AN INVESTIGATION OF THE DEVELOPMENT OF SUPERCELLS IN THE INDIANA AND OHIO TORNADO OUTBREAK OF 24 AUGUST 2016 USING A WRF MODEL SIMULATION

BY

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THESIS

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Abstract

On 24 August 2016, a tornado outbreak that produced 24 confirmed tornadoes impacted portions of Indiana and Ohio. Initially elevated multicellular convection transitioned to surface-based supercells between 1700 and 1830 UTC, after which time tornadoes began to occur. Such a transition is of particular interest owing to its relatively rare occurrence and because most convection-allowing models failed to predict it in this case, instead depicting a line of storms moving across the affected area.

Three surface boundaries were present prior to and during the event. The first was an outflow boundary traced back to convection in Nebraska and Iowa the previous evening, along which an elevated storm cluster formed that later developed into three discrete supercells. The second was a differential heating boundary owing to cloud shading from a leading cluster of storms. This boundary moved northeastward across Indiana and its position coincided with the formation of each of the supercells and the production of their first tornadoes. The third was a westerly wind shift behind the outflow boundary that triggered a second round of supercells, some of which also produced tornadoes. A mesoscale convective vortex, also owing to storms from the previous evening, moved across northern Illinois and Indiana during the event augmenting the low-level and deep-layer vertical wind shear.

A Weather Research and Forecasting (WRF) model simulation accurately captures the environment during this event, depicting a cluster of elevated thunderstorms developing in the vicinity of an outflow boundary during the early morning hours of 24 August. Trajectory analyses indicate that parcels entering the multicellular updrafts before 1530 UTC originate above 1 km, meaning that the convection is elevated. As diurnal heating progresses, more near-surface parcels enter the multicellular updrafts, each of which move atop convective outflow and weaken. Since nearly all of the vertical wind shear is below 1 km, mesocyclones do not form via tilting of horizontal vorticity until the storms ingest air from this layer. The storm on the southern end of this cluster becomes a supercell in the simulation around 1700 UTC, after it becomes surface based. Inflow trajectories pass through an area of high 0-1 km storm-relative helicity, likely owing to anvil shading southeast of the simulated storm. When these helicity-rich parcels enter the supercell updraft, strong near-surface rotation develops.
A novel analysis of the perturbation pressure field from the WRF model output indicates that the development of relatively large vertical perturbation pressure gradients coincide with when near-surface air begins to enter the updrafts, resulting in upward accelerations in the lowest 2 km, below the level of maximum rotation, and downward accelerations above this level. In strengthening updrafts, upward-directed perturbation pressure accelerations due to buoyancy may offset the downward-directed non-linear perturbation pressure accelerations above the level of maximum rotation, allowing the updrafts to deepen and intensify. This simulation suggests that the transition from disorganized elevated convection to surface-based supercells begins as more near-surface air gradually enters the multicellular updrafts, leading to an increase in updraft rotation in new cells and in the magnitude of the upward-directed non-linear perturbation pressure gradients, which likely aid in the development of a single dominant supercellular updraft.
Acknowledgments

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Thank you Mom, Dad, siblings, and grandparents for all your support during my long college career and thank you to my friends inside and outside the department for a fun-filled graduate experience.
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# List of Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>AGL</td>
<td>Above ground level.</td>
</tr>
<tr>
<td>APB</td>
<td>Vertical acceleration owing to gradients in the buoyancy component of the perturbation pressure.</td>
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<tr>
<td>APDL</td>
<td>Vertical acceleration owing to gradients in the linear dynamic component of the perturbation pressure.</td>
</tr>
<tr>
<td>APDN</td>
<td>Vertical acceleration owing to gradients in the non-linear dynamic component of the perturbation pressure.</td>
</tr>
<tr>
<td>APTD</td>
<td>Vertical acceleration owing to gradients in the total dynamic component of the perturbation pressure.</td>
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<tr>
<td>ASOS</td>
<td>Automated surface observation system.</td>
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<tr>
<td>BRN</td>
<td>Bulk Richardson number.</td>
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<td>BWD</td>
<td>Bulk-wind difference.</td>
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<tr>
<td>CAPE</td>
<td>Convective available potential energy.</td>
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<tr>
<td>CIN</td>
<td>Convective inhibition.</td>
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<tr>
<td>CM1</td>
<td>Cloud Model 1.</td>
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<tr>
<td>DPVA</td>
<td>Differential positive vorticity advection.</td>
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<tr>
<td>EBWD</td>
<td>Effective bulk-wind difference.</td>
</tr>
<tr>
<td>EF</td>
<td>Enhanced Fujita scale.</td>
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<tr>
<td>ESRH</td>
<td>Effective storm-relative helicity.</td>
</tr>
<tr>
<td>KCMI</td>
<td>Champaign, IL, ASOS site.</td>
</tr>
<tr>
<td>KDEC</td>
<td>Decatur, IL, ASOS site.</td>
</tr>
<tr>
<td>KDMX</td>
<td>Des Moines, IA, WSR-88D site.</td>
</tr>
<tr>
<td>KDVN</td>
<td>Davenport, IA, WSR-88D site.</td>
</tr>
<tr>
<td>KILX</td>
<td>Lincoln, IL, WSR-88D site.</td>
</tr>
<tr>
<td>KIND</td>
<td>Indianapolis, IN, WSR-88D site.</td>
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<tr>
<td>KIWX</td>
<td>North Webster, IN, WSR-88D site.</td>
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<tr>
<td>LCL</td>
<td>Lifting condensation level.</td>
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<tr>
<td>Acronym</td>
<td>Definition</td>
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<tr>
<td>LFC</td>
<td>Level of free convection.</td>
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<tr>
<td>MCS</td>
<td>Mesoscale convective system.</td>
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<tr>
<td>MCV</td>
<td>Mesoscale convective vortex.</td>
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<tr>
<td>MLCAPE</td>
<td>Mixed-layer convective available potential energy.</td>
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<tr>
<td>MLCIN</td>
<td>Mixed-layer convective inhibition.</td>
</tr>
<tr>
<td>MSL</td>
<td>Mean sea level.</td>
</tr>
<tr>
<td>MUCAPE</td>
<td>Most unstable convective available potential energy.</td>
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<tr>
<td>NAM</td>
<td>North American Mesoscale model.</td>
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<tr>
<td>RAP</td>
<td>RAPid Refresh model.</td>
</tr>
<tr>
<td>RIP</td>
<td>Read/Interpolate/Plot program.</td>
</tr>
<tr>
<td>RUC</td>
<td>Rapid Update Cycle model.</td>
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<tr>
<td>SCP</td>
<td>Supercell composite parameter.</td>
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<tr>
<td>SPC</td>
<td>Storm Prediction Center.</td>
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<tr>
<td>SRH</td>
<td>Storm-relative helicity.</td>
</tr>
<tr>
<td>UH</td>
<td>Updraft helicity.</td>
</tr>
<tr>
<td>WRF</td>
<td>Weather Research and Forecasting model.</td>
</tr>
</tbody>
</table>
List of Symbols

\( \frac{\partial}{\partial t} \) Partial derivative with respect to time.
\( \frac{\partial}{\partial x} \) Partial derivative with respect to horizontal distance in the x-direction.
\( \frac{\partial}{\partial y} \) Partial derivative with respect to horizontal distance in the y-direction.
\( \frac{\partial}{\partial z} \) Partial derivative with respect to height.
\( \nabla \) Gradient.
\( \nabla_h \) Horizontal gradient.
\( \nabla^2 \) Laplace operator.
\( \zeta \) Relative vertical vorticity.
\( \eta \) y-component of the vorticity vector.
\( \theta_e \) Equivalent potential temperature.
\( \theta_v \) Virtual potential temperature.
\( \xi \) x-component of the vorticity vector.
\( \tilde{\omega} \) Vorticity vector of the perturbation flow.
\( B \) Buoyancy.
\( e'_{ij} \) Perturbation deformation tensor.
\( p' \) Perturbation pressure.
\( \vec{S} \) Environmental vertical wind shear vector.
\( u \) Wind speed in the x-direction.
\( \bar{u} \) The difference between the density-weighted mean wind speed over the lowest 6 km of the profile and the average wind speed over the lowest 500 m.
\( v \) Wind speed in the y-direction.
\( \vec{V} \) Wind vector.
\( w \) Vertical velocity.
\( w' \) Perturbation vertical velocity.
\( z \) Height.
Chapter 1

Introduction

1.1 Event Overview

On 24 August 2016, a tornado outbreak that produced 24 confirmed tornadoes impacted portions of Indiana and Ohio. This event is of interest because it caught many meteorologists by surprise; a tornado outbreak did not become apparent until after the first tornadoes already occurred. For example, the Storm Prediction Center (SPC) Convective Outlooks issued at 1300 (Fig. 1.1a) and 1630 UTC 24 August (Fig. 1.1c) confirm that this event was difficult to forecast only hours before the first supercell developed. The 1300 UTC Outlook did not include Indiana or Ohio in the Slight Risk area and the 1630 UTC Outlook, issued less than 3 hours before the first tornadoes occurred, did not include most of the area affected by the outbreak in the Slight Risk area. There was no tornado probability area issued at either time over the outbreak region (Figs. 1.1b,d). The preliminary tornado reports are displayed in Fig. 1.1b.

During the event, a leading line of elevated storms dissipated as it entered Indiana (Fig. 1.2a; Cluster 1), while a second line persisted and transitioned from disorganized elevated multicellular convection (Fig. 1.2a; Cluster 2) to three discrete supercells in Indiana after 1800 UTC (Fig. 1.2b), all of which produced significant tornadoes (rated EF-2 or greater on the Enhanced Fujita Scale). Additional supercells developed behind the initial supercells after 2000 UTC (Fig. 1.2b) and many of these storms produced tornadoes as well. The transition of the elevated multicellular convection into the first round of surface-based tornadic supercells is unusual and of particular interest. There are several studies on how surface-based storms become elevated (e.g., Parker 2008; Geerts et al. 2017), and how supercells grow upscale into multicelluar complexes (e.g., Bluestein and Weisman 2000; Finley et al. 2001), but the transition from elevated to surface-based convection is relatively unexplored as is the transition from multicellular to supercellular convection. Furthermore, most convection-allowing models failed to predict such a transition into supercells, instead depicting a line of storms moving across the affected area.
Figure 1.1: (a) SPC Day 1 Convective Outlook and (b) probability of a tornado within 25 miles of a point issued at 1300 UTC; (c) and (d) as in (a) and (b) but issued at 1630 UTC. Light green areas in (a) and (c) indicate regions where a 10% or greater probability of non-severe thunderstorms is forecast. Red dots in (b) are preliminary tornado reports. All products are valid from issuance time through 1200 UTC 25 August.

Figure 1.2: Radar reflectivity (dBZ) mosaic at (a) 1625 and (b) 2055 UTC. The two clusters discussed in the text are labeled in (a); the three primary supercells and secondary round of supercells are labeled in (b).
1.2 Literature Review and Motivation

There are three primary modes of convective organization: ordinary or single-celled convection, multicellular convection, and supercellular convection. Much research has been done to determine the environmental parameters that are the most important in determining convective mode. Weisman and Klemp (1982) used a three-dimensional cloud model (Klemp and Wilhelmson 1978) to simulate storms in environments with varying magnitudes of westerly vertical wind shear (i.e., straight hodographs) and buoyancy. They found that environments with a 0-6 km bulk-wind difference (BWD) less than 10 m s\(^{-1}\) yield single-cell storms that persist for around 60 minutes, moderately-sheared environments with a 0-6 km BWD around 15 m s\(^{-1}\) produce multicellular convection, and environments with a 0-6 km BWD between 25 and 45 m s\(^{-1}\) produce supercells. The supercells in their simulations split into right-moving and left-moving members owing to the straight hodographs. From simulations that varied both vertical wind shear and buoyancy, it was found that storms may not form at all if the shear is too high and the buoyancy is too low. Though, when storms do initiate, larger buoyancy creates more intense storms with stronger outflow, necessitating greater shear to restrain its propagation away from the updraft. From these results, they used a form of the bulk Richardson number,

\[
BRN = \frac{B}{\frac{1}{2}u^2}
\]  

(1.1)

where \(B\) is buoyancy (i.e., convective available potential energy; CAPE) and \(\bar{u}\) is the difference between the density-weighted mean wind speed over the lowest 6 km of the profile and the average wind speed over the lowest 500 m of the profile, to classify their environments. Supercells are generally favored in environments where \(BRN\) is between 15-35, and multicells are favored where \(BRN\) is greater than 40.

