INTERNAL VS. EXTERNAL FORCINGS IN SUPERCELL INTERACTIONS AND THEIR IMPACT ON STORM MORPHOLOGY AND INTENSITY

BY

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THESIS

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ABSTRACT

This study examines supercell interactions from a suite of fifty-two idealized numerical simulations. The relative positions of two storm cells are varied for each case, and a single-cell simulation serves as the control. Despite initializing each simulation with an identical convective available potential energy and wind shear profile, small changes in the initial cell pair orientation lead to significant changes in subsequent storm morphology. Ninety-eight percent of the two-cell storm simulations produce stronger low-level mesocyclones than the single-cell control case, and low-level mesocyclone intensification is coupled with unsteady downdraft bursts in the forward flank. Downburst-driven, surface-based circulation centers form along the forward flank gust front, propagate toward the main updraft, and are stretched immediately prior to mesocyclogenesis events. These discrete rotation centers are approximately one kilometer deep and would most often be unobserved by operational radar. The external forcing associated with the initial storm cell pair orientation modulates the frequencies of the internal downbursts that drive the intensification of the low-level mesocyclone.
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Storm interactions ranging from intersecting outflows to complete mergers of initially discrete updrafts can have a profound impact on storm growth, propagation, and intensity. These interactions are capable of modifying storm morphology beyond what is characteristic for a particular environment. This remains a poorly understood area of convective behavior.

Operational meteorologists face a challenge forecasting (Moller 2011) and nowcasting (Andra et al. 2002, Roberts et al. 2006) storm severity because processes leading to favorable versus unfavorable interactions remain unidentified, and the intensification of interacting storms can occur rapidly (Lee et al. 2006). Several studies have suggested potential modes and implications for storm interaction, but aspects not quantified include how the location, intensity, and movement of storms are modulated by interactions. This additional source of uncertainty requires a better guidance for anticipating storm type and nowcasting storm intensification or decay as a consequence of storm interactions.

The objective of this numerical modeling study is to investigate how the initial orientation of two storms (external forcing) modulates processes that lead to the intensification of one or both of the storms (internal forcings). A preferred initial cell pair orientation for favorable storm interaction was identified for the environment used in this study, although this relationship is sensitive to small changes (~8 km) in the horizontal distance between the initial cell pair. Storm intensity is quantified in terms of surface vertical vorticity, and rotational properties such as longevity at specified thresholds and the thermodynamic character of surrounding air are examined. This study reveals the internal forcings responsible for low-level
mesocyclogenesis are an unsteady forward flank downdraft, tilting of horizontal vorticity along the forward flank gust front, and merging of discrete vertical rotation centers into the main updraft.

Chapter 2 contains a review of past literature relevant to the current study, the methodological and analysis techniques used to investigate storm interactions are presented in Chapter 3, and results and conclusions are outlined in Chapters 4 and 5, respectively.
CHAPTER 2
LITERATURE REVIEW

Changes in storm morphology from interactions with heterogeneities in the surrounding environment have been observed for decades. Byers and Braham (1949) were the first to document interacting storms based on observations during the Thunderstorm Project. A dramatic example of changes in storm morphology owing to interactions include a factor of twenty increase in precipitation associated with the merging of two cumulonimbi (Simpson 1971). A wide range of storm interaction modes exist, and Westcott (1984) provided a summary of key storm interaction studies and highlighted ambiguities associated with the term “merger” and its application to dissimilar storm interaction scenarios.

2.1 Storm interaction modes

Potential modes of storm interaction include a “bridge” feature that has been both observed and modeled (Simpson 1980; Tao and Simpson 1984, 1989). These studies showed that a region of enhanced low-level convergence between two interacting cells’ outflows initiated updraft development between the storms, forming a bridge-like feature in the reflectivity field (Figure 2.1). Westcott and Kennedy (1989) used reflectivity and triple-Doppler radar velocity data to illustrate the relevance of the low-level convergence field between interacting storms in modulating bridge development, differential storm motion, and initiating new cell growth. They also suggested the transport of hydrometeors aloft as a potential mechanism for reflectivity bridge development. Turpeinen (1982) emphasized the importance of the perturbation pressure
distribution in determining merger mechanisms using radar observations and numerical simulations.

2.2 Forecasting storm interactions

Atmospheric instability and vertical wind shear are fundamental to forecasting storm morphology. Weisman and Klemp (1982, 1984) used idealized numerical simulations to assess storm morphology versus convective available potential energy (CAPE) and deep-layer vertical wind shear profiles. Supercells are favored in environments with a balance of relatively large CAPE to strong vertical shear, and decreased vertical shear results in less organized pulse or multicell storms. In environments of lower CAPE, storm morphology and updraft propensity become increasingly dependent on the vertical distribution of CAPE. Strong, persistent updrafts are favored in low-CAPE environments when the instability is present at relatively lower altitudes (Kirkpatrick et al. 2011). The relationship of atmospheric instability to deep-layer vertical wind shear in forecasting storm morphology assumes the storm environment is horizontally uniform and does not account for intensity changes from interactions with other storms or mesoscale boundaries.

