Observations of Elevated Convection Initiation Leading to a Surface-Based Squall Line during 13 June IHOP_2002

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(Manuscript received 16 March 2010, in final form 6 August 2010)

ABSTRACT

The evolution of a mesoscale convective system (MCS) observed during the International H2O Project that took place on the Great Plains of the United States is described. The MCS formed at night in a frontal zone, with four initiation episodes occurring between approximately 0000 and 0400 local time. Radar, radiosonde, and surface data together show that at least three of the initiation episodes were elevated, occurring from moist conditionally unstable layers located above the boundary layer, which had been stabilized by previous MCSs. Initiation occurred in northwest–southeast-oriented lines where a southerly nocturnal low-level jet terminated, generating elevated convergence. One initiation episode was observed using the S-band dual-polarization Doppler radar (S-Pol) and occurred at the intersection of this convergence zone with a propagating wave. Calculations of the Scorer parameter were consistent with wave trapping. Downdrafts from the developing convection generated both waves and bores, which propagated ahead of the cold pool, initiating further convection. Between 0700 and 1000 local time, the structure and orientation of the MCS evolved to a southwest–northeast-oriented squall line, which built a cold-pool outflow that could lift near-surface air to its level of free convection. The weaker cold pool in the eastern part of the domain was consistent with the greater impacts of a previous MCS there. To the authors’ knowledge, this case study provides the first detailed observational investigation of elevated initiation leading to surface-based convection, a process that appears to be an important mechanism for the generation of long-lived MCSs from elevated initiation.

1. Introduction

Nocturnal convective storms in the United States are poorly forecast compared with convection during the day (Davis et al. 2003). One aspect of nocturnal warm-season precipitation that contributes to the difficulty of its prediction is the greater occurrence of elevated convection at night (e.g., Wilson and Roberts 2006). Here, we refer to convection where the conditionally unstable source air is located above the boundary layer as “elevated” (Glickman 2000) and storms fed from the boundary layer as “surface based,” although in reality there is a continuum between these two extremes (Corfidi et al. 2008). Elevated convection can occur above a near-surface stable layer, such as a nocturnal inversion, or above a sloping frontal surface. Elevated storms can be initiated from air above the boundary layer (“elevated initiation”) or can evolve from surface-based storms (e.g., Corfidi et al. 2008; Parker 2008). In the latter case this most often occurs from either the preconvective environment experiencing nocturnal cooling or the storms traversing a frontal zone over which the static stability varies.

Compared with surface-based convection, it is often more difficult for the downdrafts from elevated convection to reach the surface and form a cold-pool outflow, since the source air from elevated storms tends to originate from above more stable air. If downdrafts do not reach the surface, the initially negatively buoyant downdrafts must instead reach their level of neutral buoyancy and perhaps overshoot this. Systems with no surface cold pool can still generate persistent lifting of conditionally unstable air, however, by generating a wave or bore in the stable air, as observed by Browning et al. (2010) and
Marsham et al. (2010) and noted in simulations (Crook and Moncrieff 1988; Schmidt and Cotton 1990; Buzzi et al. 1991; Dudhia et al. 1987; Stoelinga et al. 2003; Parker 2008; Schumacher and Johnson 2008; Schumacher 2009).

The frequency of elevated initiation varies widely between regions. In the maritime climate of the United Kingdom such events are relatively unusual; for example, only one of the eighteen intensive observation periods from Convective Storm Initiation Project (CSIP) featured elevated initiation (Browning et al. 2007). Elevated convection is more common in the United States (e.g., Colman 1990; Horgan et al. 2007).

In the central United States there is a nocturnal thunderstorm maximum (Wallace 1975), which at least partly results from a nocturnal maximum in the initiation of elevated convection (Wilson and Roberts 2006). Hane et al. (2008) showed that while most mesoscale convective systems (MCSs) affecting Oklahoma and Kansas during the late morning dissipate during this period (1300 to 1700 UTC), approximately 28% remain steady or intensify; the MCS considered in this paper is such a system.

Elevated convection in the central United States is often organized into large MCSs with high equivalent potential temperature ($\theta_e$) air that sustains convection embodied within a nocturnal low-level jet (LLJ; e.g., Maddox 1983; Cotton et al. 1989; Laing and Fritsch 2000). Convection associated with this conditionally unstable airstream often occurs north of quasi-stationary east–west-oriented surface fronts (Kane et al. 1987), where mesoscale lifting within or above the frontal surface (Trier and Parsons 1993) helps focus the convection.

During the International H2O Project (IHOP_2002; Weckwerth et al. 2004), which took place in the Great Plains of the United States, Wilson and Roberts (2006) recorded almost 50% of initiation episodes as being elevated, which they defined as having no detectable surface convergence in radar or mesonet data. Most of these elevated episodes occurred at night or during the early morning. Wilson and Roberts (2006) noted that systems that produced gust fronts tended to live longer than those that did not, with only 31% of their elevated initiation episodes producing detectable gust fronts, compared with 79% of surface-based episodes. The median lifetime of elevated episodes was approximately 4 h, but all 10 of the systems from elevated episodes that produced gust fronts lived longer than this.

In idealized two-dimensional simulations, Parker (2008) found that surface-based convection became entirely elevated when nocturnal surface cooling exceeded $\Delta T \leq -10$ K, and that borelike features developed in the stable boundary layer with further cooling. Propagation speeds were greater for bore-based storms than for storms with a surface-based cold-pool outflow. Therefore, these results show that predicting either when a system will no longer be able to generate a cold-pool outflow or when an elevated system can form such an outflow is important for predicting the subsequent propagation and evolution of the system (Parker 2008), as well as the likelihood of severe near-surface winds (Horgan et al. 2007). Whether a cold pool can form likely depends on both the depth and stability of the air that downdrafts must penetrate, as well as microphysical processes and environmental humidity, which together influence the diabatic cooling that contributes to the downdrafts (Blanchard 1990; Horgan et al. 2007).

When a gravity current (e.g., a cold-pool outflow) encounters a strong low-level inversion either large-amplitude bores or gravity waves can be generated, which subsequently propagate ahead of the gravity current (Crook 1988; Klemp et al. 1997). Carbone et al. (1990) suggest a progression from density current to bore to solitary gravity wave influencing convection in an observed squall line over Oklahoma. Knupp (2006) provided a detailed study of such a transition during IHOP_2002. At the surface, gust fronts tend to show almost coincident temperature decrease and wind change, slightly led by a change in pressure (Fujita 1963; Goff 1976; Engerer et al. 2008). Bores generate a long-lived pressure rise and often a wind shift, with temperature changes varying from negligible to small increases as warm air is mixed downward (Clarke et al. 1981; Knupp 2006). Solitary waves show oscillations in winds and pressure with no long-lasting vertical displacement of air, whereas bores initiate long-lasting vertical displacements. The lifting of environmental air by density currents (e.g., Morcrette et al. 2006), bores (e.g., Koch and Clark 1999), and waves (e.g., Marsham and Parker 2006) can each lead to the initiation of deep convection, with the most complete theoretical understanding probably existing for initiation by density currents (Rotunno et al. 1988).