Weisman and Klemp (1984) conducted a similar set of simulations with varying wind shear, only with directionally-varying vertical wind profiles that yielded clockwise-curved hodographs. The hodographs were all semi-circles and varied in size. Again, supercells did not form if the vertical wind shear was too weak, regardless of the hodograph curvature. In the simulations with sufficient vertical wind shear for supercells, the right-moving supercell became dominant while the left-moving storm weakened. For a given clockwise-curved hodograph, more storm-relative helicity (SRH) exists for the right-moving supercell, allowing for the development of stronger rotation within the updraft (Davies-Jones 1984; Lilly 1986). Areas of rotation within storms yield areas of low pressure as seen through the pressure diagnostic equation (Rotunno and...
Klemp 1982; Klemp and Rotunno 1983):

\[ \nabla^2 p' = -\epsilon_{ij}^{\prime} + \frac{1}{2} |\vec{\omega}'|^2 - 2 \vec{S} \cdot \nabla_k w' + \frac{\partial B}{\partial z} \]  

(1.2)

In this equation, \( p' \) is the perturbation pressure, \( \epsilon_{ij}^{\prime} \) is the perturbation deformation tensor, \( \vec{\omega}' \) is the vorticity vector of the perturbation flow, \( \vec{S} \) is the environmental vertical wind shear vector, \( w' \) is the perturbation vertical velocity, and \( B \) is buoyancy. The first two terms on the right side of this equation are the non-linear dynamic terms. The first of these is the contribution from deformation, meaning that convergence is associated with positive pressure perturbations. The second non-linear term is the contribution from rotation, where rotation of either sign on the storm scale results in negative pressure perturbations. This term allows for supercells to split as upward-directed perturbation pressure gradient accelerations develop on both the left and right flanks of the storm (relative to the storm-motion vector) owing to the counter-rotating vortices that form via tilting of environmental horizontal vorticity (Fig. 1.3). If the hodograph is curved clockwise, however, the streamwise component of the horizontal vorticity increases (i.e., SRH increases; Davies-Jones 1984; Lily 1986) resulting in stronger rotation within right-moving storms, yielding stronger upward-directed non-linear perturbation pressure gradients where the rotation increases with height and additional amplification of the rotation via stretching. The third term is the linear dynamic term, yielding negative pressure perturbations downshear of updrafts and positive pressure perturbations upshear. In environments with unidirectional shear, perturbation highs and lows are stacked in the vertical, and thus no strong vertical perturbation pressure gradient develops (Fig. 1.4a). Given a clockwise turning hodograph, however, this term results in a perturbation high pressure near the surface and perturbation low pressure aloft on the right side of the storm, driving an upward-directed vertical perturbation pressure gradient there, aiding in updraft intensity and maintenance and resulting in updraft motion to the right of the mean winds (Fig. 1.4b). The last term is the pressure contribution from buoyancy arising from density differences (Doswell and Markowski 2004). Negative buoyancy perturbation pressures generally occur where buoyancy increases with height and vice versa for positive perturbations (e.g., Warren et al. 2017). To investigate the vertical accelerations that result from these pressure perturbations, it is useful to examine the vertical derivative of (1.2) given by:

\[ \nabla^2 \frac{\partial p'}{\partial z} = -\frac{\partial}{\partial z} \epsilon_{ij}^{\prime} + \frac{1}{2} \frac{\partial}{\partial z} |\vec{\omega}'|^2 - 2 \frac{\partial}{\partial z} (\vec{S} \cdot \nabla_k w') + \frac{\partial^2 B}{\partial z^2} \]  

(1.3)

Upward-directed accelerations arise when pressure perturbations described in conjunction with (1.2) decrease with height.
Figure 1.3: The promotion of storm splitting through the second non-linear term in (1.3) in the case of a straight hodograph. (a) A vortex couplet develops aloft via tilting of environmental horizontal vorticity owing to vertical wind shear, yielding upward-directed perturbation pressure gradients (blue shaded arrows) on the storm-relative left and right flanks of the updraft. (b) A precipitation-driven downdraft forms in the center of the original updraft, leading to storm splitting and the formation of two vortex couplets aloft. Vortex lines are in black and their associated rotation is given by the pink circular arrows, with the sign of vertical vorticity indicated. Brown arrows indicate storm-relative surface winds, and the storm-relative vertical wind profile is displayed on the left side of each panel. Transparent blue arrows indicate storm-relative trajectories, and in (b), the dashed transparent blue arrows indicate storm-relative trajectories after storm splitting and the development of deviant motion off the hodograph. From Markowski and Richardson (2010).
Figure 1.4: Pressure perturbations owing to the linear term in the pressure diagnostic equation surrounding an updraft in an environment with (a) a straight hodograph and (b) a hodograph that curves clockwise with height. Centers of high and low perturbation pressure are labeled, areas of rotation aloft are indicated, green arrows indicate horizontal perturbation pressure gradients, shaded blue arrows indicate vertical perturbation pressure gradients, transparent blue arrows indicate storm-relative trajectories, and the storm-relative vertical wind profile is displayed on the left side of each panel. From Markowski and Richardson (2010).
Broadly, current understanding of storm maintenance is that single-cell storms are usually relatively short-lived because they rain into their updrafts (Byers and Braham 1948). Supercells can persist much longer because strong storm-relative winds aloft, endemic to strongly-sheared environments, blow precipitation away from the updrafts, which also become more intense through storm-scale lift generated by vertical perturbation pressure gradients, as described above. Squall lines, multicellular storms, and mesoscale convective systems (MCSs) are maintained by the repeated initiation of new cells by lift generated along gust fronts, assuming that they are surface based (e.g., Rotunno et al. 1988). Elevated MCSs can be maintained by convergence and lift owing to nocturnal low-level jets (Coniglio et al. 2010; Shapiro et al. 2018), or perhaps by gravity waves or bores (Parker 2008; Geerts et al. 2017).

The Weisman and Klemp studies utilized an idealized model in which convection was initiated with a warm bubble; this initiation mechanism is generally not applicable to the real atmosphere because convection usually forms along atmospheric boundaries. Dial et al. (2010) considered convective initiation along drylines, prefrontal troughs, and cold fronts east of the Rocky Mountains from 2003 to 2007 to determine why storms initiating along boundaries sometimes form solid lines of convection and other times form broken lines of discrete supercells. They found that if the cloud-layer shear vector is oriented more normal to a boundary, discrete storms are more likely, and if the cloud-layer shear vector is oriented more parallel to a boundary, lines of storms are more likely (Fig. 1.5). In the parallel shear case (Fig. 1.5a), precipitation is blown downshear along the initiating boundary toward other storms. New convection develops between converging storm outflows as well as along the boundary itself and, as the outflows merge, new updrafts form in a linear fashion. In the perpendicular shear case (Fig. 1.5b), storms remain discrete because precipitation that is blown downshear from the updrafts does not impede on other updrafts. Storms also are more likely to move off of the initiating boundary in this case, removing themselves from any new storms that may initiate along the boundary.

Thompson et al. (2004a) developed the supercell composite parameter (SCP), using thresholds of parameters found to discriminate between environments supportive of supercells and those that are not.

\[
SCP = \left( \frac{MUCAPE}{1000 \ \text{J kg}^{-1}} \times \frac{EBWD}{20 \ \text{m s}^{-1}} \times \frac{ESRH}{50 \ \text{m}^2 \text{s}^{-2}} \right)
\]

In this equation, \( MUCAPE \) is the most unstable CAPE, \( EBWD \) is the effective bulk wind difference defined over the layer from the inflow base upward to 50% of the equilibrium level height of the most unstable parcel (Thompson et al. 2004c), and \( ESRH \) is the effective storm-relative helicity, defined over a continuous layer between the lowest level at which a parcel has greater than 100 J kg\(^{-1}\) CAPE and less than 250 J kg\(^{-1}\) CIN upward to where one of these constraints is no longer met (Thompson et al. 2004b). The \( EBWD \) term is
set to zero if $EBWD$ is less than 10 m s$^{-1}$ and it is set to 1 if $EBWD$ is greater than 20 m s$^{-1}$. An SCP of 2 is the best discriminator between environments favorable for supercells and those that are not, but this index does not consider the presence or mode of convection.

In reality, storm mode is likely a continuum and it is possible for storms to transition from one mode to another. A case study by Burgess and Curran (1985) gives an overview of a solid line of storms becoming a broken line of supercells. The authors mention that this evolution took forecasters by surprise, much like the tornado outbreak of 24 August 2016 did. On 26 April 1984, isolated storms initiated along a north-south oriented dryline in Oklahoma and were overtaken by a cold front around sunset, which resulted in a solid line of storms. Overnight, the line broke into discrete tornadic supercells. Their analysis indicates that this second transition was made possible by an increase in hodograph curvature after sunset owing to an approaching shortwave trough. This occurred in tandem with increasing warm air advection, allowing for the redevelopment of a capping inversion, such that only dominant updrafts within the line survived to realize the increased SRH. This event differs from 24 August 2016 because there was not a dryline or any frontal boundaries present during the event in question, but there was similarly an approaching shortwave trough in the form of a mesoscale convective vortex (MCV). Other than this case study, the author is unaware of any other in-depth studies of transitions from linear or multicellular convection to discrete supercells.

There have been several studies of the much more common transition from discrete supercells to lines of storms, however. For example, Finley et al. (2001) simulated a high-precipitation supercell transitioning into a bow echo. In that study, new convection initiated along the rear-flank gust front of the simulated supercell and merged with it, resulting in more precipitation and a stronger rear-flank cold pool, and thus a
faster moving rear-flank gust front. Convergence increased at the gust front and more convection initiated along it until the storms formed a bow echo. Bluestein and Weisman (2000) demonstrated how storms that form along a boundary collide with each other, and/or the outflows merge, yielding upscale growth into a line of storms depending on the orientation of the shear vector to the boundary along which the storms form (similar to Dial et al. 2010). They suggest that a shear vector oriented at 45° to the boundary is most favorable for limiting cell and outflow collisions that can lead to upscale growth. Coniglio et al. (2010) used Rapid Update Cycle (RUC) model analyses to create environmental composites for 94 MCSs, of which 39 were classified as rapidly-developing MCSs and 38 as slowly-developing MCSs. They found that in the rapid upscale growth cases, the initial storms often formed downstream of 500 hPa shortwaves, in the right-entrance region of 200 hPa jet streaks, and at the leading edge of warm air advection at 850 hPa, as these are all sources of quasi-geostrophic lift. The upscale growth usually began on the western edge of a low-level jet at 500 m above ground level (AGL), likely owing to increased convergence and moisture advection there. They also found that a stronger low-level jet, larger CAPE and downdraft CAPE, and weaker shear in the 3-10 km layer was more favorable for rapidly-developing MCSs.

The 2015 Plains Elevated Convection at Night (PECAN) field project (Geerts et al. 2017) gathered data on nocturnal MCSs over the Great Plains of the United States. Typically, surface-based storms formed during the late afternoon, grew upscale into MCSs, and then became elevated as the boundary layer stabilized after sunset. Many studies prior to and since this project investigated the transition from surface-based to elevated convection. For example, Parker (2008) utilized an idealized model to simulate the transition from surface-based storms to an elevated MCS. To accomplish this, artificial cooling was applied in the lowest 1 km of the model to simulate nocturnal cooling. Results indicate that up to 10 K of cooling may be needed for the MCS to become fully elevated. Also, as the temperature deficit decreases between the environment and the cold pool, lifting at the gust front may be enhanced as the horizontal vorticity owing to the low-level environmental vertical wind shear and that created by the buoyancy gradient across the gust front come into approximate balance (Rotunno et al. 1988). Even though PECAN sampled a few elevated MCSs that became surface based, this transition does not seem to be a topic of intense research stemming from this field project. Furthermore, there were no cases of multicellular convection transitioning into discrete supercells during the project as occurred on 24 August 2016.

