $p_t$, and $\alpha$ is the angle between $\mathbf{V}_{gb}$ and $\mathbf{V}_{gs}$. Because of the large magnitude of mid- to lower-tropospheric divergence (Fig. 7a), (7) is expected to provide only a rough estimate of the actual horizontal temperature advection. Nevertheless, assuming the actual horizontal winds at 0000 and 0300 UTC from FP4 are geostrophic, substitution into (7) and using $\bar{v} = \bar{T}(1000/p)^{10/2}$, with $p = (p_b + p_t)/2$, yields a 800–710-hPa 3-h change of

![Fig. 11. As in Fig. 9, but for (a) the fixed profiling site FP4 (location shown in Fig. 2) in case 1, (b) the mobile profiling site MP4 (location shown in Fig. 5) in case 4, and (c) the fixed profiling site FP5 (location shown in Fig. 6) in case 5.](image)

### Table 2: Thermodynamic parameters for 50-hPa-deep averaged air parcels based at the listed pressure levels from PECAN CI proximity soundings illustrated by the red profiles in Figs. 9, 11, and 15.

<table>
<thead>
<tr>
<th>50-hPa avg parcel levels (hPa)</th>
<th>CAPE (J kg$^{-1}$)</th>
<th>Lifted index (°C)</th>
<th>CIN (J kg$^{-1}$)</th>
<th>$B_{min}$ (°C)</th>
<th>LFC (hPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 1 (6 Jun) FP4 at 0300 UTC</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>941 (surface)</td>
<td>870</td>
<td>-4.5</td>
<td>-131</td>
<td>-4.2</td>
<td>675</td>
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<tr>
<td>780</td>
<td>1800</td>
<td>-6.6</td>
<td>-2</td>
<td>-0.5</td>
<td>750</td>
</tr>
<tr>
<td>Case 2 (26 Jun) FP3 at 0300 UTC</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>939 (surface)</td>
<td>1960</td>
<td>-7.9</td>
<td>-187</td>
<td>-4.8</td>
<td>650</td>
</tr>
<tr>
<td>790</td>
<td>2410</td>
<td>-8.9</td>
<td>-24</td>
<td>-1.6</td>
<td>670</td>
</tr>
<tr>
<td>Case 3 (5 Jul) FP3 at 0600 UTC</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>937 (surface)</td>
<td>1840</td>
<td>-6.6</td>
<td>-280</td>
<td>-5.3</td>
<td>650</td>
</tr>
<tr>
<td>875</td>
<td>1760</td>
<td>-6.4</td>
<td>-139</td>
<td>-3.7</td>
<td>650</td>
</tr>
<tr>
<td>800</td>
<td>1000</td>
<td>-4.8</td>
<td>-94</td>
<td>-3.1</td>
<td>630</td>
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<tr>
<td>Case 4 (10 Jul) MP4 at 0600 UTC</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>968 (surface)</td>
<td>1080</td>
<td>-4.6</td>
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<tr>
<td>830</td>
<td>1570</td>
<td>-5.5</td>
<td>0</td>
<td>-0.1</td>
<td>820</td>
</tr>
<tr>
<td>Case 5 (16 Jul) FP5 at 0400 UTC</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>896 (surface)</td>
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<td>-7.6</td>
<td>-78</td>
<td>-2.7</td>
<td>720</td>
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<tr>
<td>730</td>
<td>1080</td>
<td>-5.3</td>
<td>-1</td>
<td>-0.5</td>
<td>700</td>
</tr>
</tbody>
</table>
$
\Delta \theta = 1.7 \text{ K}$ resulting from horizontal advection. This contribution approximately cancels the $\Delta \theta = 1.5 \text{ K}$ estimated cooling from vertical motions, consistent with the small observed temperature change in the developing MAUL layer (Fig. 11a).