A critical factor for forecasting storm maintenance and propagation is the orientation of the deep-layer wind shear vector to cells initiating along a boundary. Bluestein and Weisman (2000) identified the optimum orientation for storm maintenance is 45°. At this angle, left-movers associated with splitting storms weaken over outflow associated with right-movers to the north, allowing right-moving storms to persist without colliding into left-moving storms. This finding was consistent with previous work by Lilly (1979, Figure 2.2). In their numerical modeling study of long-lived squall lines, Rotunno et al. (1988) determined squall lines
comprised of quasi-steady supercells are associated with deep-layer shear at an angle to the line of cells, which mitigates cell interference along the line.

2.3 Updraft buoyancy and storm interactions

Lemon (1976) analyzed a case in which a cell that initiated along a supercell’s flanking line merged with the parent storm, resulting in the invigoration of the parent storm’s updraft. Potential mechanisms outlined for the observed increase in buoyancy and vertical acceleration of the parent supercell’s updraft include the ingestion of undiluted updraft air comprised of higher potential temperatures, increased convergence within the mesocyclone due to increased buoyancy aloft, and microphysical alterations associated with the merging of the flanking line cumulus with the supercell updraft. An increase in buoyancy and decrease in pressure aloft enhanced surface convergence and intensified the updraft.

2.4 Constructive vs. destructive storm interactions

Storm interactions can be constructive or destructive. Using dual-Doppler analysis on a tornadic supercell, Wurman et al. (2007) suggested surface convergence associated with storm mergers can temporarily encourage tornadogenesis, but mergers can also increase the amount of cool air in the vicinity of the mesocyclone and cause the parent circulation to weaken after a short period of time. Using three-dimensional reflectivity data, Westcott (1994) determined the age of merging storm cores to be relevant for anticipating the outcome of a merger event. Younger cores, and those nearby newly mature storms, are more likely to grow in size. Kogan
and Shapiro (1996) showed the outcome of a merger between two interacting storms depended on the distance between the two initial perturbations, and Stalker and Knupp (2003) established a relationship between the planetary boundary layer depth and cell separation distance to cell merger potential and its affects on updraft strength and precipitation.

2.5 Storm-boundary interactions

Storm-boundary interactions are also capable of impacting storm morphology, and can result in the intensification of a storm and possibly tornadoes in cases synoptic conditions are not suggestive of tornadic storms (Maddox et al. 1980). Storm motion relative to the boundary is of importance, as supercell motion that is “closely aligned” with a boundary enhances the changes for long-lived supercells (Bunkers et al. 2006). Weaver and Nelson (1982) documented a case in which rapid storm intensification followed gust front interaction with a neighboring storm, followed twenty-five minutes later by the only reported tornado in the region. Wilson et al. (1988) identified a case of cell initiation and tornadogenesis associated with the collision of a weak cold front and an outflow boundary from older convection that had moved off to the east. Interactions with baroclinic boundaries excluding rear flank and forward flank gust fronts have been shown to locally enhance the horizontal vorticity in the surrounding storm environment, resulting in low-level mesocyclogenesis via tilting and stretching (Markowski et al. 1998). These baroclinic boundaries (e.g. old outflow boundaries) of enhanced horizontal vorticity are also hypothesized to be necessary for significant tornadoes (Rasmussen et al. 2000, Figure 2.3, Table 2.1). Vertical vorticity generated at the intersection of two mesoscale surface boundaries becoming collocated with an updraft and undergoing subsequent stretching resulting in an F1
tornado are evidence of the importance of mesoscale boundaries in nonsupercell tornadogenesis (Brady and Szoke 1989).

2.6 Storms interactions and tornadogenesis

A body of literature currently exists linking storm interaction to tornadogenesis. Lee et al. (2006) examined cell mergers and associated tornado incidence based on the merger-prolific April 19th, 1996 tornado outbreak. 54% of the 39 tornadoes in Illinois occurred within 15 minutes of a cell merger, and 55% of the tornadoes from that subset formed within five minutes of the merger time. Rotunno and Klemp (1985) highlighted the importance of horizontal vorticity along the forward flank for low-level mesocyclogenesis in their idealized numerical simulation of a supercell, however Markowski et al. (1998) identified 70% of tornadoes observed in the VORTEX field campaign were associated with boundaries in the storm’s surrounding environment. Markowski et al. concluded ambient horizontal vorticity associated with environmental vertical shear and horizontal vorticity generated along a storm’s own outflow boundaries are not sufficient for the genesis of significant tornadoes, but rather horizontal vorticity associated with mesoscale boundaries in the surrounding environment are critical to enhancing horizontal vorticity available for tilting and stretching. Goodman and Knupp (1993) analyzed an event involving tornado intensification as a result of the parent supercell’s interaction with squall line outflow to its west. Finley et al. (2001) performed a real-data simulation of a high-precipitation supercell that experienced five “daughter cell mergers” and intensified briefly after each merger. The updraft associated with the merged supercells was stronger than the two individual supercells’, and low-level vertical vorticity stretching intensified
during and just the merger process. The greater intensity associated with a merged updraft in 
Finley et al. (2001) was also noted in the numerical simulation studies of Wilkins et al. (1976) 