In this paper we describe the evolution of an MCS that crossed Oklahoma on 13 June 2002. The MCS formed from a merger of four separate episodes of elevated nocturnal convection initiated ~ (100–200) km apart, which became organized into northwest–southeast-oriented lines. Eventually these storms produced downdrafts that reached the surface and formed an extensive cold-pool outflow. This process aided the reorganization of convection into a surface-based squall line that persisted until evening. There are several studies of systems that have made the transition from surface-based to elevated (Cotton et al. 1983; Wetzel et al. 1983; Bernardet and Cotton 1998; Bernardet et al. 2000) and of cold-pool formation from elevated systems (Trier et al. 2006); such transitions are known to be important for the evolution of convection (e.g., Parker 2008). There are, however, very few studies
of the transition from elevated to surface-based convection (e.g., Trapp et al. 2001; Bryan and Weisman 2006; Corfidi et al. 2008; Grim et al. 2009).

For the current case study, section 2 describes the data and models used, section 3 the observed development, and section 4 the synthesis of the properties of the cold-pool outflow generated by the MCS.

2. Data and analysis methods

In this study both operational measurements and measurements from instruments deployed especially for the IHOP_2002 field campaign are used. Locations of radars and radiosonde sites are shown in Fig. 1. Times are expressed in Universal Coordinated Time, which is local time (LT) plus 6 h, and heights are above mean sea level (MSL) or above ground level (AGL).

Data from National Weather Service (NWS) radars (e.g., Crum et al. 1993) were combined with data from the S-band dual-polarization Doppler radar (S-Pol; Lutz et al. 1995) to create a single composite low-level reflectivity field. The radar sites are shown by black squares in Fig. 1. All radars performed series of surveillance (SUR) scans, which in some cases have been built up into reconstructed range–height indicator (RHI) scans. The scan cycle resulted in a maximum of a 5-min time separation between SURs used to reconstruct RHIs. Doppler winds from all radars are shown as ground-relative winds throughout the paper.

Radiosonde sites are shown by plus signs in Fig. 1. Data from the Oklahoma mesonet (Brock et al. 1995; McPherson et al. 2007) were used to provide 5-min means of air temperature (at heights of 1.75 and 9 m), humidity (at 1.5 m), wind speed (at 2 and 10 m), wind direction (at
10 m), rainfall, and air pressure. One-degree gridded operational analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) were also used, together with 11-μm infrared imagery from the Geostationary Operational Environmental Satellite-West (GOES-W).

3. Development of the MCS

The MCS studied in this paper formed overnight between 0500 and 1000 UTC 13 June 2002 in northern Texas, the Oklahoma Panhandle, and close to the Oklahoma–Kansas border within a northeast–southwest-oriented frontal zone (Fig. 1). Near the frontal zone, cells beginning at each of the four “initiation episodes” (IEs; with IE1 occurring from 0530 UTC, IE2 from 0700, IE3 from 0900, and IE4 from 1000) consolidated into a single MCS (Fig. 2). These initiation episodes occurred southeast of a preexisting MCS (MCS A) centered in eastern Colorado at 0600 UTC (Fig. 1).

Fig. 1 shows a decaying MCS (MCS B) in eastern Oklahoma and Kansas, the initiation of which was
Table 1. Parcel ascent properties from radiosondes launched close to the MCS. The first three are close to initiation episodes; the latter four show the environment that the MCS propagated into or toward. “Lift” is the LFC minus the height of the source air. “Min lift” is the minimum lifting required to exploit CAPE of 300 J kg\(^{-1}\) or more. Heights are AGL. Soundings are from Vici (Oklahoma; VCI, <50 km from IE3 at 0847 UTC), Lamont (Oklahoma; LMN, within 50 km of IE4 at around 1000 UTC), Norman (Oklahoma; OUN), Morris (Oklahoma; MRS), and Forth Worth–Dallas (Texas; FWS).

<table>
<thead>
<tr>
<th>Location (alt, m)</th>
<th>Time (UTC)</th>
<th>Max CAPE (J kg(^{-1}))</th>
<th>Height of max CAPE (km AGL, hPa)</th>
<th>Lift for max CAPE (km)</th>
<th>(\Theta) of max CAPE (°C)</th>
<th>Min lift (km AGL)</th>
<th>Height of min lift (km AGL)</th>
<th>Surface CAPE (J kg(^{-1}))</th>
<th>Lift from surface (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>VCI (690)</td>
<td>0829</td>
<td>1844</td>
<td>0.6 (871)</td>
<td>1.5</td>
<td>33.8</td>
<td>0.2</td>
<td>2.8</td>
<td>250</td>
<td>4.0</td>
</tr>
<tr>
<td>LMN (308)</td>
<td>0828</td>
<td>1303</td>
<td>1.2 (846)</td>
<td>1.3</td>
<td>34.4</td>
<td>0.3</td>
<td>2.4</td>
<td>42</td>
<td>5.2</td>
</tr>
<tr>
<td>LMN (308)</td>
<td>1126</td>
<td>1125</td>
<td>1.3 (834)</td>
<td>1.0</td>
<td>34.3</td>
<td>0.2</td>
<td>2.4</td>
<td>17</td>
<td>4.8</td>
</tr>
<tr>
<td>OUN (396)</td>
<td>1118</td>
<td>2004</td>
<td>0.94 (866)</td>
<td>0.7</td>
<td>32.9</td>
<td>0.7</td>
<td>0.9</td>
<td>33</td>
<td>5.1</td>
</tr>
<tr>
<td>MRS (220)</td>
<td>1130</td>
<td>531</td>
<td>2.1 (776)</td>
<td>0.5</td>
<td>36.5</td>
<td>0.2</td>
<td>2.9</td>
<td>0</td>
<td>—</td>
</tr>
<tr>
<td>FWS (179)</td>
<td>1106</td>
<td>1340</td>
<td>0.1 (980)</td>
<td>1.63</td>
<td>27.4</td>
<td>0.1</td>
<td>0.9</td>
<td>33</td>
<td>5.1</td>
</tr>
<tr>
<td>FWS (179)</td>
<td>2306</td>
<td>364</td>
<td>0.0 (987)</td>
<td>2.6</td>
<td>24.5</td>
<td>0.2</td>
<td>2.4</td>
<td>364</td>
<td>2.6</td>
</tr>
</tbody>
</table>

discussed by Weckwerth et al. (2008) and Champollion et al. (2009). A separate MCS moved through the Texas Panhandle and dissipated by 0300 UTC (not shown). Together these two MCSs led to decreased near-surface air temperatures for several hundred kilometers ahead of the quasi-stationary synoptic front located from northern Kansas through southeast Colorado (Fig. 1). For instance, temperatures over much of Oklahoma, eastern Kansas and the Texas Panhandle were 2° to 5°C cooler than in the Oklahoma Panhandle, where the MCS-modified air was less apparent (Fig. 1).