The $\Delta q = 3.8 \text{ g kg}^{-1}$ moisture increase at FP4 (Fig. 11a) results in an elevated maximum of equivalent potential temperature $\theta_e$ of 346 K located at 760 hPa within the MAUL. Since $\theta_e$ is approximately conserved outside of regions of heating and vigorous mixing, a vertical cross section taken through a broad zone of 800-hPa warm advection in the lowest RMS ensemble member for case 1 (Fig. 14a) suggests the moisture source for the MAUL is in the lower troposphere, closer to the leading edge of the front (Fig. 14b). The rearward-sloping $\theta_e$ maximum is consistent with the effects of both vertical...
moisture advection and differential horizontal moisture advections in the plane of the cross section downstream from (and above) the core of the nocturnal LLJ (Fig. 14c). This results in the most unstable air parcels with small negative buoyancy being located near FP4 by late evening, more than 1 km above the surface (Fig. 14a), which is consistent with the local evolution in the soundings (Fig. 11a).

c. No surface front (case 3)

As noted in section 2b(3), the lack of a well-defined surface front in case 3 (5 July) is an aspect that distinguishes it from the CI in the previously examined four cases. The CI of interest occurs late at night as a north–south-oriented band (Fig. 4c) in relatively warm surface conditions. PECAN soundings from FP3, located east of the southern edge of subsequent CI (Fig. 4), indicate a general cooling and moistening in midlevels from 650 to 525 hPa (Fig. 15) prior to this CI. This thermodynamic evolution at FP3 is broadly consistent with the modest mesoscale ascent through a similar layer (Fig. 7c, solid red curve) estimated from averaging the 0300 and 0600 UTC FP2–FP5–FP3 radiosonde triangles (Fig. 4).

Several layers in the 0600 UTC FP3 sounding have significant CAPE, including 50-hPa-deep layers based at the surface, a layer starting at 875 hPa located above the nearly isothermal layer, and 800-hPa potential temperature (2-K contour interval) and horizontal winds. In (b) the equivalent potential temperature (solid blue lines with 2.5-K contour interval) and in (c) the horizontal winds along the plane of the cross section (solid blue lines with 3 m s\(^{-1}\) contour intervals, negative values dashed) are shown. The location of the sounding from the fixed profiling site FP4 (Fig. 11a) is indicated by the white cross symbol in (a).
favorable conditions for CI, the smallest-RMS $\omega$ ensemble member 10 (Fig. 8c, blue curve) has a 700-hPa moisture maximum located between FP3 and FP5 (Fig. 16a). A model sounding from this mixing ratio maximum has more favorable conditions consisting of a 100-hPa-deep approximately dry-adiabatic, but nearly saturated ($\text{RH} = 80\%-90\%$), layer based near 700 hPa at 0600 UTC (Fig. 16b).

Reif and Bluestein (2017) discuss the importance of warm advection for late-night CI that occurs in the absence of surface boundaries and illustrate warm advection during this particular case (their Fig. 21a). In our analysis, warm advection is evident beneath and upstream of the simulated moisture maximum and the observed CI location in a vertical cross section (Fig. 17b) along transect WE from Fig. 16a. Similar to conditions in the smallest-RMS $\omega$ member from the previously discussed case 1 (Fig. 14b), there is an upward-sloping region of maximum $\theta_e$ ($x = 150-400$ km in Fig. 17d) into the midlevel region of high relative humidity in the smallest-RMS $\omega$ member from the current case.

Back trajectories emanating from this region of high relative humidity at 680 hPa (Fig. 18b), 30–60 min prior to observed CI (Fig. 4c), are representative of the air that ends up composing the lower part of the elevated moist layer denoted by the red $T$ and $T_d$ curves in Fig. 16b. These trajectories originate 5 h earlier near the top of the late afternoon PBL (Fig. 18c, solid curve), where it is particularly deep and energetic with $\theta_e \approx 350$ K (Fig. 18a), approximately 150 km west-southwest of their final locations (Fig. 18b). Though $\theta_e$ decreases $\approx 2-2.5$ K along the 5-h trajectories, the stability above their final locations is smaller (Fig. 18d), thus facilitating CI.