On 24 August 2016, initially disorganized elevated multicellular convection transitioned to discrete surface-based tornadic supercells. Utilizing the WRF model, such a transition is simulated and the analysis presented herein aims to explain this unusual occurrence. Our hypothesis is that when the storms were elevated, they were relatively unorganized due to little shear above the weakly-stable boundary layer. Owing
to solar heating, storms became surface based and began to ingest strongly-sheared air in the lowest 1 km of the atmosphere, allowing for a transition to supercells. Slight cooling below 1 km AGL owing to weak convection ahead of the storms and anvil shading may have limited vertical mixing and enhanced the 0-1 km SRH (Markowski et al. 1998; Frame and Markowski 2010) in the inflow region of the storms. As rotation developed within the updrafts, non-linear dynamic pressure perturbations formed and allowed for vertical accelerations (1.3), aiding in the transition to surface-based convection. Chapter 2 contains an analysis of the synoptic and mesoscale environment and storm evolution on 24 August 2016. Chapter 3 presents the model configuration and methodology, including that for the trajectory analysis and perturbation pressure decomposition. Chapter 4 discusses the results, and Chapter 5 presents the conclusions.
Chapter 2

Observed Synoptic and Mesoscale Environment and Storm Evolution

On the evening of 23 August 2016, a complex of storms developed near the Nebraska/Iowa border ahead of a cold front. An MCV formed within this convection and continued to move eastward, while the storms dissipated over Iowa around 0900 UTC. By tracking the outflow boundary from this convection using a combination of radar and surface observations, an outflow speed of 14.5 m s\(^{-1}\) was calculated (Fig. 2.1). Ahead of the outflow boundary, widespread elevated convection developed over eastern Iowa and northwestern Illinois after 0700 UTC, likely in response to warm air advection at 850 hPa and differential positive vorticity advection (DPVA) owing to the approaching MCV (not shown). This convection organized into a line stretching from northern to central Illinois around 1130 UTC (Cluster 1, Fig. 2.2). Using the outflow speed calculated above, convection redeveloped within 20 km of the expected outflow location around 1400 UTC in northern Illinois (Cluster 2), yielding two lines of elevated storms (Fig. 2.2). This analysis, along with the shape of Cluster 2 on radar and visible satellite imagery, strongly suggests that this outflow boundary triggered the storms that eventually became tornadic supercells in Indiana.

Cluster 2 eventually devolved into three discrete supercells after it crossed into Indiana after 1800 UTC (Fig. 2.3), all of which produced significant tornadoes. The first supercell produced an EF-2 tornado near Crawfordsville, IN, at 1838 UTC as well as a EF-0 tornado near Indianapolis at 2018 UTC, (Crawfordsville supercell in Fig. 2.3, although was already outflow dominant at the time of the image; Fig. 2.4). The Crawfordsville tornado tracked 8.6 km with maximum estimated winds of 54 m s\(^{-1}\) and caused major damage to a few homes and barns. The second supercell to form produced an EF-3 tornado near Kokomo, IN, at 1920 UTC (Kokomo supercell in Figs. 2.3, 2.4). This tornado tracked 13.9 km with maximum estimated winds of 68 m s\(^{-1}\) and resulted in 20 reported injuries as it moved through Kokomo. At least 80 homes were destroyed and many businesses were damaged. The supercell that produced this tornado went on to produce eight additional tornadoes in far eastern Indiana and northwestern Ohio, west of Lima (Fig. 2.4). The third supercell produced an EF-3 tornado near Woodburn, IN, at 2127 UTC (Woodburn supercell in Figs. 2.3, 2.4). The Woodburn tornado tracked 8.3 km with maximum estimated winds of 72 m s\(^{-1}\). Many homes and barns were damaged, including a home that was removed from its foundation and a
barn that was completely leveled. This tornado also threw a dump truck a distance over 400 m. The parent supercell produced four more tornadoes in northwestern Ohio. Lastly, a second round of supercells (Fig. 2.3) formed around 2000 UTC and also produced tornadoes in north-central Indiana (all tornado statistics are from Storm Data; NCDC 2016).

2.1 Synoptic Environment and Mesoanalysis

The 1200 UTC sounding from Lincoln, IL (Fig. 2.5), was launched into precipitation, as indicated by the nearly moist-adiabatic lapse rates and deep saturation. The wind profile depicts strong shear between the surface and 950 hPa, where wind speeds increase from 10 to 40 kts and veer from southeasterly to southwesterly, and only weak shear above 950 hPa, with flow generally from the southwest between 30-40 kts. The 1200 UTC 925 hPa analysis indicates that 40 kt winds extended from Missouri into Illinois, northern Indiana, and Michigan (Fig. 2.6). The 1200 UTC 500 hPa analysis displays wind speeds between 40-50 kts on the south side of the MCV, the presence of which is inferred from the slight troughing in the isolines and relatively warmer temperatures near the Illinois/Iowa border (Fig. 2.7a).

By 1800 UTC, just before the first tornadoes occurred, SPC mesoanalyses showcase an environment favorable for the development of supercells and tornadoes. At 500 hPa, the MCV had moved into northern
Figure 2.2: Radar reflectivity (dBZ) from the Davenport, IA (KDVN), WSR-88D at 1407 UTC. Pink lines indicate the locations of the outflow boundary at 0406, 0712, and 1407 UTC. The dashed red line is the eastern extent of the outflow boundary at 0406 UTC and the solid red lines indicate the distance traveled by the outflow boundary, given in the blue text. The position of the outflow boundary at 1407 UTC was determined from a combination of radar, satellite, and surface observations. The expected distances traveled using the calculated outflow speed closely match the actual distances (text box at the top). Cluster 1 and Cluster 2 are labeled with black text.
Illinois and wind speeds greater than 40 kts existed on the southeast side of the MCV near the Illinois/Indiana border (Fig. 2.7b). The 0-1 km SRH was greater than 100 m$^2$ s$^{-2}$ over most of Indiana (Fig. 2.8a), favorable for the development of rotating updrafts. A pocket of 0-6 km bulk shear between 35-40 kts, owing to the stronger winds at 500 hPa on the south flank of the MCV, was centered near the Illinois/Indiana border (Fig. 2.8b). Plentiful instability (greater than 1000 J kg$^{-1}$ mixed-layer CAPE; MLCAPE) with little capping (less than 25 J kg$^{-1}$ mixed-layer convective inhibition; MLCIN) was present (Fig. 2.8c) with lifting condensation level (LCL) heights less than 750 m AGL providing a favorable environment for tornadoes (Fig. 2.8d).

After sunrise, visible satellite imagery was used in conjunction with surface observations to track several mesoscale boundaries important to this event. Figures 2.9 and 2.10 are examples of visible satellite and surface observations at roughly 1400 UTC with subjectively-analyzed boundaries superposed. The outflow boundary from the evening storms in Nebraska that was tracked through Iowa overnight can be identified by Cluster 2 on visible satellite (Fig. 2.9) and also by south-southwesterly surface winds ahead of the boundary and slower, more westerly winds behind it (Fig. 2.10). A differential heating boundary is marked by the southern edge of the cloud cover associated with Cluster 1 on visible satellite and a shift from faster
Figure 2.4: Tornado tracks from the 24 August 2016 outbreak. Colors indicate Enhanced Fujita scale rating. The towns nearest the three significant tornadoes in Indiana are labeled with bold text.
Figure 2.5: 1200 UTC sounding launched from Lincoln, IL. Winds are plotted in knots, with a flag = 50 kts, a barb = 10 kts, and a half barb = 5 kts.

Figure 2.6: 925 hPa analysis with height (solid contour; m), temperature (dashed red contour; °C), dewpoint (green contour; light green shading > 14°C, dark green shading > 18°C), and wind (barbs; kts) at 1200 UTC.
Figure 2.7: 500 hPa analysis with height (solid contour; m), temperature (dashed red contour; °C), and wind (barbs; > 40 kts shaded blue) at (a) 1200 and (b) 1800 UTC.

Figure 2.8: 1800 UTC analysis of (a) 0-1 km SRH (blue contours; m² s⁻²) and storm motion of a right-moving supercell (barbs; kts), (b) 0-6 km bulk shear (barbs; kts; > 30 kts contoured every 10 kts), (c) 100 mb mixed-layer CAPE (red contours, J kg⁻¹), MLCIN (> 25 J kg⁻¹ shaded) and surface winds (barbs; kts), and (d) LCL height (m AGL).
south-southwesterly surface winds outside of the cloud cover to slower, southerly winds and slightly cooler temperatures beneath the clouds. These boundaries meet at a triple point over central Illinois. A secondary wind shift is also apparent over eastern Iowa, indicated by generally southerly winds ahead of the boundary and westerly winds behind it (Fig. 2.10); this boundary appears to be the trigger for the second round of supercells that formed over Indiana during the afternoon of 24 August. The outflow boundary moved southeastward, the differential heating boundary moved northeastward as Cluster 1 dissipated, and the secondary wind shift moved eastward. The evolution of Cluster 2 into tornadic supercells is presented in the following section.

2.2 Radar Observations and Storm Evolution

From 1600 to 1630 UTC, additional small cells developed south of Cluster 2, which was a broken line of cells located near the Illinois/Indiana border at this time (Fig. 2.11a). There were no sharp gust fronts or surging outflow in surface observations (Fig. 2.10) or WSR-88D radial velocity data associated with either cluster of
storms (not shown). Convection continued to develop south of Cluster 2 after 1630 UTC, forming a small solid line near Danville, IL, by 1656 UTC (Fig. 2.11b), while Cluster 1 had nearly dissipated in Indiana. Clearing was also evident on visible satellite imagery over western Indiana in the wake of dissipating Cluster 1 (Fig. 2.12). After 1700 UTC, the differential heating boundary began to advance northeastward while the MCV moved eastward across northern Illinois (Fig. 2.12).

The southern end of Cluster 2 continued to exhibit multicellular structure from 1700 to 1730 UTC (Fig. 2.13a) with weak rotation developing at roughly 3500 m AGL within each cell as it moved northeastward within the cluster (Fig. 2.13b). From 1730-1800 UTC, the southernmost storm in Cluster 2 strengthened and developed rotation between 1200 - 3000 m AGL. As this rotation developed, Cluster 2 split into a southern group of cells west of Crawfordsville, IN (Cluster 2A), and a northern line west of Kokomo, IN (Cluster 2B; Fig. 2.13c). Small cells continued to develop to the south of and merge with Cluster 2A, and by 1800 UTC, there was one dominant area of rotation within Cluster 2A as small cells continued to merge into its southern end (Fig. 2.13c,d). At 1830 UTC, there was rotation above low level convergence on the southern end of the line and the storm began to acquire supercell characteristics, such as a mesocyclone, deviant rightward motion, and a hook echo (Crawfordsville supercell; Fig. 2.14). Small cells continued to
Figure 2.11: Radar reflectivity (dBZ) from the Lincoln, IL (KILX), WSR-88D at (a) 1628 and (b) 1656 UTC. The approximate domain is indicated in Fig. 2.3.

Figure 2.12: Visible satellite at 1655 UTC. Colored lines are as in Fig. 2.9, and the pink circle indicates the MCV location.
merge with this storm from the south, and following one such merger at 1832 UTC, rotation increased and the Crawfordsville supercell produced an EF-2 tornado from 1838 to 1848 UTC near Crawfordsville, IN, (Fig. 2.4). This storm continued moving to the east-southeast, passing just north of Indianapolis where it produced another brief tornado (Fig. 2.4).

As the Crawfordsville supercell produced its first tornado, a storm on the southern end of Cluster 2B began to acquire rotation between 1000 - 2000 m AGL and developed a hook echo west of Kokomo, IN (Kokomo supercell; Fig. 2.15). During the formation of the Kokomo supercell, the northeastern portion of Cluster 2B separated from the Kokomo supercell just before 1850 UTC (Cluster 2C; Fig. 2.15). At 1911 UTC, a small cell west of the Kokomo supercell merged with it and rotation increased at 2000 m AGL (Fig 2.16). The Kokomo supercell produced an EF-3 tornado near Kokomo from 1920 to 1934 UTC (Fig. 2.4).
Figure 2.14: (a) Radar reflectivity (dBZ) and (b) radial velocity (kts) from the Indianapolis, IN (KIND), WSR-88D at 1841 UTC depicting the Crawfordsville supercell. The radar site is southeast of the storm and the domain is indicated in Fig. 2.13d.