2.7 Low-level mesocyclone intensification

Additional components analyzed in this study include mechanisms associated with the 
intensification of the low-level mesocyclone. Klemp and Rotunno (1983) used a three-
dimensional cloud model to generate enhanced resolution simulations of a mature supercell in its 
tornadic phase. They determined the tilting and stretching of horizontal vorticity generated from 
the environmental wind shear as well as along baroclinic gradients along the forward flank 
contributes most to the intensification of low-level vertical vorticity, and the rear-flank 
downdraft occurs as a result of the inverse pressure gradient associated with the intensification of 
the low-level rotation. In their study of tornadogenesis and decay, Wicker and Wilhelmson 
(1995) also identified tilting of the environmental and baroclinically-generated vorticity along 
the forward flank by the updraft to be responsible for the genesis of the mesocyclone and tornado 
vortex, and argued tornadogenesis occurs when mid-level rotation increases. An increase in mid-
level rotation causes a pressure gradient that draws air upward and enhances convergence below 
cloud base.

Shabbott and Markowski (2006) used in situ measurements of forward flank downdraft 
outflow to determine that outflow associated with nontornadic supercells was more negatively 
buoyant, originated from higher altitudes, and consisted of larger baroclinic gradients and 
horizontal vorticity than outflow from tornadic supercells (Figure 2.4). These large baroclinic
gradients along the leading edge of the forward flank gust front may be associated with strong storm-relative helicity, where strong vertical shear produces drop size sorting and a polarimetric radar $Z_{DR}$ arc signature (Kumjian and Ryzhkov 2009).

Results from the study herein align with those suggesting that when horizontal vorticity generated along baroclinic gradients is added to the environment vorticity (Wicker and Wilhelmson 1995), the potential for a stronger mesocyclone increases. Interactions with additional environmental heterogeneities (i.e. storm interactions) affect the degree to which the intensification can occur. In the experimental design for this study (Chapter 3), a wide range of storm interaction scenarios are created, and preferred storm and boundary orientations are identified for a particular environment.
Figure 2.1: Tao and Simpson (1984). Left panel is a vertical cross section at T=156 min. Heavy black contours outline the cloud, and dark shading represents liquid water content > 1 gkg$^{-1}$. Right panel is a vertical cross-section of convergence with a contour interval of 2.0 x $10^{-3}$ s$^{-1}$. Stippled contours represent areas of divergence. Note cloud bridge and its co-location with an area of weak convergence.
Figure 2.2: Rotunno et al. (1988). Lilly’s proposal for a line of supercell thunderstorms existing at an angle to the shear so that each supercell could propagate without colliding with a neighbor. The shear profile is indicated on the left; the relative winds at low, middle and high levels are indicated by the L, M, and H symbols, respectively. The stippled region indicates the ‘hook-shaped’ rain area at the surface and the barbed line represents the micro-cold front. (Adapted from Fig. 15 of Lilly, 1979).
Figure 2.3: Echo centroid tracks for five 1-h periods illustrated by the different line types shown in the legend. Boundary positions are depicted by thick lines according to the symbols in the legend, and apply to the start of the 1-h periods. Numbers in circles correspond to tornadoes in Figure X (Rasmussen et al. 2000).
<table>
<thead>
<tr>
<th>Number</th>
<th>Location (distance in mi)</th>
<th>Time</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>29 ENE Mosquera, NM</td>
<td>2245</td>
<td>F0 tornado destroyed 30–40 power poles.</td>
</tr>
<tr>
<td>2</td>
<td>3 NE Olton, TX</td>
<td>2300</td>
<td>F0 (no damage) tornado reported by the fire department. The report is plausible given the presence of an intense echo just to the NE.</td>
</tr>
<tr>
<td>3</td>
<td>4 NE NRA Vista, NM–2 E Romero, TX</td>
<td>2320</td>
<td>Over open country; no damage. Verified through chaser reports. Rated F1.</td>
</tr>
<tr>
<td>4</td>
<td>1 N Glen Rio, NM</td>
<td>2320</td>
<td>F0 tornado verified through chaser reports as being 6–8 mi NNE of Glen Rio.</td>
</tr>
<tr>
<td>5</td>
<td>3 SW to 6 NNE Friona, TX</td>
<td>2342</td>
<td>F3 VORTEX intercept.</td>
</tr>
<tr>
<td>6</td>
<td>1 NW Edmondson, TX</td>
<td>2343</td>
<td>F0 (no damage) tornado reported by storm spotters. This report is plausible based on WSR-88D data.</td>
</tr>
<tr>
<td>5 E Har, TX</td>
<td>2345</td>
<td>Spotter report; no damage. Same event as 6.</td>
<td></td>
</tr>
<tr>
<td>7.5 N Dismid, TX</td>
<td>0022</td>
<td>Not observed by VORTEX teams in the vicinity; assumed to be erroneous.</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>3 3–2 NE Dismid, TX</td>
<td>0059</td>
<td>F2 VORTEX intercept.</td>
</tr>
<tr>
<td>8</td>
<td>Near Tulia, TX</td>
<td>0100</td>
<td>Storm chaser video shows a very small, brief tornado.</td>
</tr>
<tr>
<td>9</td>
<td>11 N Dismid, TX</td>
<td>0100</td>
<td>VORTEX intercept of a significant tornado. Not in Storm Data.</td>
</tr>
<tr>
<td>2 NE Edmondson, TX</td>
<td>0100</td>
<td>F0 (no damage) tornado reported by storm spotters. Not consistent with radar echoes; spotters may have seen the more distant tornado near Tulia at the same time because there was not intervening precipitation.</td>
<td></td>
</tr>
<tr>
<td>3 N Nazareth, TX</td>
<td>0102</td>
<td>Not observed by VORTEX teams in the vicinity; assumed to be erroneous. Observer probably was seeing the Dismid tornado and misjudged the location.</td>
<td></td>
</tr>
<tr>
<td>3 N Swisher County, TX</td>
<td>0102</td>
<td>Location does not exist; this F0 tornado cannot be utilized in the analysis owing to uncertainty of location.</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>9 NE Tulia, TX</td>
<td>0118</td>
<td>F1 tornado with 11-mi path length reported in Storm Data. This tornado was not observed by any chasers; the source of the report is unknown. Report is consistent with Doppler data.</td>
</tr>
<tr>
<td>3 SW Denver City, TX</td>
<td>0130</td>
<td>F0 (no damage) tornado reported by the public. Experienced storm chasers did not observe a tornado with this storm; this report is deemed to be erroneous.</td>
<td></td>
</tr>
</tbody>
</table>