Figure 2a shows cell clusters from three of these four initiation episodes (IE1, IE2, and IE3). By 1120 UTC, IE4 was initiating along the Oklahoma–Kansas border and IE1, IE2, and IE3 had evolved into northwest–southeast lines (Fig. 2b). Sunrise occurred in central Oklahoma at 1114 UTC and data from the Oklahoma mesonet showed significant warming of the near-surface air beginning around 1200 UTC. Between 1120 and 1300 UTC the orientation of the line of storms formed from IE3 (line W) in western Oklahoma, rotated counter-clockwise (Figs. 2b,c), and merged with the line in Texas, producing a northeast–southwest-oriented MCS leading edge by 1500 UTC (Fig. 2d). The northernmost northwest–southeast line E in Figs. 2c and 2d remained distinct from line W for several hours (Fig. 2d) before eventually merging to form a single squall line (section 3c), which dissipated at around 2100 UTC (not shown).

Radiosonde data (Table 1, discussed in section 3a) from close to the initiation episodes reveal that most, if not all, were elevated and likely fueled by air with potential temperature \(\theta\) of around 34°C (\(\pm 0.4°C\)). Figure 3 shows the ECMWF analysis of the height of the 34°C \(\theta\) surface, which together with a wind shift delineates a lower-tropospheric frontal zone. Cells in Fig. 2a formed in convergence lines at the northern end of the low-level jet (Augustine and Caracena 1994), where the ECMWF analysis shows upward motion magnitudes exceeding 0.1 Pa s\(^{-1}\). Thus, strong southerly winds with a low-level jet transport moisture northward, up the sloping \(\theta\) surface (Fig. 3) in a manner similar to that discussed in the simulations of Trier et al. (2006).

### a. The environment and initiation of the elevated convection

Radiosondes launched from Vici at 0829 UTC and Lamont at 0828 and 1126 UTC (Fig. 4; Table 1) were in the region affected by previous MCSs (Fig. 1) and characterize the environment of IEs 2, 3, and 4. These soundings had stable stratification from the surface to between 900 and 780 hPa, and distinct elevated layers with high relative humidity and significant CAPE. The Vici 0829 UTC sounding (Figs. 4a,c) was close in space and time (<50 km and 15 min) to IE3. The Lamont soundings at 0828 (Figs. 4b,d) and 1126 UTC were within 50 km of IE4, which initiated at around 1000 UTC.

The 0.6 to 1.3 km AGL heights of maximum CAPE were similar for all three of the soundings close to the initiation episodes (top three rows of Table 1) and vertical displacements of 1.0 to 1.5 km were required for lifted air parcels to reach their level of free convection (LFC). The heights for the location of minimum lifting of 0.2–0.3 km required to exploit at least 300 J kg\(^{-1}\) of CAPE were higher (2.4–2.8 km AGL). In particular, for the Vici sounding (close to IE3, which was well observed by the S-Pol) the two layers of CAPE were at 0.4–1.0 km and 1.6–3.0 km AGL (respectively labeled A and B in Fig. 4a). The upper layer needed at least 0.9 km of lifting to reach its LFC and the lower layer, at least 1.0 km. The hodographs from 0830 UTC indicate southerly or south-easterly low-level jets, with maximum speeds of 8 m s\(^{-1}\) at 900 hPa at Vici (Fig. 4c) and 9 m s\(^{-1}\) at approximately 950 hPa at Lamont (Fig. 4d).

Although only one of the four initiation episodes (IE3; section 3a) had radar and radiosonde data sufficient to observe it in detail, all three radiosondes close to the
Initiation episodes required substantial vertical displacements (4.0 to 5.2 km) for surface air to reach its LFC, with much less lifting required for the elevated conditionally unstable layers (Table 1). We therefore conclude that IE2, IE3, and IE4, which initiated above the cold near-surface air left by previous MCSs (Figs. 1 and 2), must have been elevated. IE1, which initiated closer to the warmer near-surface air over the Oklahoma Panhandle (Figs. 1 and 2), may have been elevated or surface based. The location of IE1 within the region of convergence generated by the low-level jet suggests it may have been elevated, but the proximity of IE1 to MCS A (Fig. 1) suggests that a secondary initiation mechanism related to an MCS-generated circulation (e.g., Carbone et al. 1990) is possible. IE3 and IE4 initiated along two distinct northwest–southeast-oriented lines, where an elevated convergence line generated by the LLJ is likely. The lack of radar and surface data near IE2 makes its initiation mechanism more difficult to determine.

**INITIATION OBSERVED USING THE S-POL RADAR**

Only IE3 (Figs. 2a, b) could be observed using the S-Pol radar. During IE3 cells were initiated where an apparent southeastward-propagating gravity wave intersected a northwest–southeast oriented convergence line (Fig. 5). The wave source could not be identified, but the deep convection and cold-pool outflows to the west are possible sources. The first reflectivities exceeding 6 dBZ occurred at 0847 UTC in a 4° SUR scan (not shown) at ranges of approximately 70 and 90 km (corresponding to heights of 5 and 6.5 km AGL), and Fig. 5a shows weak cells approximately 100 km southeast of the radar in the 0911 UTC 1.2° scan. The wave (shown by southwest–northeast banding in Fig. 5a) could be most clearly seen in 1.2° and 2° scans from the S-Pol radar but could not be seen in the 0° scans. The convergence line is visible in the 1.2° Doppler winds at 0911 UTC (approximately along the dashed black line in Fig. 5b) and was almost stationary after 0600 UTC. It often displayed two convergence maxima, approximately 30 km apart (e.g., in Fig. 5b the second maximum is indicated by the dotted black line), and there was evidence of cells forming on both maxima. The convergence line was less clear in the 0.5° scan and much less clear in the 0° scan (but still detectable) and was also just visible in the limited surface wind data available. The fact that there was some surface signature of this elevated initiation episode suggests that some episodes classified as surface-based by Wilson and Roberts (2006) could have been elevated.