Along the first half of the trajectories, there are no significant net decreases in pressure (Fig. 19a) and no significant relative humidity increases (Fig. 19b). However, CIN decreases during the first few hours of these trajectories (Fig. 19d), as the trajectories move toward a region where the overlying air is colder with $\Delta \theta_e \approx -2$ K (cf. dashed and solid curves in Fig. 18c). After 0300 UTC, each trajectory undergoes sustained pressure decreases of 25–40 hPa (Fig. 19a), indicating $\approx 300$–500-m upward displacements (not shown), as they move through the region of warm advection (Fig. 16a). These gradually ascending trajectories have relative humidity rises of 15%–30% and CIN approaches zero (Fig. 19d), which helps permit CI in the environment of moderate CAPE of $\approx 1000$ J kg$^{-1}$ (Fig. 19c).

A common aspect of case 1 and the current case 3 is the association of upward moisture transports with warm advection occurring in a baroclinic zone. Together, these processes contribute to a favorable environment for elevated CI in nearly saturated conditionally unstable layers.

In case 1, the baroclinic zone and associated warm advection are linked to the interaction of the LLJ and a quasi-stationary surface front. In contrast, the baroclinicity in the current case is weaker and more widespread, and the relevant warm advection occurs in the plane opposite LLJ orientation and 100–200 hPa above it (Figs. 17a,b).

5. Summary

This study uses radiosonde measurements collected during the Plains Elevated Convection At Night (PECAN) 2015 field campaign to diagnose mesoscale vertical motions and assess their role in nocturnal convection initiation (CI) over the central United States. Examination of physical processes contributing to nocturnal CI was
significant component of PECAN (Geerts et al. 2017). In
the current study our analysis is limited to cases where CI
occurred at least 50 km away from ongoing convection
and persisted for $\geq 1.5$ h while organizing upscale. We
additionally required that CI occur within the area de-
finite by radiosonde triangles meeting certain size and
structural criteria (section 2), and that radiosonde mea-
surements were available prior to the onset of CI in order
to evaluate mesoscale vertical motions in the preconvective
environment.

These requirements resulted in five qualifying CI
cases that are examined in this paper. However, these
five cases encompass several characteristic types of
large-scale environments in which nocturnal CI com-
monly occurs, including

- synoptic forcing with a strong cold front and midlevel
  short wave,
- interaction of the nocturnal low-level jet with a quasi-
  stationary surface front, and
- lower- to midtropospheric warm advection in the
  absence of surface boundaries.

Though there are variations among the five cases in
kinematically diagnosed mesoscale vertical motions
prior to CI, the five cases examined herein share the
common characteristic of maximum ascent in the lower
to middle troposphere. These diagnosed mesoscale
vertical motion profiles are broadly consistent with their
counterparts from ensembles of 0–6-h low-resolution
WRF Model forecasts, where averages of vertical mo-
tion are calculated over the same mesoscale regions as
the observations. This comparison provides confidence
in the representativeness of the mesoscale vertical motions
diagnosed from the radiosondes and also indicates
the usefulness of such ensembles for short-range fore-
casts of where mesoscale conditions are likely to be most
favorable for CI during both the daytime-to-nocturnal
transition and overnight.

In each of the three environmental categories noted
above, elevated convection is strongly supported by the
thermodynamic vertical structure diagnosed from either
the radiosondes or model soundings at times and loca-
tions close to the observed CI. Furthermore, diagnosed

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**FIG. 17.** Vertical cross sections of relative humidity (color shading) and potential temperature (dashed black lines
with 2-K contour interval) along transects (a),(c) SN and (b),(d) WE in Fig. 16a. Horizontal winds parallel to the
cross sections are depicted in (a) and (b) by the blue solid lines with the contour intervals of 2.5 m s$^{-1}$ in (a) starting
at 10 m s$^{-1}$ and the contour intervals of 2 m s$^{-1}$ in (b) starting at 2 m s$^{-1}$. Equivalent potential temperature is
depicted in (c) and (d) by the blue solid lines with contour intervals of 2.5 K starting at 340 K.
lower- to midtropospheric mesoscale ascent in the examined cases of $-3$ to $-10 \mu \text{b s}^{-1}$, when acting over several hours, result in upward vertical displacements ($\sim 0.3$–$1 \text{ km}$) that are broadly consistent with thermodynamic evolution in most of these CI proximity soundings.