Table 2.1: Table corresponding to tornadoes depicted by colored circles in Figure 2.3 above. (Rasmussen et al. 2000)
Figure 2.4: Shabbott and Markowski (2006). Three-dimensional schematic of a numerically simulated supercell thunderstorm in westerly mean shear, viewed from the southeast, at a stage when low-level rotation is intensifying. The cylindrical arrows depict storm-relative winds. The thin lines are vortex lines, with the sense of rotation indicated by the circular arrows. The heavy barbed line marks outflow the boundary. The orientation of the horizontal buoyancy gradient, $\Delta h B$, is also indicated. [Adapted from Klemp (1987)].
CHAPTER 3
METHODOLOGY

3.1 Model configuration

The three-dimensional, non-hydrostatic ARW Weather Research and Forecasting model (Skamarock et al. 2008) was used to generate a suite of fifty-two idealized numerical storm interaction simulations. A single-cell simulation served as the CONTROL. This cell was placed 3 km to the south of the model domain center. The subsequent fifty-one simulations were initialized with a second cell positioned to the southwest of the CONTROL cell at a varied location (Figure 3.1). This configuration was chosen based on results from a previous study by Jewett et al. (2006) showing the preferred configuration for favorable storm interaction (i.e. maximum surface rotation) is a cell pair oriented southwest to northeast. The parameter space was restricted to two cells in order to minimize the ambiguities associated with a multi-cell interaction.

All fifty-two horizontally homogeneous simulations were initialized using an identical idealized sounding (Section 3.3), and storms were generated using the warm bubble method. The CONTROL cell utilized a +3.0°C temperature perturbation centered at 1.5 km AGL. Each second cell to the southwest was initialized with +2.0°C perturbations, since most interacting storms in the real atmosphere are not identical in age and maturity. The initial temperature perturbations had an 8500 m horizontal radius and 1500 m vertical radius. The simulations were integrated for five hours within a 138 km x 151 km x 18 km model grid stretched vertically with 90 levels, and 540 m horizontal grid spacing. Vertical resolution in the low levels was more than sufficient, with 20 of the 90 levels present below 2 km AGL. The highest vertical resolution in
the stretched vertical grid was 96.2 m, and the model top was treated as a surface of constant pressure (Skamarock et al. 2008). A time step of 1.5 seconds was used for model integration based on tests for high-resolution simulations to be completed later. Model output was saved every minute.

3.2 Model physics

Operating within an idealized framework, longwave and shortwave radiation, surface layer and land surface physics, boundary layer processes, and heat and moisture surface fluxes, the Coriolis force, and surface friction were turned off in the suite of storm interaction simulations. Neglecting surface friction would not be ideal for simulations of tornadoes, but the horizontal grid spacing used in this study resolves mesocyclogenesis, not tornadogenesis.

Preventing storms from interfering with domain boundaries required modifications to storm motion in the initial sounding. Open boundary conditions were used to allow wave-like features to propagate out of the model domain. Additional model settings include a third-order Runga-Kutta time integration scheme and a turbulent kinetic energy (TKE) eddy-resolving diffusion scheme.