Figure 5c shows a reconstructed RHI through the wave, along the direction of the solid yellow lines in Figs. 5a and 5b. Using the vertical variations in contours of reflectivity...
to estimate wave displacements, Fig. 5c suggests that the wave has a wavelength of approximately 20 km, and an amplitude of 750 m, at 5 km AGL. The estimated vertical displacements induced by the wave increase with height, at least up to 6 km; at a range of 120 km wave amplitudes are perhaps 1 km at a height of 6 km and 400 m at a height of 3 km. For the 0829 VCI sounding, this wave amplitude is greater than the 0.2-km minimum lift required to lift air to its LFC from 2.8 km AGL (Table 1). Thus, the cells from IE3 may have been initiated from the upper moist layer at approximately 2.8 km AGL (650 hPa; Fig. 4a), where the wave induced lifting was greater than

Fig. 4. Skew-T and hodograph plots from radiosondes at (a),(c) Vici (VCI) at 0829 UTC (close to IE3), and (b),(d) Lamont (LMN) at 0828 UTC (close to IE4). Hodographs are labeled with pressures in hundreds of hPa. Layers A and B in (a) are referred to in the text.
for the lower layer at 840 hPa, but CAPE would subsequently be exploited from both the elevated moist layers.

Although the wave was evident between 0830 and 0953 UTC, it was impossible to visually track wave crests from one SUR scan to another. A Hovmöller diagram (Fig. 6) of reflectivity along the yellow line shown in Fig. 5 toward the IE3 cells initiating in Fig. 5a showed weak reflectivity signals (approximately $\sim 5$ dBZ) propagating at speeds between 3 and 9 m s$^{-1}$ for approximately 1 h. These are in some cases later connected to stronger reflectivities from clouds (i.e., are consistent with initiation by the wave). Figure 6 does not allow the wave source to be identified but, as already noted, the deep convection and cold-pool outflows to the west are possible sources.

Following Crook (1988), we investigate mechanisms that may have trapped gravity wave energy in the lower troposphere by examining the Scorer parameter calculated from the 0828 UTC Vici sounding (Figs. 4a,c). Waves with horizontal wavenumber $k$ and horizontal phase speed $c$, propagating through an atmosphere with Brunt–Väisälä frequency $N$ with base-state winds $u$, satisfy
\[
\frac{d^2w}{dz^2} + m^2w = 0, \quad \text{where} \quad m^2 = \ell^2 - k^2,
\]

where \( \ell^2 = N^2/\left[(u - c)^2 - u_{wz}(u - c)\right] \) is the Scorer parameter, \( w \) is the vertical velocity, and \( u_{wz} = \frac{d^2u}{dz^2} \). If \( m^2 \) is positive then an oscillatory solution exists for \( w(z) \), whereas if \( m^2 \) is negative, then \( w(z) \) decays with height (i.e., wave trapping occurs). Figures 4a and 7a show that the stable stratification topped by almost dry adiabatic layers between 800 and 450 hPa supports wave trapping as \( N^2 \) largely decreases with height from the surface to 6000 m (leading to a decrease in the first term in the Scorer parameter with height). Figure 7b shows the northwesterly wind profile, which is similar to that shown in Fig. 3 of Crook (1988), where the wind curvature from a low-level jet is found to favor wave propagation in the direction opposite to the jet. Figure 7b therefore suggests that in this case the low-level jet will favor trapping of southeastward-propagating waves via the second (wind curvature) term of the Scorer parameter. Figure 7c shows the Scorer parameter for waves traveling southeast at 5 m s\(^{-1}\) (approximately the observed velocity), with the solid and dashed lines respectively indicating values with and without the observed winds. Figure 7c indicates that the low-level jet should enhance trapping, since it generates strongly negative values of the Scorer parameter between 2 and 3 and between 3.5 and 4.5 km.

Additional support for this interpretation of the observed feature being a trapped wave is obtained by comparing the observed wave velocity (3 to 9 m s\(^{-1}\)) with...
the theoretical phase velocity for waves in a duct with Brunt–Väisälä frequency $N$ capped by a deep neutral layer, described by Nappo (2002, p. 100). Using $N^2 = 3 \times 10^{-4}$ s$^{-2}$ (Fig. 7a) and horizontal wavelengths between 10 and 25 km (Fig. 5c), the fundamental mode has a phase velocity between 11.7 and 14.3 m s$^{-1}$. Considering the opposing 900-hPa flow of 9 m s$^{-1}$, the observed velocities are consistent with this theoretical value.

We therefore conclude that IE3 occurred when elevated conditionally unstable air was lifted to its LFC by a combination of a convergence line and a propagating wave, although the wave source could not be identified. Convection then developed along the convergence line (Fig. 2b).

### b. Observations of the cold pool, bores, and waves in north-central Oklahoma and subsequent secondary initiation

In this section the development of the cells that had formed during IE4 (Fig. 2b) is discussed, focusing on the generation of bores and waves by the downdrafts from the previous convection and the subsequent secondary initiation [here, as used by Bennett et al. (2006) and Browning et al. (2007), “secondary initiation” is used to refer to initiation of deep convection by feature that was itself generated by previous deep convection].

As discussed in section 3a, the first cells within IE4 initiated along a northwest–southeast-oriented convergence line close to the Oklahoma–Kansas border at around 1000 UTC. These cells moved slowly, passing over the VNX radar site (see Fig. 1 for location) at 1120 and 1300 UTC (Figs. 2b,c). Figures 8a and 8c show the cells over the VNX radar at 1243 and 1253 UTC. There were four mesonet stations close to the VNX radar (OK27, OK67, OK59, and OK16) that together with the radar data are used to examine the outflows from the convection. Figure 8b shows a southward bulge of yellow–orange–red winds directed away from the VNX radar associated with the convection situated over the radar (Fig. 8a). At this time convection also occurs over mesonet stations OK27 and OK67 (Fig. 8a). The banded structures in the red–orange Doppler winds toward the south and east 10 min later (Fig. 8d) are suggestive of bores or waves generated by downdrafts from the convection interacting with the stable air at the surface. The evolution of the two SURs in Fig. 8 suggests that these bores or waves may have been responsible for secondary initiation of the cells between the radar and OK16. This process therefore appears to be similar to that discussed by Fovell et al. (2006), where waves generated by a squall line initiate cells ahead of the surface gust front, except that in this case the cold pool is either weak or absent.