Figure 2.15: Radar reflectivity (dBZ) from the Indianapolis, IN (KIND), WSR-88D at 1850 UTC. Cluster 2C is circled and the domain is indicated in Fig. 2.13d.
Cluster 2C generally continued east-northeastward and was in range of the North Webster, IN (KIWX), WSR-88D radar by 2000 UTC. Stronger convection formed on the southern edge of Cluster 2C in an area of low-level convergence by 2041 UTC (Fig. 2.17). Rotation developed, increased, and a hook echo formed at 2100 UTC (Woodburn supercell). The hook echo became well defined (Fig. 2.18) as rotation deepened and increased further, and the supercell produced an EF-3 tornado near Woodburn, IN, from 2127 to 2139 UTC (Fig. 2.4). West of these three supercells, additional cells formed around 2000 UTC near of the secondary wind shift (Fig. 2.3). Many of these cells became supercells and produced tornadoes as well.

Figure 2.19 suggests that the development of each supercell coincided with the northeastward movement of the differential heating boundary as each tornado occurred shortly after the parent storm acquired supercellular characteristics. Along the boundary, the presence of baroclinically-generated horizontal vorticity likely enhanced SRH and the ingestion of this helicity-rich air by the updrafts may have been the impetus for supercell development and tornadogenesis. It is possible that numerical weather prediction models would have needed to capture this differential heating boundary to accurately depict supercells across this area.

Updraft helicity (UH) is commonly used to identify rotating updrafts in high-resolution numerical weather prediction output. Since a supercell mesocyclone is a rotating updraft, a streak of continuous UH accumulated over time is typically indicative of a simulated supercell, as long as a cellular mode is maintained within
Figure 2.17: (a) Radar reflectivity (dBZ) and (b) radial velocity (1.0° scan; kts) from the North Webster, IN (KIWX), WSR-88D at 2041 UTC. Circle in (b) indicates an area of convergence at roughly 900 m AGL. The radar is northwest of the area and the approximate domain is indicated in Fig. 2.3.

Figure 2.18: (a) Radar reflectivity (dBZ) and (b) radial velocity (kts) from the North Webster, IN (KIWX), WSR-88D at 2127 UTC depicting the Woodburn supercell. The radar is northwest of the storm and the approximate domain is indicated in Fig. 2.3.
Figure 2.19: As in Fig. 2.4 but with only the first tornado produced by each of the three initial supercells indicated (tornado start times are in black text boxes). Surface observations and visible satellite imagery were used to locate the differential heating boundary (solid red lines) at 1807 and 1907 UTC. The dashed red line is an estimate of the location of the differential heating boundary at 2107 UTC using only surface observations because widespread cloud cover at this time precluded the use of visible satellite imagery.
UH is given by

\[ UH = \int_{z_1}^{z_2} w \zeta dz \]  

where \( w \) is vertical velocity, \( \zeta \) is relative vertical vorticity, and the integral is typically taken between \( z_1 = 2 \) km and \( z_2 = 5 \) km. The SPC Storm Scale Ensemble of Opportunity (Jirak et al. 2012), initialized at 1200 UTC on 24 August, produced only weak, short UH streaks over Indiana, suggesting that rotating updrafts were possible, but that a tornado outbreak did not appear likely from the lack of significant (> 100 m² s⁻²; Clark et al. 2013) UH streaks over Indiana (Fig. 2.20). The High Resolution Rapid Refresh (HRRR) model is initialized every hour, and each model run from the morning of 24 August failed to produce significant updraft helicity over Indiana (output from 1600 and 1700 UTC runs is shown in Fig. 2.21). None of these models produced any UH streaks over Ohio. The failure of the convection-allowing models to produce simulated supercells certainly contributed to the difficulty in forecasting this event.
Figure 2.21: Accumulated UH (m$^2$ s$^{-2}$) through 0300 UTC 25 August from the HRRR model initialized at (a) 1600 UTC and (b) 1700 UTC 24 August.
Chapter 3
Methodology

3.1 Model Configuration

The WRF model, version 3.8.1 (Skamarock et al. 2008) was used to simulate the 24 August 2016 storms. The outer domain measures $1695 \times 1695 \times 20$ km with 60 vertical levels. The horizontal resolution of the outer domain is 3 km, the vertical resolution is stretched from 50 m to 250 m below 2 km and is 377.5 m above 2 km, and the timestep is 3 seconds. An inner domain with dimensions of $682 \times 682 \times 20$ km and 60 vertical levels is centered over Illinois and Indiana with 1 km horizontal resolution, the same vertical resolution as the outer domain, and a timestep of 1 second (Fig. 3.1). Both domains utilize open boundary conditions.

The model was initialized with the 12-km resolution 0600 UTC 24 August 2016 North American Mesoscale (NAM) model analysis. Data were provided for the lateral boundary conditions of the outer domain every six hours from the 1200 and 1800 UTC 24 August NAM analyses, and the 0000 UTC 25 August NAM analysis. The inner domain was initialized at 1400 UTC and the simulation terminated at 0000 UTC.

The Milbrandt-Yau two-moment cloud microphysics parameterization was used for both domains (Milbrandt and Yau 2005). Within the Milbrandt-Yau microphysics, the break-up diameter of rain drops was increased from the default of 300 $\mu$m to 450 $\mu$m, which is more representative of convective clouds. For example, in a study that fitted microphysical observations of convective clouds to numerical model output, the break-up diameter used was 600 $\mu$m (Verlinde and Cotton 1993). The Rapid Radiative Transfer Model for General Circulation Models was employed for both longwave and shortwave radiation (Mlawer et al. 1997; Iacono et al. 2000). The Revised MM5 Monin-Obukhov surface-layer scheme (Jiménez et al. 2012) and the Unified Noah land-surface model were used (Livneh et al. 2011). The MYNN 2.5 level TKE scheme was utilized for the boundary layer (Nakanishi and Niino 2006) along with the 2D Smagorinsky first-order turbulence closure model (Xue et al. 2000). There was no convective parameterization used for either domain. Diffusive damping was active 3000 m from the model top to mitigate reflecting waves. The model configuration is summarized in Table 3.1.
Figure 3.1: Domains used in the WRF simulation. The outer domain is 1695×1695 km with a horizontal resolution of 3 km. The nested inner domain is 682×682 km with a horizontal resolution of 1 km. For both domains, there are 60 vertical levels, the vertical resolution is stretched from 50 m to 250 m below 2 km, and is 377.5 m above 2 km.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Outer Domain</th>
<th>Inner Domain</th>
</tr>
</thead>
<tbody>
<tr>
<td>Domain Size</td>
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<td>682 × 682 × 20 km</td>
</tr>
<tr>
<td>Horizontal resolution</td>
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<td>1 km</td>
</tr>
<tr>
<td>Vertical resolution below 2 km</td>
<td>50 m to 250 m</td>
<td>-</td>
</tr>
<tr>
<td>Vertical resolution above 2 km</td>
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<td>-</td>
</tr>
<tr>
<td>Time step</td>
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<td>1 s</td>
</tr>
<tr>
<td>Boundary conditions</td>
<td>open</td>
<td>-</td>
</tr>
<tr>
<td>Initialization time</td>
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<td>1400 UTC 24 August 2016</td>
</tr>
<tr>
<td>Cloud microphysics</td>
<td>Milbrandt-Yau 2-moment, 450 µm breakup diameter</td>
<td>-</td>
</tr>
<tr>
<td>Longwave and shortwave radiation</td>
<td>Rapid Radiative Transfer Model for GCMs</td>
<td>-</td>
</tr>
<tr>
<td>Surface layer</td>
<td>Revised MM5 Monin-Obukhov</td>
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<tr>
<td>Land-surface</td>
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<td>2D Smagorinsky first-order closure</td>
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</tr>
<tr>
<td>Damping</td>
<td>Diffusive damping, 3000 m from model top</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 3.1: Model physical and computational parameters. A dash for the inner domain indicates that the parameter is the same as that for the outer domain.
Analyses of the model output include the use of trajectories to determine if simulated storms are elevated or surface based. Trajectories were calculated using the Read/Interpolate/Plot (RIP) program from output saved every 2 minutes. The trajectory time step is 1 minute and data were linearly interpolated between each timestep (as recommended by the RIP User’s Guide Section 6a).

3.2 Perturbation Pressure Decomposition

A decomposition of perturbation pressure was completed following Rotunno and Klemp (1982) and Weisman and Rotunno (2000). To the author’s knowledge, this is the first study that performed a perturbation pressure decomposition from WRF model output. Perturbation pressure decomposition has generally been performed from Cloud Model 1 (CM1) output rather than from WRF output because CM1 uses flat vertical levels, whereas WRF uses semi-terrain following (eta) levels. Code exists to perform the perturbation pressure decomposition from CM1 output, so to avoid tampering with the working code, we interpolated the WRF output to prescribed flat vertical levels like CM1 output. We used 34 vertical levels for $w$ and 33 vertical levels for $u$, $v$, and all scalar values. The vertical grid for $w$ spanned from 0 - 19800 m above mean sea level (MSL) with a vertical resolution of 600 m. The grid for the other variables started at 300 m MSL, just above the elevation of the surface, and extended to 19500 m MSL with a vertical resolution of 600 m. All values of $w$ were set to zero at the bottom vertical level. A caveat with this methodology is that relatively flat terrain is required such that the new prescribed flat levels do no intersect the model surface. Vastly changing terrain may also create pressure perturbations owing to orographic effects not associated with the storm.

In previous studies utilizing perturbation pressure decomposition, CM1 is typically initialized with a horizontally homogeneous environment determined by an input sounding, which is used as the base state for the perturbation pressure decomposition. The WRF simulation analyzed herein, however, is not initialized with a horizontally homogeneous environment. Thus, the base state is defined by taking the horizontal average of the vertical profile over an 11×11 km grid from the inner domain that is centered 90 km south and 30 km west of the maximum UH location associated with the storm of interest (red dot in Fig. 3.2). These averaged base-state profiles were quality checked to assure that this grid does not include active convection or precipitation. The perturbation pressure decomposition code returns vertical accelerations owing to the total dynamic terms [the sum of the first three terms on the right side of (1.3)], the non-linear dynamic terms [the sum of the first two terms on the right side of (1.3)], the linear term [the third term on the right side of (1.3)], and the buoyancy term [the last term on the right side of (1.3)] after inverting the
Figure 3.2: MLCAPE (J kg$^{-1}$; shaded), 0-6 km BWD (kts; barbs), and the locations of the ten base-state profiles used in the sensitivity test (blue dots). Vertical accelerations from location A are displayed in Fig. 3.3 and those from location B are displayed in Fig. 3.4. The location of the base state used for the analysis in section 4.4 is labeled with a red dot.

Laplace operator on the left side of (1.3).

To determine the sensitivity of the vertical perturbation pressure gradient accelerations to the location of the base state, tests for were conducted by selecting two locations west, five locations southwest, and two locations south of a point 95 km southwest of the storm of interest, for a total of ten base-state profile locations (Fig. 3.2). Base-state locations were chosen to be in upwind locations free of clouds and precipitation and far enough away from the storm as to not be directly influenced by it.

Choosing a different base state in an inhomogeneous environment may alter the magnitude of the pressure perturbations, but the pattern should remain the same (Doswell and Markowski 2004). In this case, the base-state profiles were quite similar and produced nearly identical vertical accelerations. Two examples, chosen at random, of the accelerations produced owing to each group of terms in (1.3) at 1.5 km MSL at
1600 UTC are provided in Figs. 3.3 (base state located at point A in Fig. 3.2) and 3.4 (base state located at point B in Fig. 3.2). Figures 3.3a and 3.4a depict that the vertical acceleration owing to gradients in the buoyancy component of the perturbation pressure (APB) are nearly identical. Figures 3.3b and 3.4b suggest that the vertical acceleration owing to gradients in the linear dynamic component of the perturbation pressure (APDL) are only about 0.01 m s$^{-2}$, significantly weaker than the other components near the updraft region of the storm. Vertical perturbation pressure gradient accelerations are dominated by those owing to gradients in the non-linear dynamic component of the perturbation pressure (APDN; Figs. 3.3c and 3.4c), which account for nearly all of the total dynamic vertical accelerations, and are almost identical between the two base states (APTD; Figs. 3.3d and 3.4d). The dominance of APDN at low levels is consistent with the results of Coffer and Parker (2017), who used CM1 to examine vertical dynamic accelerations in tornadic and non-tornadic supercells.