Cloud and precipitation processes were parameterized using the double-moment Morrison microphysics scheme (option 10 in WRF v.3.3). Morrison microphysics are double-moment (predicting number concentration as well as mass) in precipitating hydrometeors: rain, ice, snow, graupel/hail. The dense precipitating ice category flag was set to hail in expense of graupel, per recommendation for simulations of deep continental convection (Morrison et al. 2009). Three other microphysics schemes were evaluated prior to completing the suite of storm
interaction simulations, with the aforementioned scheme chosen based on a qualitatively more realistic supercell storm structure. For more on the model physics used in this study see Tables 3.1 and 3.2 in Section 3.5.

3.3 Sounding

The sounding used to initialize all fifty-two storm interaction simulations was taken from a Pennsylvania State University National Center for Atmospheric Research Mesoscale Model (MM5) real-data simulation of the April 19th, 1996 tornado outbreak (Figure 3.2). It is representative of the pre-convective environment east of the dryline in Illinois (Figure 3.3) and exhibits a well-mixed, mostly dry-adiabatic boundary layer and 53 knot 0-6 km vertical wind shear. The vertical wind shear in the cloud-bearing layer is unidirectional, indicating both left-moving and right-moving members of a storm split would persist and be of comparable intensity (Wilhelmson and Klemp 1978). Surface-based convective available potential energy (CAPE) is approximately 2500 Jkg\(^{-1}\), and convective inhibition is 77 Jkg\(^{-1}\). Storm initiation on this day occurred at approximately 2100 UTC along a dryline, warm front, and warm-front occlusion with frequent storm splits, merging, and subsequent tornadogenesis characterizing this event (Lee et al. 2006).

3.4 Analysis methods

Consistent with the research objective to identify external and internal forcings pertaining to favorable storm interactions, analysis techniques ranged from suite-wide comparisons to statistical analyses within individual storms. Fields included maximum and minimum vertical
velocity as well as surface vertical vorticity, wind speed, pressure, and minimum surface relative humidity (Table 3.2). Qualitative assessments of storm morphology were based on 2 km simulated reflectivity. Within a diagnosed cold pool, the average, minimum, and lowest 5% threshold perturbation temperatures were computed every minute for the 52 simulations. The duration for which a surface rotation center maintained an intensity of 0.005, 0.02, and 0.05 s\(^{-1}\) was tracked to assess intensity versus longevity, and identify how the strength and duration of the surface rotation centers change as a function of the initial cell pair orientation.

As part of the diagnosis of sub-storm processes, the ten strongest surface rotation centers were tracked within each simulation, and relevant variables and statistics were computed in relation to the rotation center for its lifespan (Table 3.3). Variables computed at each minute for the ten strongest rotation centers include the vertical vorticity value associated with the rotation center and thermodynamic properties of air within 2 and 4 km of the center, motivated by the study of Markowski (2002). Vorticity-tracking allows the character of the ten strongest surface rotation centers to be evaluated in terms of their kinematic and thermodynamic properties and understand how these properties impact the intensity and longevity of the rotation centers. The ultimate goal was to deduce how these storm-internal properties are modulated by the external forcing associated with initial cell pair orientation.
3.5 Chapter 3 figures and tables

Figure 3.1: Schematic of initial cell pair configurations. The red dot represents the position of the CONTROL cell relative to the position of a second cell in the fifty-one two-celled simulations. The CONTROL cell is located 3 km south of center in the full 138 km x 151 km model domain.
<table>
<thead>
<tr>
<th>Model parameterizations</th>
<th>Treatment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time integration scheme</td>
<td>3(^{rd}) order Runge-Kutta</td>
</tr>
<tr>
<td>Turbulence and mixing</td>
<td>Turbulent kinetic energy (TKE)</td>
</tr>
<tr>
<td>Numerical diffusion</td>
<td>6(^{th}) order: non-dimensional strength = 0.1; up-gradient diffusion prevention</td>
</tr>
<tr>
<td>Vertical diffusion</td>
<td>Acts on all fields (not limited to perturbations)</td>
</tr>
<tr>
<td>Microphysics</td>
<td>Morrison; double-moment; hail flag turned on</td>
</tr>
<tr>
<td>Friction</td>
<td>None; free-slip</td>
</tr>
<tr>
<td>Surface fluxes</td>
<td>N/A</td>
</tr>
<tr>
<td>Radiation</td>
<td>N/A</td>
</tr>
<tr>
<td>Sfc/boundary layer</td>
<td>N/A</td>
</tr>
</tbody>
</table>

Table 3.1: Model integration, diffusion, and microphysics schemes

<table>
<thead>
<tr>
<th>Advection – moisture and scalar (monotonic)</th>
<th>Order</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal momentum</td>
<td>5(^{th})</td>
</tr>
<tr>
<td>Vertical momentum</td>
<td>3(^{rd})</td>
</tr>
<tr>
<td>Horizontal scalar advection</td>
<td>5(^{th})</td>
</tr>
<tr>
<td>Vertical scalar advection</td>
<td>3(^{rd})</td>
</tr>
</tbody>
</table>