The modifications to the thermodynamic vertical structure are found to support elevated CI in two different ways. In the case with the strongest synoptic forcing (26 June) vertical motions are strongest, and this mesoscale ascent reduces CIN to values where CI may have been more easily triggered by finescale circulations associated with nearby preexisting convection. However, in the majority of the examined cases mesoscale vertical motions contribute to an evolution from significantly subsaturated conditions to deep, elevated conditionally unstable layers with high relative humidities that support the observed CI. In two of these cases, moist absolute instability (Bryan and Fritsch 2000) is observed within portions of the elevated destabilizing layers.
The development of nearly saturated conditionally unstable layers or moist absolutely unstable layers (MAULs) can be well explained by the local cooling and moistening associated with diagnosed mesoscale vertical motions in some cases (10 and 16 July), if such mesoscale ascent were to locally persist for several hours. However, in other cases (6 June and 5 July) observations together with analyses of selected ensemble forecast members reveal that horizontal temperature and moisture transports are also important in the development of these layers. The development of such layers can occur under different environmental conditions, including situations where the nocturnal low-level jet interacts with surface frontal zones, but also in environments where clearly defined surface frontal zones or convective outflow boundaries are not present. A common aspect of these otherwise disparate environments is mesoscale ascent in the presence of warm advection, which allows the vertical transport of water vapor that promotes elevated MAULs or nearly saturated conditionally unstable layers.

Such layers can obviate the need for finescale (e.g., $L \sim 1–10 \text{ km}$) lifting and may explain some instances of nocturnal CI in the apparent absence of near-surface boundaries, which have been documented in earlier field experiments such as IHOP (e.g., Wilson and Roberts 2006). Further work that combines high-resolution observations and convection-allowing simulations is necessary to understand the progression from mesoscale ascent destabilizing the thermodynamic environment to the onset of CI in these different varieties of nocturnal CI types.

**Acknowledgments.** The authors acknowledge Chris Davis, George Bryan, Rita Roberts, and Morris Weisman (each of NCAR) for helpful discussion on aspects of this research, and Richard Rotunno and Tammy Weckwerth (each of NCAR) for providing helpful internal reviews of the manuscript. The paper also benefited from the comments and suggestions of John Peters (Naval Postgraduate School) and two anonymous reviewers. We also thank all participants of PECAN for their hard work during the field campaign. Special appreciation is extended to the Colorado State University, North Carolina State University, University of Oklahoma, University of Wisconsin–Madison, University of Alabama in Huntsville, and NCAR/EOL mobile sounding teams, and to David Turner (NOAA/National Severe Storms Laboratory) for overseeing the planning of the Fixed Profiling (FP) sounding sites for PECAN. These collective

![Thermodynamic Quantities along 0600-0100 UTC 5 July Back Trajectories (Case 3, Member 10)](image-url)

**Fig. 19.** (a) Pressure (hPa), (b) relative humidity (%), (c) CAPE (J kg$^{-1}$), and (d) CIN (J kg$^{-1}$) along the color-coded back trajectories plotted in Fig. 18b.
efforts resulted in an outstanding radiosonde dataset, without which the current research would not have been possible. The field data used in this study were obtained from NCAR/EOL (http://data.eol.ucar.edu) under the sponsorship of the National Science Foundation. The current research was performed as part of NCAR’s Short Term Explicit Prediction (STEP) program, which is supported by National Science Foundation funds for the U.S. Weather Research Program (USWRP). We also acknowledge high-performance computing support from Yellowstone (ark:/85065/d7w3xhc) provided by NCAR’s Computational and Information Systems Laboratory, sponsored by the National Science Foundation.

REFERENCES


Clark, R., 2016: FP3 Ellis, KS radiosonde data, version 2.0. UCAR/NCAR—Earth Observing Laboratory, accessed 1 October 2016, doi:10.5065/D6GMR5DZ.


Wagner, T., E. Olson, and W. Feltz, 2016: Mobile PISA 3 UW/SSEC SPARC radiosonde data, version 2.0. UCAR/NCAR—Earth