The mean and standard deviation of each component of the 1.5 km vertical perturbation pressure gradient accelerations from the entire sample of base-state locations in Fig. 3.2 are provided at 1600 and 1700 UTC in Figs. 3.5 and 3.6, respectively. The means of APB, APDL, and APDN at 1600 UTC (Figs. 3.5a,c,e) are similar to the examples provided in Figs. 3.3a-c and 3.4a-c. The standard deviations of APB and APDN (Figs. 3.5b,f and 3.6b,f) are both an order of magnitude smaller than the mean accelerations (different color scales are used for means and standard deviations in Figs. 3.5 and 3.6), indicating small variation relative to the mean among the sample and little sensitivity to base-state location. The APDL has the largest relative standard deviation, up to 0.01 m s$^{-2}$ in an area with 0.01 m s$^{-2}$ vertical acceleration (Figs. 3.6c,d), but APDL is also approximately and order of magnitude less than the total acceleration, which is dominated by APDN (Figs. 3.3 and 3.4). The standard deviations of APDL and APDN are nearly identical (Figs. 3.5d,f and 3.6d,f) because the perturbation pressure decomposition code first calculates the total dynamic perturbation pressure, then calculates the linear dynamic perturbation pressure, and treats the non-linear dynamic perturbation pressure as a residual. Thus, if APDL deviates from the mean at a location, the APDN must also deviate from the mean at the same location by nearly the same magnitude. It is advised to choose a base-state location with a vertical wind shear profile representative of the near-storm environment because APDL depends on the environmental vertical wind shear and appears to be the most sensitive to base-state location.

Changes in bulk shear on the order of 10 kts or in MLCAPE on the order of 1000 J kg$^{-1}$ (in an environment with around 3000 J kg$^{-1}$ MLCAPE) have little effect on the results of the perturbation pressure decomposition. Having completed these sensitivity tests, we are confident that our perturbation pressure decomposition is suitable for further analysis in section 4.4.
Figure 3.3: Vertical accelerations (m s$^{-2}$; shaded) at 1.5 km MSL owing to vertical gradients in the (a) buoyancy (APB), (b) linear dynamic (APDL), (c) non-linear dynamic (APDN), and (d) total dynamic (APTD) components of the perturbation pressure. The 40 dBZ simulated reflectivity contour at 1 km is in black. The perturbation pressure decomposition was performed using the base state centered at location A in Fig. 3.2. All panels are at 1600 UTC.
Figure 3.4: As in Fig. 3.3 but the perturbation pressure decomposition was performed using the base state centered at location B in Fig. 3.2. All panels are at 1600 UTC.
Figure 3.5: The mean (a) APB, (c) APDL, and (e) APDN (m s$^{-2}$; shaded) and standard deviation (m s$^{-2}$; shaded) of (b) APB, (d) APDL, and (f) APDN at 1.5 km produced from the sample of ten base-state locations in Fig. 3.2. The 40 dBZ simulated reflectivity contour at 1 km is in black. All plots are at 1600 UTC.
Figure 3.6: As in Fig. 3.5 but at 1700 UTC.
Chapter 4

WRF Model Simulation

4.1 Overview of Simulation

The WRF model accurately simulates the mesoscale environment on 24 August 2016, capturing the MCV that moved across northern Illinois (compare Fig. 2.7a with Fig. 4.1a and Fig. 2.7b with Fig. 4.1b). Although the 500 hPa winds associated with the simulated MCV are slightly weaker than were observed, the location and movement of the MCV match observations fairly well. Simulated surface winds and temperatures also match observations at 1400 UTC; for example, the model depicts temperatures near 80°F in southern Illinois dropping to the lower 70s°F in northern Illinois and Indiana (Figs. 2.10, 4.2a). The simulation also captures the differential heating boundary near the central Illinois/Indiana border and associated wind shift from southwesterly winds in southern Illinois and Indiana to southerly over central Indiana north of the boundary (compare Figs. 2.10 and 4.2a near the differential heating boundary). The simulated outflow temperature beneath active convection at 1400 UTC is around 70°F (Fig. 4.2a in the region of divergent winds north of the differential heating boundary in central Illinois). Observations suggest a similar outflow temperature indicated by the 70-73°F observations in this region. The model produces slightly higher dewpoints than were observed, with widespread 74-75°F dewpoints over central and southern Illinois, while observed dewpoints ranged from 71-75°F (compare Figs. 2.10 and 4.2b). The simulated 70°F isodrosotherm also extends farther east into Indiana, where observed dewpoints are in the upper 60s°F.

The simulation does not capture the two distinct clusters of storms that moved across Illinois in the morning. Instead, the simulation produces one well-defined cluster and some scattered, leading convection (Fig. 4.3a), that eventually dissipates, as Cluster 1 does in the observations. Figure 4.2a indicates that there is not a strong cold pool associated with this cluster, with only a 5°F temperature deficit in the outflow relative to the ambient air. Surface observations at 1400 UTC indicate a temperature deficit of roughly 2°F (Fig. 2.10) with a temperature of 75°F in the precipitation-free environment at Decatur, IL (KDEC), and 73°F in active convection at Champaign, IL (KCMI). Furthermore, the simulated surface winds in the outflow are not surging outward away from the storm, consistent with the lack of a surging gust front.
Figure 4.1: 500 hPa height (contoured; m AGL) and winds (barbs; >40 kts shaded blue) at (a) 1200 UTC and (b) 1800 UTC from the outer domain. The black box in (b) is the inner domain.

Figure 4.2: (a) Surface temperature (shaded; °F), surface winds (barbs; kts), and differential heating boundary location (solid blue line) valid at 1400 UTC from the inner domain. The blue dashed line encompasses a region of rain-cooled air and cloud shading from small cells southeast of the main cluster. (b) Surface dewpoint temperature (shaded; °F) and surface winds (barbs; kts) at 1400 UTC. The black box in (b) is the zoomed area in Figs. 4.3, 4.6, 4.7, and 4.8.
associated with Cluster 2 in radar and surface observations and in the observed 1200 UTC KILX sounding (Fig. 2.5). In the simulation, there is about 4°F of cooling at the surface due to the leading convection (the weak cells southeast of the main cluster in Fig. 4.3a) augmenting the differential heating boundary southeast of the cluster (dashed and solid boundaries in Fig. 4.2a).

Several severe weather parameters at 1800 UTC, around the time that the observed storms started to become supercells, were closely simulated as well. The simulation captures an area of 0-1 km SRH greater than 100 m² s⁻² in eastern Illinois and northern Indiana (compare Fig. 2.8a and Fig. 4.4a) co-located with 0-6 km bulk shear on the order of 30 kts (compare Fig. 2.8b and Fig. 4.4b). The SRH is much higher in the simulation immediately downstream of the cluster, possibly because the relatively coarse 13-km resolution Rapid Refresh (RAP) model analysis is unable to capture storm-generated modifications to the near-storm environment (e.g., Parker 2014). The simulation produces MLCAPE greater than 3000 J kg⁻¹ over central Illinois (Fig. 4.4c), whereas the 1800 UTC mesoanalysis only exhibits around 2000 J kg⁻¹ MLCAPE in this area (Fig. 2.8c). This discrepancy is likely due, in part, to a slightly moister environment in the WRF simulation as discussed above. The differential heating boundary results in a sharp instability gradient with 3000 J kg⁻¹ of MLCAPE and little MLCIN south of the boundary to less than 1000 J kg⁻¹ of MLCAPE and greater than 25 J kg⁻¹ of MLCIN north of it (near the southern end of the storm cluster in Fig. 4.4c).

A time series of maximum relative vertical vorticity (ζ) at 1 km and mean vertical velocity (w) at 4 km for the southernmost storm in the cluster is provided as a reference in Fig. 4.5; this storm becomes a supercell in the simulation. From 1500-1600 UTC, there are several weak updrafts and a brief peak in 1 km ζ just after 1500 UTC associated with a shallow vortex along a gust front (not shown). Between 1600-1700 UTC, a dominant updraft steadily intensifies, with a corresponding increase in 1 km ζ. After 1700 UTC, the mesocyclone becomes disorganized as a central downdraft develops, likely owing to mesocyclone cycling. Shortly after 1800 UTC, a dominant updraft redevelops and rapidly intensifies, followed by a peak in 1 km ζ at 1900 UTC, after which time the storm becomes outflow dominant and weakens.

Figures 4.3, 4.6, 4.7, 4.8, 4.9, and 4.10 provide an overview of the evolution of convection within the inner domain from 1500 to 2100 UTC every 30 minutes. A multicellular cluster is located over east-central Illinois at 1500 UTC (Figs. 4.3a and 4.9a). The 0.1 kg m⁻² contour of total column integrated ice is used as a proxy for the extent of the anvil cloud, which extends east and southeast from the storms. As noted above, there are small scattered cells southeast of the main cluster with associated cloud cover and latent cooling keeping temperatures roughly 2-3°F lower there than in the ambient environment south of the cluster (Fig. 4.3), resulting in a circuitous differential heating boundary southeast of the primary cluster (Fig. 4.2a). A surface wind shift is present along the primary differential heating boundary. Outside of the anvil cover, winds
Figure 4.3: Simulated 1 km reflectivity (dBZ; shaded), surface temperature (°C; red contours), total column integrated ice of 0.1 kg m$^{-2}$ (black contour), and surface winds (barbs; kts) valid at (a) 1500, (b) 1530, and (c) 1600 UTC. The light gray line is the Illinois/Indiana state line for reference. The blue boxes are the zoomed areas at the corresponding times in Fig. 4.9.
Figure 4.4: (a) 0-1 km storm-relative helicity \( (m^2 \text{s}^{-2}) \); (b) 0-6 km bulk wind difference (kts; > 30 kts shaded); (c) 500 m mixed layer CAPE (color fill; J kg\(^{-1}\)), MLCIN > 25 J kg\(^{-1}\) (hatching) from the inner domain at 1800 UTC. The 40 dBZ simulated reflectivity contour at 1 km (black) is overlaid on all panels for reference.
Figure 4.5: Time series of maximum 1 km $\zeta$ every 2 minutes (red) and mean 4 km $w$ (black) calculated over a 5×5 km grid centered on the maximum 1 km $\zeta$.

are southwesterly and faster, likely due to vertical mixing of stronger southwesterly momentum downward (Fig. 2.5). Under the anvil, vertical mixing is weaker owing to reduced solar heating, so surface winds are slower and more backed. As a result, 0-1 km SRH is enhanced beneath the anvil (Fig. 4.11; Frame and Markowski 2010, 2013; Parker 2014). A gradient in virtual potential temperature ($\theta_v$) is also co-located with the anvil edge and it is possible that horizontal vorticity is generated baroclinically along this gradient, augmenting the ambient 0-1 km horizontal vorticity.

Figures 4.9a,b demonstrate the multicellular structure at 1500 and 1530 UTC as many individual updrafts at 4 km are identifiable. A small area of relative vertical vorticity ($\zeta$) on the order of 0.01 s$^{-1}$ at 1 km is present on the right flank of a larger cell in the cluster at these times (Figs. 4.3a,b and 4.9a,b). This vorticity increases to 0.015 s$^{-1}$ and becomes located along a gust front from 1530 to 1600 UTC, indicated by a temperature gradient and wind shift southwest of the cluster (Figs. 4.3c and 4.9c). From 1600 to 1630 UTC, additional cells develop to the immediate south of and merge with the strengthening cell on the southwestern end of the cluster, while the differential heating boundary strengthens southeast of the cluster as solar heating increases temperatures outside of the anvil cover (Figs. 4.3c, 4.6a, and 4.9c,d; compare the relative locations of the 24°C and 28°C isotherms in Figs. 4.3a and 4.6a).