Table 3.2: Advection schemes
Figure 3.2: Initial sounding and hodograph used in storm interaction suite
Figure 3.3: Lee et al. (2006)
<table>
<thead>
<tr>
<th><strong>Analysis variables</strong></th>
<th><strong>Units</strong></th>
<th><strong>Description</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>VortMax</td>
<td>s⁻¹</td>
<td>Maximum surface vertical vorticity</td>
</tr>
<tr>
<td>VortMin</td>
<td>s⁻¹</td>
<td>Minimum surface vertical vorticity</td>
</tr>
<tr>
<td>T05</td>
<td>min</td>
<td>Time an individual surface vertical rotation center &gt; 0.005 s⁻¹</td>
</tr>
<tr>
<td>T20</td>
<td>min</td>
<td>Time an individual surface vertical rotation center &gt; 0.02 s⁻¹</td>
</tr>
<tr>
<td>T50</td>
<td>min</td>
<td>Time an individual surface vertical rotation center &gt; 0.05 s⁻¹</td>
</tr>
<tr>
<td>Wmax</td>
<td>ms⁻¹</td>
<td>Maximum vertical velocity</td>
</tr>
<tr>
<td>Wmin</td>
<td>ms⁻¹</td>
<td>Minimum vertical velocity</td>
</tr>
<tr>
<td>Amax</td>
<td>pts</td>
<td># grid points with maximum column vertical velocity &gt;= 10 ms⁻¹</td>
</tr>
<tr>
<td>Amin</td>
<td>pts</td>
<td># grid points with maximum column vertical velocity &gt;= -2 ms⁻¹</td>
</tr>
<tr>
<td>SfcWnd</td>
<td>ms⁻¹</td>
<td>Maximum surface wind speed – perturbation from base state</td>
</tr>
<tr>
<td>WndUpa</td>
<td>ms⁻¹</td>
<td>Maximum wind perturbation at/above 2 km AGL</td>
</tr>
<tr>
<td>Area0</td>
<td>pts</td>
<td># grid points with surface wind perturbation &gt; 20 ms⁻¹</td>
</tr>
<tr>
<td>Area3</td>
<td>pts</td>
<td># grid points with wind perturbation at 3 km AGL &gt; 15 ms⁻¹</td>
</tr>
<tr>
<td>WindGR</td>
<td>ms⁻¹</td>
<td>Maximum ground-relative surface wind</td>
</tr>
<tr>
<td>Pmin</td>
<td>mb</td>
<td>Minimum surface pressure perturbation</td>
</tr>
<tr>
<td>Pmax</td>
<td>mb</td>
<td>Maximum surface pressure perturbation</td>
</tr>
<tr>
<td>CPavg</td>
<td>°C</td>
<td>Average surface temperature perturbation in diagnosed cold pool</td>
</tr>
<tr>
<td>CPmin</td>
<td>°C</td>
<td>Minimum surface temperature perturbation in diagnosed cold pool</td>
</tr>
<tr>
<td>CP05%</td>
<td>°C</td>
<td>5% level surface temperature perturbation in diagnosed cold pool</td>
</tr>
<tr>
<td>Rn02</td>
<td>pts</td>
<td># grid points with surface rain total &gt; 0.2 in</td>
</tr>
<tr>
<td>Rn25</td>
<td>pts</td>
<td># grid points with surface rain total &gt; 1 in</td>
</tr>
<tr>
<td>Qpr02</td>
<td>pts</td>
<td># grid points with surface rain mixing ratio &gt; 0.2 gkg⁻¹</td>
</tr>
<tr>
<td>Qpr20</td>
<td>pts</td>
<td># grid points with surface rain mixing ration &gt; 2.0 gkg⁻¹</td>
</tr>
<tr>
<td>RHmin</td>
<td>%</td>
<td>Minimum surface relative humidity</td>
</tr>
<tr>
<td>RHar</td>
<td>pts</td>
<td># grid points with surface RH &lt;= 25%</td>
</tr>
</tbody>
</table>