Figure 9 is a Hovmöller diagram of Doppler velocity and reflectivity along a line from VNX to OK16 (53 km southeast of VNX), with the pressure signal from OK16 (Figs. 8a,c) superimposed. The full time series of data from OK16 is shown in Fig. 10a. The pressure increase of around 1 hPa at 1230 UTC was accompanied by a wind direction change from southerlies to northeasterlies and was associated with the surface convergence line crossing the station. In Fig. 9 the stepped rise of about 2 hPa from around 1330 to 1400 UTC is associated with a propagating feature (10 m s$^{-1}$) in the Doppler winds and the reflectivity. This feature corresponds to the orange bulge in Doppler winds (Figs. 8b,d) and the reflectivity signal corresponds to the cells that appear to result from secondary initiation. When the feature (gray shading in Fig. 10a) reached OK16 at ~1330 UTC, a wind shift to northerlies, a wind speed increase from 5 to 8 m s$^{-1}$, and no significant surface temperature or water vapor mixing ratio ($q_v$) change was observed. This surface signature is more typical of a bore than a density current outflow, which would likely have a temperature decrease and perhaps a larger wind speed increase. Therefore, although the bore was no longer obvious in the SUR by the time it reached OK16, together the radar, surface, and sounding data are consistent with secondary initiation by a bore generated by downdrafts from the elevated convection interacting with the stable nocturnal boundary layer. A distinct cold-pool outflow was never observed at OK16 (Fig. 10a) because the remainder of the pressure rise at OK16 from 1400 to 1600 UTC coincided with heavy rainfall, increasing $q_v$, and a 1°C warming.

Mesonet stations to the south (OK59) and east (OK67) of VNX show signatures somewhat similar to those from OK16. OK59, 40 km south of VNX and under the orange band leading the outflow in Fig. 8d, recorded a three-step pressure increase of approximately 5 hPa from 1200 to 1445 UTC (Fig. 10b). The second step of ~1 hPa indicated by gray shading in Fig. 10b was accompanied by a wind speed increase from approximately 2 to 4 m s$^{-1}$, a wind direction change, and surface warming of less than 1°C. These surface changes are more typical of a bore than a cold-pool outflow. The main pressure increase of 3.5 hPa beginning at 1330 UTC was accompanied by the onset of rain, gradually increasing wind speeds, gradual drying, and a weak temperature decrease of ~2°C. A bore ahead of a cold pool (hypothesized from OK59 data) is consistent with the Doppler data in Fig. 8d, which shows an orange band over OK59 with the cold-pool outflow to its north.

OK67 (Figs. 8c,d) was affected by a 1-h local pressure maximum of ~2 hPa (gray shading in Fig. 10c), which was accompanied by a wind speed and direction change, with very little change in temperature or $q_v$. This short-lived pressure increase is consistent with the wave structure.
FIG. 8. SURs at (a),(b) 1243 and (c),(d) 1253 UTC from the VNX radar showing (a),(c) reflectivities and (b),(d) Doppler winds. The locations of the VNX radar and mesonet stations OK27, OK59, OK67, and OK16 are labeled in (a)–(d). Wind barbs are as defined for Fig. 2. Reconstructed (e) reflectivity and (f) Doppler RHIs toward OK59 [along yellow lines in (a) to (d)] at 1253 UTC. The area of the SURs is shown by the dashed yellow box in Fig. 2c.
(banded orange in velocity SUR) affecting OK67 in Figs. 8b and 8d. This was followed by a much larger pressure increase at 1400 UTC, which was accompanied by a cooling of up to 2°C but very little change in wind speed (Fig. 10c). Thus, as was the case for the other two mesonet stations, any cold pool accompanying the convection was very weak and was preceded by wavelike features.

c. Observed transition from elevated convection to a surface-based squall line in central Oklahoma

In this section we describe the evolution of initial elevated cells to a surface-based squall line (feature E in Figs. 2c,d). Reconstructed RHI scans showed that the depth of the outflow from the convection initiated close to the VNX radar increased significantly with time. At 1253 UTC, the reconstructed RHI toward OK59 (Fig. 8f) indicates that the orange layer of the northerly ground-relative winds, assumed to correspond approximately to the cold pool–bore, reaches a depth of only 750 m, approximately that of the stable boundary layer in the 0828 (Fig. 4b) and...

FIG. 9. Hovmöller diagram from the VNX radar through mesonet site OK16, the location of which is shown by the vertical white line. Colors show radar Doppler velocities, and black lines show radar reflectivities (both from 0.5° scans). The thick white line shows the time series of pressure data from OK16, with the full data shown in Fig. 10a.

FIG. 10. Data from mesonet stations close to the VNX radar: (a) OK16 (to the southeast of VNX), (b) OK59 (to the south), and (c) OK67 (to the east). (top) Solid lines show pressure; dashed lines, temperature; and dotted lines, the water vapor mixing ratio (q_v). (bottom) Solid lines show wind speed; dashed lines, wind direction; and dotted lines, accumulated rainfall. Labels and shading are discussed in the text.
1130 UTC (not shown) LMN soundings. Ahead of this, to the south, low-level winds are southeasterly and are forced to rise over the outflow. By 1424 UTC, however, outflow depths have deepened to approximately 2 km (Figs. 11c,d), over twice the depth observed at 1253 UTC.

The cells within the dashed red box in Fig. 11a formed ahead of the main precipitation system, along the same convergence line that led to IE4 (Fig. 2b). However, the forthcoming synthesis suggests that waves–bores emanating from the main precipitation system, which intersect the convergence line, could have contributed to the cell initiations. Figure 11 indicates that the MCS did not yet have a squall-line-type organization; high reflectivities were widespread rather than being confined to the leading edge of the outflow. The system velocity along the cross-sectional line shown in Fig. 11a (approximately along its direction of travel) was \(\approx 6 \text{ m s}^{-1}\). There is therefore zero or weak rear inflow shown in the Doppler cross sections, since the radial velocities show approximately 6 m s\(^{-1}\) outbound.

A Hovmöller diagram (Fig. 12), created along a line approximately through seven mesonet stations, perpendicular to the eventual squall-line MCS and along its direction of travel (dashed line in Fig. 2d), allows a synthesis of the evolution of the cells formed during IE4 and the subsequent development of the system. Figure 12 shows the gradually decreasing width of the system along the line, as the system evolves from a northwest–southeast-oriented line of convection to a northeast–southwest-oriented squall line. The duration of the positive surface pressure anomaly (black lines) also decreases in time, as does its magnitude (4.5 to 2.3 hPa). The wind speed anomaly increases in time as the surface cold pool strengthens (excluding the northwesternmost station at \(x = 0\) km, where the cold pool is stronger than at \(x = 90\) km, but it lags the first intense convection by over 1.5 h). There is a strong surface mesolow at the back of the system, which again decreases in magnitude with time, and there is a suggestion of a minimum in reflectivity within the mesolow. The position of the mesolow (relative to the reflectivity gradient), along with its large amplitude and associated abrupt wind shift to northwesterlies (Fig. 12b), suggests that it is likely the subsidence-induced wake low (e.g., Johnson and Hamilton 1988, Stumpf et al. 1991) commonly observed in large MCSs.