At 1700 UTC, the southwesternmost cell begins to turn right (move toward the southeast), $\zeta$ increases to 0.03 s$^{-1}$ in the notch of a hook echo, and $w$ at 4 km concentrates into one primary updraft above the 1 km vortex, indicative of a deep mesocyclone (Figs. 4.6b and 4.9e). By 1730 UTC, the hook echo is less defined
Figure 4.6: As in Fig. 4.3 only at (a) 1630, (b) 1700, and (c) 1730 UTC.
as the vortex and associated updraft move northward and dissipate as the mesocyclone cycles (Figs. 4.6c and 4.9f). Vortices continue to develop on the southern end of the storm through 1800 UTC as the updraft grows in size and strength in the simulated reflectivity notch (Figs. 4.7a and 4.9g), and 1 km $\zeta$ increases above 0.02 s$^{-1}$ by 1830 UTC (Fig. 4.9h). Another small cell develops west of the supercell at 1830 UTC (Fig. 4.7b) and merges with it at 1900 UTC, at which time a mesocyclone with concentrated $\zeta$ at 1 km up to 0.04 s$^{-1}$, a single strong updraft at 4 km, and a hook echo are present (Figs. 4.7c and 4.9i).

After 1900 UTC, the supercell weakens as it becomes outflow dominant. The simulated reflectivity decreases from 1900 to 2000 UTC (Figs. 4.7c and 4.8a,b), as does the vertical vorticity, while the updraft loses organization (Figs. 4.10a,b). The surface winds at 1930 and 2000 UTC suggest that the rear-flank gust front begins to surge southward during this period (stronger northerly surface winds west-southwest of the vertical vorticity maximum in Figs. 4.10a,b). The storm then exhibits a bowing structure as the surging outflow becomes more evident in the temperature field from 2030 to 2100 UTC (tightly packed isotherms south of the bowing cell in Figs. 4.8c,d). The updraft at 4 km also forms a bow structure as the 1 km vertical vorticity concentrates along the gust front (Figs. 4.10c,d). Similarly, the Crawfordsville supercell develops a bow and surging rear-flank gust front in radar observations around 2100 UTC (Fig. 2.3).

In summary, the storm exhibits a transition from unorganized (and elevated, as presented in section 4.2) convection to a surface-based supercell between 1500 and 1700 UTC in the simulation, as was observed from 1600 to 1830 UTC with the Crawfordsville supercell. A mesocyclone eventually develops within the southernmost storm as this cell turns right and exhibits a hook echo, indicative of a supercell. After 1900 UTC, the supercell weakens and becomes outflow dominant, as suggested by the bowing structure in reflectivity. A few cells farther north also exhibit rotation, as indicated by cumulative UH streaks $> 300$ m$^2$ s$^2$ (Fig. 4.12), which are analogous to the Kokomo and Woodburn supercells in the observations. Although the first simulated supercell develops earlier and thus farther west than was observed, we believe that this simulation depicts the events of 24 August 2016 well.

While numerous simulations of supercells within favorable kinematic and thermodynamic environments have been analyzed in the past (e.g., Klemp and Wilhelmson 1978; Weisman and Klemp 1982, 1984; Bluestein and Weisman 2000; Coffer and Parker 2015, 2017), the development of a supercell from initially disorganized elevated convection remains relatively unexplored. Thus, the remainder of the analysis focuses on the transition from disorganized elevated convection into an intense surface-based supercell.
Figure 4.7: As in Fig. 4.3 only at (a) 1800, (b) 1830, and (c) 1900 UTC.
Figure 4.8: As in Fig. 4.3 only at (a) 1930, (b) 2000, (c) 2030, and (d) 2100 UTC. The blue boxes are the zoomed areas at the corresponding times in Fig. 4.10.
Figure 4.9: Vertical velocity at 4 km (m s$^{-1}$; shaded), relative vertical vorticity at 1 km contoured every 0.005 s$^{-1}$ (green), surface winds (kts; barbs), and the 40 dBZ simulated reflectivity contour at 1 km (black) at (a) 1500, (b) 1530, (c) 1600, (d) 1630, (e) 1700, (f) 1730, (g) 1800, (h) 1830, and (i) 1900 UTC. The dark blue lines in (a-e) are the locations of the vertical cross sections at the corresponding times in Figs. 4.13, 4.16, and 4.17.
Figure 4.10: As in Fig. 4.9 but at (a) 1930, (b) 2000, (c) 2030, and (d) 2100 UTC.
0-1 km SRH, Virtual Potential Temperature, Simulated Reflectivity, and Total Column Integrated Ice

Figure 4.11: 0-1 km SRH (m² s⁻²; shaded), virtual potential temperature (K; thin black contours), 40 dBZ simulated reflectivity contour at 1 km (green), and total column integrated ice of 0.1 kg m⁻² (thick black contour) at 1600 UTC. The light gray line is the Illinois/Indiana state line for reference.
Figure 4.12: Simulated reflectivity at 1 km (dBZ; shaded) and cumulative UH streaks > 300 m$^2$ s$^{-2}$ (dark shading) at (a) 1835 UTC and (b) 2010 UTC. The red ovals in (b) indicate UH streaks from rotating cells north of the primary supercell discussed in the text.

4.2 Vertical Cross Sections

Vertical cross sections permit an examination of the depth of updrafts and mesocylones throughout their life cycles. The cross sections presented are taken either from south to north or west to east over a distance of 40 km through the area of greatest vorticity at 1 km. The vorticity equation is given by

$$\frac{\partial \zeta}{\partial t} = -\vec{V} \cdot \nabla \zeta + \xi \frac{\partial w}{\partial x} + \eta \frac{\partial w}{\partial y} + \zeta \frac{\partial w}{\partial z}$$  \hspace{1cm} (4.1)

where $\vec{V}$ is the wind vector, $\xi$ and $\eta$ are the $x$ and $y$-components of the vorticity vector, and all other terms have their standard meaning. The term on the left is the local tendency of vertical vorticity, the first term on the right is advection of vertical vorticity, the next two terms are tilting of horizontal vorticity into the vertical, and the last term is stretching of vertical vorticity. When vorticity becomes co-located with an updraft that strengthens with height, the stretching term can amplify vorticity exponentially with time.

At 1500 UTC, the south-to-north vertical cross section along the line in Fig. 4.9a indicates that there are multiple elevated updrafts rooted above 2 km MSL (Fig. 4.13a). Weak vertical vorticity of both signs straddle the updrafts above 2 km, owing to tilting of environmental horizontal vorticity. The presence of a vortex below 2 km suggests that the updrafts may be ingesting some near-surface air, but this vortex lasts for only a brief period before it moves northeastward into outflow (not shown). At 1530 UTC, a new vortex with vertical vorticity on the order of 0.01 s$^{-1}$ develops below 1 km (Fig. 4.13b) and persists for longer
than the previous vortex. Most vertical wind shear, and thus horizontal vorticity, is confined below 1 km (compare areas south of $y = 17$ km in Fig. 4.14 with Fig. 2.5; areas north of $y = 17$ km in Fig. 4.14 are either within convection or outflow and are thus not representative of the ambient environment). Although weak vertical vorticity develops in the elevated updrafts, it is expected that much stronger near-surface vertical vorticity would result as updrafts ingest more air from below 1 km and tilt the stronger horizontal vorticity generated by vertical wind shear in this layer. Indeed, Fig. 4.15a confirms that weak tilting is occurring at 0.5 km at 1530 UTC. A trajectory analysis (section 4.3) confirms that storms were beginning to become surface based just after this time.

Multicellular structure is still evident at 1550 UTC (Fig. 4.13c), but the updrafts are intensifying near and below 2 km MSL. The northernmost updraft still exhibits a vortex couplet straddling it, whereas cyclonic vertical vorticity is co-located with the next two updrafts to the south (at 4 km MSL near $y = 26$ km, and at 1 km MSL near $y = 23$ km in Fig. 4.13c). Ten minutes later, at 1600 UTC, an 8 km deep surface-based updraft develops that is co-located with vertical vorticity greater than 0.015 s$^{-1}$, extending from near the surface to over 2 km MSL (Fig. 4.13d). The ingestion of air with high 0-1 km SRH (Fig. 4.11) is consistent with the development of a dominant cyclonic vortex within an updraft as opposed to a vortex couplet straddling it (Davies-Jones 1984). Figure 4.15b illustrates that both tilting and stretching are active near the primary updraft and along the rear-flank gust front at 1600 UTC.

By 1620 UTC, multicellular structure is still evident, but the individual updrafts are larger, deeper, and stronger, with maximum updraft speeds greater than 20 m s$^{-1}$ (Fig. 4.13e). As the updrafts move northward and become elevated atop outflow, they retain the vorticity acquired from ingesting the strongly-sheared near-surface air. These updrafts merge into one deep, rotating updraft at 1630 UTC (Fig. 4.16a), extending to roughly 16 km MSL and exhibiting vertical vorticity greater than 0.015 s$^{-1}$ from 1-6 km MSL. Meanwhile, tilting and stretching increase southwest of the updraft, near the rear-flank gust front, as a new updraft develops there (Fig. 4.15c).

This new updraft quickly acquires vorticity greater than 0.015 s$^{-1}$ in the lowest 1 km by 1640 UTC while the deep updraft evident at 1630 UTC moves northward, becomes elevated atop outflow, and weakens (Fig. 4.16b). The new updraft is more upright than the previous updrafts (compare Fig. 4.16a and Fig. 4.16c). By 1650 UTC, this updraft extends to roughly 12 km MSL with a mesocyclone roughly 8 km in diameter and nearly 10 km deep (Fig. 4.16c). At 1700 UTC, the updraft tilts toward the east (Figs. 4.16d and 4.17). Despite the tilted updraft, vorticity greater than 0.03 s$^{-1}$ is present at 1 km MSL. A large area of positive tilting exists on the flanks of the updraft near the hook echo (Fig. 4.15d) as well as a large, concentrated circular area of stretching beneath the updraft (compare the size and shape of stretching
Figure 4.13: South-to-north vertical cross sections of vertical velocity (m s\(^{-1}\); shaded), vertical vorticity (s\(^{-1}\); green; positive values solid and negative values dashed) contoured every 0.005 s\(^{-1}\) (zero contour omitted for clarity), and plane-parallel wind vectors (m s\(^{-1}\); arrows) at (a) 1500, (b) 1530, (c) 1550, (d) 1600, and (e) 1620 UTC. Units on the axes are km.
This analysis indicates that the cluster of elevated storms begins to ingest air from below 1 km between 1500-1530 UTC as tilting generates vertical vorticity below 1 km. New updrafts become co-located with vertical vorticity and intensify from 1550 to 1620 UTC. By 1630 UTC, one deep rotating updraft exists as a new updraft forms to its south. This new updraft quickly grows into another deep rotating updraft from 1640 to 1700 UTC with vertical vorticity at 1 km exceeding 0.03 s\(^{-1}\). The transition from unorganized, elevated convection to a surface-based supercell is not instantaneous in this case, but rather requires roughly 90 minutes and multiple updrafts before one deep rotating updraft with strong rotation below 1 km develops, similar to what often occurs with the initiation of surface-based storms in convection-free environments. In the next section, we utilize trajectory analyses to confirm that the storms indeed transitioned from elevated to surface based prior to the development of supercellular characteristics.

### 4.3 Trajectory Analysis

A west-to-east oriented line of 100 trajectories was initiated south of the storm of interest every 1 km at 250 m MSL, and was integrated forward in time to examine if any near-surface parcels enter an updraft. If the
Figure 4.15: Tilting of horizontal vorticity ($s^{-2}$) at 0.5 km (shaded), stretching of vertical vorticity at 0.5 km contoured every 0.0001 $s^{-2}$ (green), and the 40 dBZ reflectivity contour at 1 km (black) at (a) 1530, (b) 1600, (c) 1630, and (d) 1700 UTC. The blue line is the location of the vertical cross section at corresponding times in Fig. 4.13 or 4.16.
Figure 4.16: As in Fig. 4.13 but at (a) 1630, (b) 1640, (c) 1650, and (d) 1700 UTC.
Figure 4.17: As in Fig. 4.13d but for a west-to-east cross section at 1700 UTC.