Table 3.3: Data analysis variables
<table>
<thead>
<tr>
<th>Tracked Variables</th>
<th>Units</th>
<th>Description (at x,y,1 location of sfc vorticity max)</th>
</tr>
</thead>
<tbody>
<tr>
<td>VortSfc</td>
<td>s⁻¹</td>
<td>Surface vorticity</td>
</tr>
<tr>
<td>VrtDepth</td>
<td>m</td>
<td>Vorticity maximum depth &gt;= 200 s⁻¹</td>
</tr>
<tr>
<td>Usfc</td>
<td>ms⁻¹</td>
<td>U wind component at vort max location</td>
</tr>
<tr>
<td>Vsfc</td>
<td>ms⁻¹</td>
<td>V wind component at vort max location</td>
</tr>
<tr>
<td>Wind2km</td>
<td>ms⁻¹</td>
<td>Maximum wind speed within 2 km</td>
</tr>
<tr>
<td>VortH2km</td>
<td>s⁻¹</td>
<td>Horizontal vorticity within 2 km</td>
</tr>
<tr>
<td>Wlow</td>
<td>ms⁻¹</td>
<td>Lowest-level vertical velocity</td>
</tr>
<tr>
<td>Wmin</td>
<td>ms⁻¹</td>
<td>Minimum vertical velocity for all heights</td>
</tr>
<tr>
<td>Wmax</td>
<td>ms⁻¹</td>
<td>Maximum vertical velocity for all heights</td>
</tr>
<tr>
<td>P</td>
<td>mb</td>
<td>Surface pressure perturbation</td>
</tr>
<tr>
<td>Convg</td>
<td>s⁻¹</td>
<td>Surface convergence</td>
</tr>
<tr>
<td>RHsfc</td>
<td>%</td>
<td>Relative humidity at the surface</td>
</tr>
<tr>
<td>Tsfc</td>
<td>°C</td>
<td>Surface temperature</td>
</tr>
<tr>
<td>Tgrad</td>
<td>°Ckm⁻¹</td>
<td>Surface temperature gradient</td>
</tr>
<tr>
<td>Qhail</td>
<td>gkg⁻¹</td>
<td>Surface hail mixing ratio</td>
</tr>
<tr>
<td>Qra</td>
<td>gkg⁻¹</td>
<td>Surface rain mixing ratio</td>
</tr>
<tr>
<td>Eth_2km_max</td>
<td>K</td>
<td>Max θₑ within 2 km</td>
</tr>
<tr>
<td>Eth_2km_min</td>
<td>K</td>
<td>Min θₑ within 2 km</td>
</tr>
<tr>
<td>Eth_4km_max</td>
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<td>Max θₑ within 4 km</td>
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<tr>
<td>Eth_4km_min</td>
<td>K</td>
<td>Min θₑ within 4 km</td>
</tr>
<tr>
<td>Vth_2km_max</td>
<td>K</td>
<td>Max θᵥ within 2 km</td>
</tr>
<tr>
<td>Vth_2km_min</td>
<td>K</td>
<td>Min θᵥ within 2 km</td>
</tr>
<tr>
<td>Vth_4km_max</td>
<td>K</td>
<td>Max θᵥ within 4 km</td>
</tr>
<tr>
<td>Vth_4km_min</td>
<td>K</td>
<td>Min θᵥ within 4 km</td>
</tr>
</tbody>
</table>

Table 3.4: Variables measured relative to surface rotation maximum location for each data time of the rotation center’s lifespan, for the strongest ten in each simulation.
CHAPTER 4
RESULTS

Fifty-two idealized storm interaction simulations were generated using the WRF model (v3.3) and an identical sounding. The only change made among the fifty-one two-cell simulations was the initial position of the second thermal perturbation (and resulting cell initiation location) relative to the first. A single-cell simulation acted as the control case. The objective was to determine how external forcings (i.e. the initial cell pair orientation) affect internal storm processes (i.e. downdrafts, baroclinic vorticity generation). The overall nature of intensification within the storms in these simulations is episodic. The cyclic behavior is consistent with observations from the April 19th, 1996 Illinois tornado outbreak (Lee et al. 2006). Supercells on that day were steady and long-lived, and tornadoes were brief. Given the sounding used to initialize these simulations was taken from a real-data simulation of April 19th, 1996 immediately prior to convective initiation, it is appropriate that the general storm behavior be qualitatively similar.

4.1 Isolated cell control case

The control storm splits at approximately 60 minutes (Figure 4.6), and the left-mover exists the domain by 2 hours (Figure 4.7). At 90 minutes, the right-mover attains a steady-state updraft intensity oscillating between 40-60 ms$^{-1}$ (Figure 4.1), the cold pool strengthens in terms of spatial coverage (Figure 4.2), and after a rapid decline as precipitation processes developed, the average negative cold pool temperature perturbation leveled off, but continued to cool.
gradually (Figure 4.3). Winds at the surface also reach their relative maximum of 33 ms\(^{-1}\) at 90 minutes, and gradually increased to an intensity of 46 ms\(^{-1}\) near the time of the strongest surface rotation (Figure 4.4). Rotation at the surface fluctuated significantly as the storm split and the right-mover matured. At 180 minutes, the surface rotation steadily intensified until reaching its peak of 0.0493 s\(^{-1}\) at 3 hours 38 minutes (Figure 4.5, 4.8, 4.9). Select two-cell runs will be discussed in more detail in the following sections.

4.2 External forcing: Initial cell pair orientation

In the 51 simulations that followed the control, a second thermal perturbation was placed in the model initial condition to examine cell interaction. Altering the orientation of the initial cell pair had significant effects on storm morphology as the interactions progressed in time. Storm interactions were shown to impact the ability of an interacting storm to maintain its supercell organization versus a multi-cell structure, the storms’ cold pool behavior, the orientation of the forward flank gust front relative to the main updraft, and the degree to which baroclinic horizontal vorticity was generated.