Reflectivities generally increase with time, particularly at 1430 and 1530 UTC (Fig. 12). It is shown later that the second increase at 1530 UTC appears to correspond to the transition to surface-based convection. The largest reflectivities are observed just behind the leading edge of the surface cold-pool outflow (the diagonal black solid line in Fig. 12), which is detectable from the temperature decreases (Fig. 12a) and wind speed increases (Fig. 12b). From 0 to 2 h ahead of the cold pool, a small pressure increase is observed, evident between the dashed and solid lines in Fig. 12, especially from \(x = 0\) to 200 km. Since the pressure rise evolves toward a pressure rise and fall, we interpret this feature as a bore, generated by the cold pool interacting with the stable near-surface air, evolving to a wave. The reflectivity in Fig. 12 shows that the wave or bore is associated with the initiation of convection ahead of the cold pool, similar to that discussed by Fovell (2005). The initiation is particularly widespread between 1530 and 1600 UTC; the cold pool of the MCS then caught up with the slower-moving cells initiated ahead of it and so these cells lead to little change in the overall speed of the MCS. Ahead of this wave or bore, between the dashed and dotted lines, there is some evidence of a propagating wave in the pressure data. Sporadic cells of high reflectivity within this zone suggest convection initiation by the wave, but because of data limitations we are unable to confirm this.

A radar fine line propagating ahead of the precipitation system at 1524 UTC (Fig. 13a) is associated with the cold-pool outflow diagnosed from the surface mesonet data (cf. Fig. 12). Radar cross sections of Doppler velocities from the KLX radar (not shown) showed the northwesterly ground-relative winds, assumed to correspond to the outflow, reaching a depth of over 3.5 km in places, which persisted for the next hour. By 1624 UTC there was widespread secondary initiation in bands oriented approximately 45° to the squall line (Fig. 13b). This banded secondary initiation structure is reminiscent of that simulated and described by Bryan et al. (2007), although the \(\approx 20\)-km horizontal spacing between the rolls is much greater in the current case. Similar but less pronounced structures are also observed 3 h earlier farther west, in the western half of the system (shown in Fig. 18a).

Bryan et al. (2007) showed that roll circulations could form in the leading deep convection of a squall line if a conditionally unstable layer of air was lifted in quasi-horizontal flow over a cold pool in the presence of vertical shear. Such lifting to saturation was responsible for...
forming a moist absolutely unstable layer (MAUL) that supported deep convection.

Of the soundings ahead of the MCS, the Norman (OUN) 1118 UTC sounding (Fig. 14) is most representative of the environment that the storms propagated into. In both the OUN 1118 UTC and the Morris (MRS) 1130 UTC soundings the maximum CAPE and height of minimum lifting to overcome convective inhibition (CIN) were elevated (Table 1). Farther south, at Dallas (FWS; Fig. 1), which is located ahead of the southeastward-moving MCS as it was later weakening, the maximum CAPE was essentially at the surface at both 1106 and 2306 UTC. There was much more CAPE at OUN than MRS. The MRS sounding shows the “onion” structure typical of a profile after deep convection and so this difference between the soundings is consistent with the effects of the preceding MCS B (Fig. 1; Weckwerth et al. 2008; Champollion et al. 2009) being much larger at MRS. The surface CAPE at OUN would have become more significant as the boundary layer warmed after sunrise at 1114 UTC, although shading from the system’s cirrus anvil (e.g., Marsham et al. 2007a,b) is expected to have reduced this close to the MCS (the anvil had reached by OUN at 1245 UTC).

![Fig. 11. SURs and reconstructed RHIs from VNX showing (a),(c),(e) reflectivities and (b),(d),(f)Doppler winds. RHIs are toward TLX along the line shown in the SURs. (a),(b) 1424 UTC SURs, with the VNX and TLX radars and the OUN radio sounding site labeled in each. The dashed yellow box in (a) indicates the area shown in Figs. 8a–d; the cells within the dashed red box are referred to in the text. (c),(d) 1424 UTC RHIs; (e),(f) 1444 UTC RHIs.](image-url)
FIG. 12. Hovmöller diagram along the direction of travel of the MCS (dashed yellow line in Fig. 2d) and approximately through seven mesonet stations. Colors show radar maximum radar reflectivities within a region 50-km north of the line. For each mesonet station, the thick black lines show pressure perturbations from the 1000–2000 UTC station mean (scaled so that 1 hPa is plotted as 10 km). (a) The dashed black line shows temperature (scaled so that 1 K is plotted as 5 km). (b) For the wind arrows, the short barb = 2 kt, long barb = 5 kt, and pennant = 10 kt. The diagonal black lines are discussed in the text.
If we assume that the profile above the boundary layer at OUN did not change significantly between 1118 and 1600 UTC, the LFC of surface air parcels at around 1600 can be estimated using a combination of the mesonet data (to define surface air properties) and the sounding. Under this assumption, Fig. 15a shows that at 1530 UTC the LFC of surface air in the region ahead of the squall line, where secondary initiation occurred, was 3.5 km. Therefore, it appears that the sudden widespread initiation observed at this time in Figs. 12 and 13b occurred when the surface air had warmed sufficiently for the lifting provided by the system to be sufficient for near-surface air to reach its LFC. Before this time a cold pool was present, but it is not expected to have been able to lift near-surface air to its LFC, although a contribution of near-surface air before this time cannot be ruled out. The estimated CAPE for the surface air parcels at this time of widespread initiation was around 500 J kg$^{-1}$ and is much less than the 2004 J kg$^{-1}$ for air at 1.34 km AGL, which is also expected to have contributed to the initiation, which occurred both directly over the cold-pool outflow and with the bore ahead of it. The lack of initiation around Morris (MRS) compared with Norman (OUN) in Fig. 13 is consistent with the greater effects of the preceding MCS B there (Figs. 1 and 14; Table 1).