Updrafts are elevated, these trajectories should not enter them. Figures 4.18a,b indicate that trajectories initiated at 1500 UTC bound for the updraft region of the cluster at 1522 UTC are forced upward to a height of 1-2 km by a gust front, and then descend. A few parcels eventually enter weaker updrafts farther north after ascending over the gust front. These trajectories, however, are not indicative of near-surface air directly entering deep updrafts. Trajectories initiated at 1610 UTC bound for the updraft region at 1630 UTC are ingested by a deep updraft (Figs. 4.18c,d) and rotation is evident in the westernmost trajectory paths (Fig. 4.18d), lending confidence that the transition from elevated to surface-based convection occurs between 1520 and 1630 UTC, and permits our analysis to be focused between these times.

Backward trajectories were utilized to determine at what level parcels entering the updrafts originated. We hypothesize that, initially, trajectories originate at some level above the surface, likely above 1 km, and that this level approaches the surface over time. A 3×3 km grid of trajectories, each separated by 1 km, was initialized, centered on the location and height of the strongest updraft and the trajectories were then integrated backward in time to 1400 UTC, when the inner domain was initialized. A robustness check was conducted by initializing a 5×5×4 km box of backward trajectories also centered on the strongest updraft; the 3×3 km grid of trajectories shown herein is representative of this larger set. Figure 4.19 demonstrates that at 1530 UTC, the parcels that enter the strongest updraft all originate above 1 km MSL. Figure 4.20a gives a horizontal projection of the trajectories. Almost all the trajectories come from the same area, west-southwest of the storm at 1530 UTC, consistent with the ambient 925 hPa flow regime (Fig. 2.6). A model sounding from the starred location in Fig. 4.20a at 1510 UTC, while the parcels are at this location and 20
minutes before they are within the maximum updraft, is displayed in Fig. 4.21a. Although the sounding indicates that the level of free convection (LFC) for an undilute surface parcel is below 1 km, such a parcel does not become significantly buoyant until it ascends above 850 hPa, or about 1.5 km AGL, owing to the poor lapse rates between 925-850 hPa. Parcels lifted from around 850 hPa, however, do not encounter such a neutrally-buoyant layer, explaining why the storms are elevated at 1530 UTC and why the source level of the trajectories at this time is around 1.5 km (Fig. 4.19). Since most of the vertical wind shear is below 1 km, similar to what was observed (compare Figs. 2.5, 4.14, and 4.21a), these elevated storms do not acquire significant rotation via tilting of environmental horizontal vorticity (Figs. 4.13a,b, 4.15a).

At 1540 UTC, the back trajectories indicate that some parcels originating below 1 km are being ingested into the updrafts (Fig. 4.22a). Back trajectories at 1542, 1544, and 1546 UTC (Figs. 4.22b-d), however, indicate that the majority of parcels entering the updrafts during this period are still from above 1 km. By 1548 and 1550 UTC (Figs. 4.22e,f), some parcels from below 1 km are again feeding into the updrafts, implying that the convection is beginning to become rooted closer to the surface, although parcels from many levels still enter the updrafts. At 1550 UTC (Fig. 4.20b), the parcels from below 1 km enter the updrafts from the southwest, while parcels from above 1 km flow from the west-southwest, owing to the veering winds with height. A model sounding at 1530 UTC from the location shown in Fig. 4.20b indicates that near-surface parcels exhibit more positive buoyancy just above their LFC than 20 minutes earlier (compare Fig. 4.21a and Fig. 4.21b) owing to more diabatic solar heating, supportive of more near-surface air entering the updrafts.

Even though some of the trajectories originate below 1 km at 1550 UTC (Fig. 4.22f), not all of the trajectories do so. Figures 4.23a,b,c all indicate that there are some trajectories from below 1 km and some from above 1 km entering the updrafts between 1552 and 1556 UTC; it is not until 1558 UTC (Fig. 4.23d) that almost all of the trajectories originate below 1 km; the convection is continually surface-based thereafter (not shown) suggesting that the transition from elevated to surface-based convection is not instantaneous. This analysis is consistent with that presented in section 4.2, which depicts vertical vorticity of 0.015 s$^{-1}$ developing below 2 km by 1550 UTC (Fig. 4.13c), and a deeper area of 0.015 s$^{-1}$ vertical vorticity co-located with an updraft greater than 20 m s$^{-1}$ at 1600 UTC (Fig. 4.13d), when the storm is surface based. The vertical vorticity could then be amplified through stretching (Figs. 4.15b-d). Figure 4.24 indicates that parcels entering the 1 km updraft at 1600 UTC originate from a broad area with 0-1 km SRH between 150-200 m$^2$ s$^{-2}$ and pass through an area of enhanced 0-1 km SRH $> 500$ m$^2$ s$^{-2}$ beneath the anvil southeast of the storm. It is possible that horizontal vorticity generated baroclinically along the temperature gradient at the edge of the anvil contributes to this enhanced SRH (Fig. 4.11).
Figure 4.18: A horizontal projection of forward trajectories initiated along a west-to-east line south of the storm cluster and bound for the updraft region at (a) 1522 and (c) 1630 UTC. $w > 3 \text{ m s}^{-1}$ at 1 km is shaded red and $w < -3 \text{ m s}^{-1}$ is shaded blue. The 40 dBZ simulated reflectivity contour at 1 km is green and the black line is the Illinois/Indiana border. Projections of the trajectories on a west-to-east vertical cross section are provided at the corresponding times in (b) and (d). Only trajectories that intersect the updraft are shown.
Figure 4.19: A south-to-north vertical projection of the nine trajectories centered on the strongest updraft at 1530 UTC.

Figure 4.20: As in Figs. 4.18a,c except for the nine back trajectories centered on the strongest updraft at (a) 1530 and (b) 1550 UTC. The star in (a) is the location of the model sounding taken at 1510 UTC in Fig. 4.21a and the star in (b) is the location of the model sounding taken at 1530 UTC in Fig. 4.21b. The black line is the Illinois/Indiana border.
Figure 4.21: Model soundings taken at (a) 1510 UTC from the location indicated in Fig. 4.20a and (b) 1530 UTC from the location indicated in Fig. 4.20b. The dashed white line is the most unstable parcel process curve and the dashed red curve is the environmental virtual temperature. Although the LFC for an undilute surface parcel in (a) is below 1 km (yellow arrow), such a parcel does not become significantly buoyant until above 1 km (blue arrow). In (b), a parcel is more buoyant just above its LFC (yellow arrow). Soundings were generated using SHARPpy (Blumberg et al. 2017).
Figure 4.22: As in Fig. 4.19, but for trajectories centered on the strongest updraft at (a) 1540, (b) 1542, (c) 1544, (d) 1546, (e) 1548, and (f) 1550 UTC.
Figure 4.23: As in Fig. 4.19 but for trajectories centered on the strongest updraft at (a) 1552, (b) 1554, (c) 1556, and (d) 1558 UTC.
0-1 km SRH, Virtual Potential Temperature, Simulated Reflectivity, and Total Column Integrated Ice

Figure 4.24: 0-1 km SRH (m$^2$ s$^{-2}$; shaded), 40 dBZ simulated reflectivity contour at 1 km (green), and total column integrated ice of 0.1 kg m$^{-2}$ (thick black contour) at 1600 UTC. Thin black lines and arrows are trajectories bound for the 1 km MSL updraft at 1600 UTC. The light gray line is the Illinois/Indiana state line for reference.
If enough strongly-sheared air from below 1 km enters the updraft before the storm is entirely surface-based, the transition to surface-based convection could be expedited owing to the creation of upward-directed vertical perturbation pressure gradients stemming from rotation aloft generated through tilting and amplified via subsequent stretching of vorticity [second non-linear term in (1.3)]. Any such perturbation pressure gradient that may have developed likely aided near-surface parcels in overcoming the moist-neutral layer between 850-700 hPa. The development of vertical perturbation pressure gradients is analyzed in the following section.

4.4 Perturbation Pressure Analysis

In section 4.3, we demonstrated that updrafts become surface based between 1545-1600 UTC, coinciding with diurnal surface heating and that updraft and mesocyclone intensity increases after this occurs. In this section, we explore vertical accelerations owing to vertical perturbation pressure gradients during this transition period. Figures 3.3c and 3.4c depict that APDN dominates the other contributions to the vertical perturbation pressure gradients at 1600 UTC, meaning that if accelerations owing to vertical perturbation pressure gradients aid in the development of surface-based updrafts, those due to the APDN term are likely the most important. As near-surface air enters updrafts, the updrafts acquire rotation through tilting of the horizontal vorticity generated by the vertical wind shear in the 0-1 km layer. Rotation of either sense yields negative non-linear dynamic perturbation pressures [second term on the right side of (1.2)]. Rotation increasing with height then causes upward-directed perturbation pressure gradients and thus upward accelerations, possibly permitting near-surface air to ascend through a moist-neutral layer (e.g., Fig. 4.21a).

Positive (upward) APDN at 900 m is generally co-located with $\zeta$ at 2 km of either sign (areas of positive APDN are near the green and pink contours in Fig. 4.25). An updraft with $w > 5$ m s$^{-1}$ at 1 km is present near the intersection of the two orange lines from 1546-1550 UTC (Figs. 4.13c, 4.25a-c), but it does not persist (Fig. 4.25d). At 1552 UTC, a new updraft with $w > 5$ m s$^{-1}$ at 1 km develops in an area of upward APDN at 900 m near the intersection of the blue lines in Fig. 4.25d. This updraft strengthens from 1552-1600 UTC (Figs. 4.25d-h) and acquires $\zeta > 0.015$ s$^{-1}$ at 1 km (Fig. 4.13d). This is also likely the first updraft ingesting mostly near-surface air (Fig. 4.23).

The orange lines in Figs. 4.25a-c correspond to the south-to-north and west-to-east vertical cross sections in Figs. 4.26a,b, 4.27a,b, and 4.28a,b and are taken through the updraft that dissipates. Figs. 4.26a,b indicate that $\zeta$ is greater than 0.01 s$^{-1}$ near 1 km, but there is little $\zeta$ above this level, inducing negative APDN in the updraft from 1.5-5 km at 1546 UTC, which becomes more apparent at 1548 UTC (Figs. 4.27a,b). By
Figure 4.25: APDN at 900 m (m s$^{-2}$; shaded), vertical velocity at 1 km (contoured every 5 m s$^{-1}$; black), vertical vorticity at 2 km (contoured every 0.005 s$^{-1}$; positive values in green, negative values in pink, and zero contour suppressed for clarity), and 40 dBZ simulated reflectivity contour at 1 km (thin dark green line) at (a) 1546, (b) 1548, (c) 1550, (d) 1552, (e) 1554, (f) 1556, (g) 1558, and (h) 1600 UTC. Orange lines in a-c are the locations of the south-to-north and west-to-east vertical cross sections in Figs. 4.26a,b, 4.27a,b, and 4.28a,b. Blue lines are the locations of the vertical cross sections in Figs. 4.26a,b, 4.27a,b, 4.28a,b, 4.29, 4.30, and 4.31.
1546 UTC (Figs. 4.25a-c) at 1546 UTC.

1550 UTC (Figs. 4.28a,b), the negative APDN weakens the updraft in the 1.5-3 km layer to less than 5 m s\(^{-1}\) and cuts it off from the stronger updrafts aloft. This appears to be a failed surface-based convective initiation attempt.

During this same time period, a new updraft develops southwest of the dissipating updraft discussed above. Vertical cross sections near this new updraft are indicated by the blue lines in Figs. 4.25a-c. At 1546 UTC (Figs. 4.26c,d), before the updraft at 1 km is greater 5 m s\(^{-1}\), there is an elevated updraft above 3 km with \(\zeta > 0.01\) s\(^{-1}\), yielding positive APDN up to 0.04 m s\(^{-2}\) extending from near the surface to 4.5 km. The area of \(\zeta\) at 4.5 km increases at 1548 UTC (Figs. 4.27c,d), likely owing to stretching. Despite this, APDN decreases slightly, but remains positive through 5 km. The decrease in the magnitude of APDN is likely due to increased \(\zeta\) near the surface, decreasing the vertical gradient in \(\zeta\). A small updraft greater than 5 m s\(^{-1}\) develops at 1.5 km at 1548 UTC (Fig. 4.27d), which connects to the updraft aloft and extends through 8 km by 1550 UTC (Figs. 4.13c, 4.28c,d). APDN decreases slightly from 1548-1550 UTC, but remains positive from the surface to 5 km.