The system was originally oriented northwest–southeast, which corresponded to the orientation of the convergence that led to the initiation (Figs. 2, 3, and 5). The reorientation of the convection to a northeast–southwest-oriented squall line after sunrise (Figs. 2 and 13) occurred as the cold pool deepened and was likely influenced by the
interaction of the environmental vertical shear with the developing cold pool. Rotunno et al. (1988) used idealized two-dimensional simulations to explain the organization of convection along the downshear edge of cold pools as the result of a horizontal vorticity balance between the cold-pool buoyancy gradient and the environmental vertical shear. This preference for development on the downshear edge of the cold pool has been demonstrated further in subsequent fully three-dimensional simulations of squall lines (e.g., Weisman 1993; Skamarock et al. 1994; Trier et al. 1997). Recently, French and Parker (2010) suggested that the environmental vertical shear within the conditionally unstable layer is most critical for deep lifting at the cold-pool edge. Consistent with the above modeling studies, the $\Delta U = 10 \text{ m s}^{-1}$ ($8 \times 10^{-3} \text{ s}^{-1}$) vertical shear within the most unstable 925–800-hPa layer of the 1118 UTC OUN sounding is directed from 320°, approximately normal to the concentrated northeast–southwest-oriented line of leading-edge reflectivity (Fig. 13).

d. Observed formation of the squall line in western Oklahoma

We now return to the development of the MCS in western Oklahoma. Figures 2b and 2c show that between
1120 and 1300 UTC the orientation of the line of convection immediately to the southeast of the Oklahoma Panhandle (labeled W in Figs. 2b–d) rotated from a northwest–southeast-oriented line into a squall-line-like structure oriented from southwest to northeast. Throughout this period, the most intense convection in the western part of the system (reflectivities $>$40 dBZ) was still spatially separated from the intense convection in the eastern part of the system (labeled E and discussed in section 3c). This observed change in the orientation of the western system occurred between the S-Pol, VNX, and FDR radars (Fig. 2), so radar data are not available from low levels. However, the combination of the available radar and surface mesonet data still provides useful insights into this process.

Figures 16c–f show reconstructed RHIs intersecting the VNX radar, surface mesonet station OK84 (Fig. 17a), and the nearly stationary northwest–southeast-oriented line of convection (Figs. 16a,b). OK84 was affected by heavy precipitation at the time of both of the RHI cross sections. Unlike for the time series from mesonet stations located farther east and discussed in section 3b (Fig. 10),
at OK84 a very clear cold-pool signature was observed, with sudden temperature and $q_v$ decreases of 3.5°C and 3 g kg$^{-1}$, respectively, and a 4–10 m s$^{-1}$ wind speed increase at around 1200 UTC (dashed vertical line in Fig. 17a). The arrival of the cold pool at OK84 approximately coincides with the first observed penetration of the rear inflow toward the surface, which is shown in the 1207 UTC RHI (Fig. 16f) by the orange and red colors; this descent of the rear inflow is expected to contribute to the observed increase in near-surface winds (Mahoney et al. 2009). Here, the rear inflow may have reached the surface but the radars are unable to detect winds at ground level.

At 1210 UTC, approximately 20 min before the cold-pool outflow reached OK21 (which was located at $\sim$35 km in the RHIs; Figs. 18c,d), low-level southerlies (seen in the Doppler radar and also observed in mesonet data ahead of the system) were being lifted over a 5.5-km-deep layer of northerlies. Using the 0829 UTC Vici (VCI) radiosonde and the surface mesonet data, we estimate that the LFC of the surface air ahead of the convection was $\sim$3.5 km AGL (Fig. 15b). We therefore conclude that as this northwest–southeast line rotated to form a squall line and a clear surface cold pool, it then became able to lift surface air to its LFC. The estimated LFCs in this region where the surface-based convection occurred were lower than those approximately 75 km east of VCI (yellows to reds in Fig. 15b), which is later shown to be consistent with the relatively warm boundary layer in the west and the weaker influence of MCS B in western Oklahoma (Figs. 1, 14, and 19b). The combination of mesonet data with the Vici radiosonde used to generate Fig. 15b showed surface CAPE values ahead of the convection of only 500–600 J kg$^{-1}$. Thus, as we found for the eastern part of the MCS, when the data first suggest that surface air parcels could be lifted to their LFC, the highest CAPE values were still elevated, with 1844 J kg$^{-1}$ of CAPE at a height of 1.29 km for the Vici radiosonde (Fig. 8a; Table 1). However, surface CAPE is expected to have eventually dominated (see the FWS sounding in Table 1). As for central Oklahoma (section 3c), the Vici and Lamont radiosondes (Fig. 4) indicate a veering lower-tropospheric wind profile, with northwesterly shear that favors a northeast–southwest-oriented squall line after the surface cold pool develops (Rotunno et al. 1988).

4. Synthesis of observations of the cold-pool outflow from the MCS

Regional aspects of the cold pool generated by the MCS have been discussed in the preceding three subsections. In this section data from the individual Oklahoma mesonet stations are combined to give a broader
FIG. 18. Radar data from 1215 UTC. (a) Radar mosaic reflectivities; (b) PPI of Doppler winds from FDR. (c),(d) Reconstructed RHIs from FDR through OK21 [along the yellow line shown in (a) and (b), with FDR located outside of the area shown in the PPIs], showing (c) reflectivity and (d) Doppler winds. The cold-pool outflow was detected using data from OK21 at 1230 UTC.
overview of the evolution of the cold pool as the MCS crossed Oklahoma.

The MCS and its earlier components generated a surface cold pool through much of the precipitation life cycle; gust fronts were detectable from the IE1 cells as soon as they arrived within range of the S-Pol radar, and the mature squall line had an extensive cold pool. The cold pool was tracked by finding the time that a wind shift related to a sudden cooling occurred for each of the mesonet stations, consistent with climatology of cold pools studied using the Oklahoma mesonet by Engerer et al. (2008). If the wind shift was not clear, the change in wind speed was used. The temperature fall, pressure rise, and wind speed change from 2 h before to 3 h after this time was then calculated for each station. In this way, although the time of crossing was determined in a subjective manner, these statistics were robust to the precise time used. Figure 19 shows these statistics contoured over the domain.

The isochrones in Fig. 19a show the system traveling from northwest to southeast across the mesonet. The slower progression across northeastern Oklahoma compared with central and western Oklahoma is consistent with the impacts of the preceding MCS B on the profile there (Figs. 1 and 14; Table 1). Figure 19b shows the temperature decrease generated by the cold pool (colors), the near-surface air temperature before the arrival of the cold pool (black contours), and near-surface air temperature at 1130 UTC (pale gray contours). These pale gray contours show the colder air left around by MCS B in northeastern Oklahoma.

After 1600 UTC (bolder contour in Fig. 19a) the leading edge of the MCS was located in central Oklahoma, where it exhibited the classic surface signature of a pressure rise, wind speed increase, and temperature decrease as it moved southeastward within a daytime convective boundary layer (Fig. 19). In northern Oklahoma, the impact of the nocturnal and early morning surface cold pool was more complex and spatially varying. Just east of the Oklahoma–Texas border there were strong temperature and $q_c$ decreases, a pressure rise, and a wind speed increase as dry air descended to the surface behind the MCS.
leading edge. Farther east, in north-central Oklahoma, slightly smaller pressure signals (up to 5.5 hPa rather than 7 hPa) were accompanied by much smaller temperature decreases (less than 2°C rather than over 4°C), smaller wind speed increases (up to 4.5 m s⁻¹ rather than up to 6.5 m s⁻¹), and 1 g kg⁻¹ water vapor mixing ratio \( q_v \) increases rather than 2 g kg⁻¹ decreases.