The blue lines in Figs. 4.25d-f are the planes of the vertical cross sections in Fig. 4.29. The updraft near
Figure 4.27: As in Fig. 4.26, but along the lines in Fig. 4.25b at 1548 UTC.

Figure 4.28: As in Fig. 4.26, but along the lines in Fig. 4.25c at 1550 UTC.
the intersection of the cross sections at 1552 UTC (Fig. 4.25d) continues to strengthen through 1600 UTC (Figs. 4.25e-h). The vertical cross sections at 1552 UTC (Figs. 4.29a,b) still exhibit $\zeta > 0.01 \text{ s}^{-1}$ at 5 km and an associated column of positive APDN from the surface to this level. At 1554 UTC (Figs. 4.29c,d), the area of APDN extends upward and to the east of the updraft at 1 km (Fig. 4.29d). At 1 km, the updraft widens at 1556 UTC (Figs. 4.29e,f), and $w$ in the updraft core increases above 10 m s$^{-1}$. As the near-surface updraft intensifies, $\zeta$ at 1 km becomes greater than 0.01 s$^{-1}$ and APDN in the surface-to-2 km layer increases to 0.06 m s$^{-2}$. The rotation creating non-linear dynamic pressure perturbations does not need to be oriented vertically (i.e., horizontal vorticity can also create negative pressure perturbations) and it can be inferred that horizontal vorticity is present on the flanks of the updraft (e.g., positive APDN near the southern flank of the updraft in Fig. 4.29e). Furthermore, non-linear dynamic pressure perturbations can also be caused by the deformation term in (1.2), and the presence of deformation can be inferred by convergence along the gust front creating positive pressure perturbations near the surface, driving upward accelerations (there is a gust front in the vicinity of the updraft; Figs. 4.3c, and 4.29e,f). Large positive values of APDN near the surface are conducive to near-surface air entering the updrafts, regardless of how the acceleration develops.

The vertical cross sections in Figs. 4.30 and 4.31 are taken along the blue lines in Figs. 4.25g,h. Near-surface $\zeta$ is greater than 0.015 s$^{-1}$ at 1600 UTC (Figs. 4.13d and 4.31), indicating that horizontal vorticity owing to the 0-1 km vertical wind shear has been tilted and stretched in the updraft (Fig. 4.15b) and that the updraft is surface-based (Fig. 4.23d). The vertical cross sections of APDN at 1558 (Figs. 4.30a,b) and 1600 UTC (Figs. 4.31a,b) indicate upward accelerations greater than 0.10 m s$^{-2}$ in the surface-to-3 km layer, which is over roughly the bottom half of the updraft. Above this layer, in the top half of the updraft, is negative APDN of similar magnitude. Large negative APDN is generally not indicative of updraft growth, but the core of the updraft intensifies from 15 m s$^{-1}$ to 20 m s$^{-1}$ from 1558 to 1600 UTC (Figs. 4.30a and 4.31a). For all the vertical cross sections presented in this section, APDL was also inspected and is much smaller than APDN, suggesting that the updraft growth is not primarily driven by APDL (e.g., Figs. 3.3b and 3.4b). APB, however, can be of similar magnitude to APDN (Figs. 4.30c,d and 4.31c,d) and positive APB is co-located with negative APDN (compare Figs. 4.30a,b with 4.30c,d and Figs.4.31a,b with 4.31c,d), offsetting the downward acceleration from APDN. This allows the updraft to persist and even deepen when $\zeta$ near the surface becomes stronger than $\zeta$ aloft, which drives downward APDN. The strengthening updraft can then further increase $\zeta$ through stretching.

This analysis suggests that the first surface-based updraft is at least partially a result of positive APDN below a rotating updraft around 5 km MSL. It takes roughly 15 minutes (from 1545-1600 UTC) for an updraft with $w > 20 \text{ m s}^{-1}$ extending from 1-9 km MSL to form (Fig. 4.31c). Once the updraft is established and
Figure 4.29: Vertical cross sections of APDN (m s\(^{-2}\); shaded), vertical velocity (contoured every 5 m s\(^{-1}\); black), and vertical vorticity (contoured every 0.005 s\(^{-1}\); positive values in green, negative values in pink, and zero contour suppressed for clarity) along the blue lines in (a,b) Fig. 4.25d at 1552 UTC, (c,d) Fig. 4.25e at 1554 UTC, and (e,f) Fig. 4.25f at 1556 UTC.
Figure 4.30: Vertical cross sections of (a,b) APDN (m s\(^{-2}\); shaded), (c,d) APB (m s\(^{-2}\); shaded). Also shown is vertical velocity (contoured every 5 m s\(^{-1}\); black) and vertical vorticity (contoured every 0.005 s\(^{-1}\); positive values in green, negative values in pink, and zero contour suppressed for clarity) along the blue lines in Fig. 4.25g at 1558 UTC.
Figure 4.31: As in Fig. 4.30 but along the blue lines in Fig. 4.25h at 1600 UTC.

surface based by 1558-1600 UTC, tilting and stretching increase the low-level $\zeta$ rapidly such that large negative APDN exists in the top half of the updraft, which is offset by positive APB in the same area. Only $0.01 \text{ s}^{-1}$ of $\zeta$ at 5 km MSL was needed to initiate this process (Figs. 4.13c, 4.26c,d), making it plausible that if some of the air in the upper portion of the strongly-sheared layer in the lowest 1 km enters an updraft (i.e., an elevated updraft ingesting a small amount of the highly-sheared air), sufficient $\zeta$ could develop aloft and aid the transition to a surface-based updraft. Although solar heating increases CAPE and reduces CIN, this process is less rapid in the anvil-shaded region. Thus, the accelerations owing to pressure perturbations may have been particularly important in allowing near-surface parcels beneath the anvil to penetrate through the moist-neutral layer.
Chapter 5

Conclusions

On 24 August 2016, unorganized elevated convection developed into three discrete, surface-based supercells, each of which produced multiple tornadoes across Indiana and Ohio. The event was unusual in that it was a challenging forecast and because multicellular convection became discrete supercells, which is a rare occurrence. Furthermore, many convection-allowing models failed to accurately capture this evolution, instead depicting a line of storms persisting across the affected area, contributing to the difficult forecast.

That morning, two clusters of elevated storms formed across Illinois, one of which developed along an outflow boundary that was traced back to convection in Nebraska and Iowa that developed the prior evening. There was also a differential heating boundary owing to the anvil of the leading cluster and that moved northeastward as this cluster dissipated. The updrafts in the trailing cluster exhibited only weak transient rotation while elevated, as there was little vertical wind shear above the 0-1 km layer. As these storms became surface based, they acquired rotation through tilting and subsequent stretching of horizontal vorticity in the 0-1 km layer. Each supercell formed and produced a significant tornado in the vicinity of the northeastward moving differential heating boundary. The presence of an MCV provided sufficient 0-6 km bulk wind shear for supercells.

A WRF model simulation accurately captures these features, including the MCV and the differential heating boundary. The simulation only depicts one initial cluster of elevated storms, which becomes a single surface-based supercell. This supercell forms about an hour before, and thus west of where radar observations indicate that the first supercell developed on 24 August 2016. In the simulation, the mesocyclone cycles after 1700 UTC, reaches peak intensity at 1900 UTC, and then becomes outflow dominant and weakens as in the observations.

Vertical cross sections of $w$ and $\zeta$ showcase multiple elevated cells in the simulation before 1550 UTC, and that the storms become surface based between 1550-1600 UTC. After this time, the updrafts become stronger and deeper until one deep, rotating updraft develops by 1700 UTC. The simulation reveals that the transition of disorganized elevated convection to a surface-based supercell is not instantaneous in this case, but instead that a few attempts at surface-based convection initiation occur before a dominant rotating
updraft develops, as is often observed when surface-based supercells form in environments initially free of precipitating convection.

Back trajectories confirm that the storms become surface based during this period. Soundings taken along trajectories before they enter the updrafts indicate the presence of a moist-neutral layer from 950-850 hPa, meaning that parcels do not become significantly buoyant until above 850 hPa (above 1 km) despite an LFC around 950 hPa. As solar heating increases parcel buoyancy in the moist-neutral layer, more parcels from the 0-1 km layer enter the updrafts.

The simulation also reproduces the differential heating boundary, the presence of which is likely owing to cloud shading and latent cooling. Since less vertical mixing occurs beneath the anvil, surface winds are more southerly there, increasing 0-1 km SRH. Back trajectories indicate that near-surface parcels entering the updrafts pass through this region. A gradient in \( \theta_v \) along the edge of the anvil also suggests that baroclinic generation of horizontal vorticity is possible there, further augmenting SRH in the storm inflow region. This analysis is consistent with the observed supercells forming in the vicinity of the differential heating boundary as it moved northeastward.

The WRF model output was manipulated to be compatible with code to perform a perturbation pressure decomposition, as has been done in many analyses of CM1 output. This new methodology required a sensitivity test of the accelerations that arise from vertical perturbation pressure gradients to the location of the base state because the initial environment in this simulation is horizontally inhomogeneous. The APDL is the most sensitive to the location of the base state, but it contributes less than 10% to the APTD in this environment. Both APB and APDN exhibit a standard deviation approximately an order of magnitude less than their means, indicating little sensitivity to the base-state location. Furthermore, the distribution of the pressure perturbations is independent of base-state location, consistent with the analysis of Doswell and Markowski (2004). Despite this small sensitivity, we recommend that the base-state location be selected such that the shear profile is representative of the ambient near-storm environment.

The development of vertical accelerations owing to perturbation pressure gradients likely aids surface parcels, particularly those near and north of the differential heating boundary, in overcoming the lack of significant buoyancy in the moist-neutral layer just above their LFC. The perturbation pressure decomposition indicates that \( \zeta \) acquired in elevated updrafts, perhaps from the ingestion of some air from just below 1 km, leads to positive APDN below the level of maximum rotation. As developing updrafts ingest more strongly-sheared near-surface air, \( \zeta \) increases and APDN becomes downward directed above the strongest rotation. In updrafts that persist, the negative APDN is offset by positive APB at these levels, allowing the updraft to grow, strengthen, and further stretch \( \zeta \). In updrafts that do not persist, upward-directed APB
does not offset the downward APDN.

The environment on 24 August 2016 was favorable for the development of supercells and even tornadoes, provided that supercellular convection was present. In this case, air from the lowest 1 km needed to be ingested by updrafts for significant rotation to form because this was the only layer where significant vertical wind shear existed. In environments such as this, in which elevated convection occurs above a weakly-stable but strongly-sheared near-surface layer, the development of weak rotation aloft can permit upward-directed non-linear dynamic perturbation pressure accelerations to form and allow for more of the helicity-rich near-surface air to enter the updrafts, and an intense rotating updraft may develop shortly afterward. In other environments characterized by strong vertical wind shear over a deeper layer, the transition from elevated to surface-based convection may occur differently than demonstrated in this case because the elevated convection may already be supercellular with significant upward-directed accelerations. The environment on 24 August 2016 was also quite moist, limiting the evaporative cooling potential in downdrafts and the production of surging outflow. If the initial storms had been more outflow dominant, it is possible that they would not have become supercells at all.

An important takeaway from this event is that environments favorable for supercells and tornadoes should not necessarily be discounted if convection-allowing models do not depict cellular convection, particularly if any ongoing convection is not outflow dominant or is poorly initialized. The development of improved techniques to assimilate radar data, and possibly the implementation of cloud microphysics parameterizations that more accurately represent the processes that occur in intense continental convective clouds, may help reduce such model errors in the future. Furthermore, the analysis of observations in real time, specifically the identification of boundaries such as the differential heating boundary, is also crucial for a better qualitative understanding of an environment. Despite this challenging forecast, we are especially grateful for the hard work and dedication of numerous operational and broadcast meteorologists who delivered timely and accurate warnings to the public during a tornado outbreak that occurred with little advance notice. Their efforts prevented any storm-related fatalities during this event and are commended.
References