It is notable that Fig. 19b shows that spatial variations in cold-pool properties observed at the surface (colors) are correlated with temperatures before the arrival of the cold pool (black contours). These pre-existing temperatures are largely controlled by the impact of MCS B, which left a region of reduced near-surface temperatures in northeastern Oklahoma (Fig. 1, and pale gray contours in Fig. 19b), as well as a more stable profile (Fig. 14; Table 1). In particular, Fig. 19b shows that the large surface temperature decreases and wind speed increases observed just east of the Texas–Oklahoma border are where preconvective air temperatures were greater than 20°C, whereas farther east, where preconvective temperatures were less than 20°C, temperature decreases were smaller (with comparable pressure increases). The cold pool in this area, diagnosed from surface mesonet data, was highly heterogeneous. Figures 17b and 17c show data from two mesonet stations (OK108 and OK48), which are 26 km apart and are, respectively, 47 and 66 km from OK84. Heavy precipitation from the MCS tracked directly over all three stations, but OK48 shows a weak cold pool with a slow temperature decrease (2.5°C in 2 h) whereas the effect at OK108 is much more rapid (3°C in 30 min). Similarly, maximum wind speeds at OK108 are larger (12 m s⁻¹ compared with 8 m s⁻¹). In addition, there is a narrow corridor of increased temperature falls from the cold pool and some increased drying oriented northwest to southeast in central Oklahoma (Figs. 19b,c), which is collocated with a strip of warmer pre-existing air from before the cold pool arrived (indicated by a dashed line in Fig. 19b). We conclude that MCS B led to a region of lower near-surface air temperatures in northeastern Oklahoma (Figs. 1 and 19b) and a more stable profile there (Fig. 14) and that this affected the surface cold-pool strength of the MCS studied in this paper. The more homogeneous effects of the cold pool in southern Oklahoma are consistent with the weaker impacts of MCS B there. These conclusions are consistent with Horgan et al.’s (2007) discussion of the impacts of low-level stability on the strength of surface cold pools and their associated winds.

5. Summary and discussion

Nocturnal convection is difficult to forecast accurately. Compared with surface-based convection, orography, and other easily detected forcing features such as boundary layer convergence lines play a less prominent role in initiating elevated convection. Elevated initiation is instead often controlled by elevated features such as low-level jets and waves, with low-level stability and jets both favoring wave trapping (see, e.g., section 3a). Nocturnal elevated initiation was common during IHOP_2002 and resulting precipitation systems were less likely to produce surface cold pools and be long lived (Wilson and Roberts 2006).

We have described the observed evolution of an MCS that formed at night in a frontal zone on 13 June 2002 during IHOP_2002. The first convective cells formed ahead of the cold front. At least three of the four initiation episodes that led to the MCS were elevated, initiating from two layers of conditionally unstable air at 650 and 870 hPa, which were located above a stable nocturnal boundary layer that had been affected by previous MCSs. The elevated system formed from the initial cells went on to form a surface-based squall line. For this case study we conclude that

- The elevated convection initiation overnight occurred along northwest–southeast-oriented convergence lines, at the terminus of a southerly LLJ (section 3a). Although initiation occurred from moist conditionally unstable layers located above the boundary layer, convergence was sometimes observed both at the surface and above the nocturnal boundary layer.
- The best observed initiation event (IE3) occurred where a southeastward-propagating wave intersected such a convergence line (section 3a). Calculation of the Scorer parameter using nearby radiosonde measurements showed that the stable layer capped by near-neutral layers combined with the southerly low-level jet to favor trapping of the wave, which could have been excited by a preexisting upstream nocturnal MCS.
- Early in the life cycle of the MCS, when downdrafts from the convection interacted with the stable nocturnal boundary layer, bores and waves propagated ahead of the cold pool and initiated further deep convection (section 3b).
- The cold-pool surface properties observed by the surface mesonet varied substantially across the area affected by the MCS at night (section 4). The weaker cold pool in the eastern part of the domain appears to be a result of the greater impacts of a previous MCS there, which led to a more stable profile and lower near-surface air temperatures. As the boundary layer warmed after sunrise, near-surface changes due to the cold pool were more uniform, with pressure increases, rapid cooling and drying, and wind speed increases typical of large squall-type MCSs.
The initially northwest–southeast-oriented lines of convection eventually formed a southwest–northeast-oriented squall line. This happened before any significant surface heating occurred in western Oklahoma (section 3d; 1300 UTC, 0700 LT), and later after some surface warming in central Oklahoma (section 3c; 1600 UTC, 1000 LT), where the impacts of the preceding MCS B were greater. In both locations the time of the reorientation of the convection approximately coincided with the development of an MCS-induced surface cold pool that was able to lift environmental surface air to its LFC, although at first the air with the most CAPE was still elevated.

To our knowledge, this paper provides the first detailed observational case study of a transition from elevated initiation to surface-based deep convection, a process that likely enhances MCS longevity. However, a single case study with limited data leaves several important questions unanswered. For instance, Wilson and Roberts (2006) found that during IHOP_2002 elevated initiation episodes that led to detectable surface gust fronts tended to generate longer-lived systems, but it is not known how often these gust fronts led to near-surface air contributing to the convection, as occurred for this case. Moreover, this study suggests that Wilson and Roberts’ (2006) use of surface convergence to define surface-based initiation may mean that elevated air may have dominated the initiation in some episodes they defined as surface based. For the current case, the available data show that the greatest CAPE was elevated during the onset of MCS reorganization into a single northeast–southwest line. The developing cold pool arising from penetrative downdrafts may then have further influenced MCS organization by helping to lift near-surface air to its LFC. The initial reorganization of the MCS into a single northeast–southwest line is a topic that we are still investigating using a numerical model. We speculate that it could be influenced by a variety of factors including interactions of internal MCS circulations with the environmental shear and larger-scale forcing associated with the evolving synoptic environment.

Acknowledgments. The authors thank all those who were involved with the IHOP_2002 field campaign. John Marsham’s visit to NCAR was funded by the British Council Researcher Exchange Programme and he would like to thank Tammy Weckwerth and Jim Wilson for hosting his visit. The authors thank Tracy Emerson for producing final versions of some of the figures and David Ahijevych, who provided the plot of NEXRAD radar mosaic and surface stations for subjective analysis in Fig. 1. Finally, we thank Dr. David Schultz and three anonymous reviewers whose comments helped to improve both the content and clarity of the paper.

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