One of the approaches taken to examine the morphological variations as a result of changing the initial orientation of the interacting cell pair was to compare the peak updraft intensity, minimum cold pool temperatures, and surface rotation among the control case and the fifty-one two-cell simulations. At any given time, the peak updraft intensities fluctuated between 40-60 ms\(^{-1}\). Considering that the same sounding was used to initialize all fifty-two simulations, it is not unexpected that storms in these simulations produced comparable peak updraft intensities (Figure 4.10).
Minimum surface temperature perturbations were computed for the diagnosed cold pool region for all 52 simulations. These were also found to be of similar magnitude (Figure 4.11). On average, the minimum cold pool temperature was -10.0°C during the mature stage of the storms.

In terms of updraft and cold pool strength, the magnitudes of these variables were quantitatively similar regardless of the orientation of the initial cell pair and subsequent storm interaction processes. Considering the identical initial environment used in all simulations, this result is also not surprising. However, it calls into question what modulates storm intensity when observed for storms similar in buoyancy and cold pool magnitude.

Despite similar magnitudes in buoyancy and cold pool extrema, a wide range of maximum intensities were generated based only on changing the initial orientation of the interacting cell pair (Figure 4.12). The majority (24 of 52) of the simulations produced maximum surface vertical vorticity values of 0.06 s⁻¹, while the weakest storms produced surface rotation of 0.04 s⁻¹. The low-end threshold for mesocyclone strength is 0.01 s⁻¹, suggesting relatively healthy rotation is produced in all simulations. The two strongest storm interaction cases produced surface rotation in excess of 0.1 s⁻¹.

A particularly significant result is highlighted in Figure 4.13. Fifty-one of the fifty-two two-cell cases produced stronger rotation at the surface than the control case containing an isolated storm. This highlights the critical need to understand how interactions can modify the evolution and intensity of storms within a given environment, since all storms were initialized using the same environmental profile, but the presence of a second cell produced stronger surface rotation 98% of the time. Results presented earlier showing similar magnitudes in updraft strength and cold pool extrema also call into question what modulates storm intensity, the
magnitudes of these variables were comparable among all simulations, yet the storms produced a wide range of peak surface rotation intensities.

The two strongest storm interaction simulations exceeding 0.1 s\(^{-1}\) surface vertical vorticity have similar initial cell pair orientations (Figures 4.14 and 4.15). This suggests that despite the broad range of maximum surface vertical vorticity intensities generated in the suite of fifty-two simulations, this relationship is likely physical, not random. The two storms each split at approximately one hour into the simulation, and WRF_20 reaches its peak surface rotation at 3 hours 18 minutes and WRF_04 at 3 hours 45 minutes. They had peak intensities of 0.11 and 0.10 s\(^{-1}\), respectively. The eastern-most cell (of the original cell pair at the start of the simulations) becomes the strongest in both of these simulations (Figures 4.16 and 4.17).

The rest of this chapter will focus on the results of three simulations: the single-cell control case, a relatively weak case WRF_03, and a relatively strong case discussed briefly above, WRF_04. The initial cell pair in WRF_03 has a horizontal separation of 20 km, and the strong case has the initial cell pair spaced 28 km apart. Despite only an 8 km difference in the horizontal distance of the two initial thermals between the strong and weak two cell cases, WRF_04 achieved a factor of two larger surface vertical vorticity magnitude than WRF_03 (Figure 4.18).

The 8 km variation in horizontal distance between the initial cell pairs in WRF_03 and WRF_04 had a profound effect on the morphology of the interacting storms (Figure 4.18). Considering the isolated control cell case, it retains much of its organization and supercellular structure throughout the duration of the simulation, with episodes of short-lived surface rotation intensification. This single storm cycled twice during the five-hour simulation, and reached its peak intensity of 0.049 s\(^{-1}\) at 3 hours 38 minutes. The second set of reflectivity panels pertaining
to the relatively weak case, WRF_03, show this interaction resulted in a more multi-cell storm structure between two and three hours into the simulation. The easternmost storm was characterized by multiple vertical vorticity intensification episodes (hereafter “cycled”) a total of three times, and reached its maximum intensity at 4 hours 11 minutes. The strong case, WRF_04, also cycled three times, but maintained its supercell organization to a larger extent than WRF_03, particularly during the 3 to 4 hour time period when it reached its peak intensity. A small perturbation in the initial spatial configuration of these cells produced storms of different morphologies within identical initial instability and shear environments.

4.3 Forward flank gust front: A mesocyclone intensification source

After determining that small changes in the orientation of the initial cell pair produced a large degree of variation in surface rotation intensity, the next phase consisted of examining internal forcings relative to individual storms. This included comparing and contrasting kinematic and thermodynamic properties near the surface that are modulated by the orientation of the initial cell pair that could explain observed differences in storm intensity.

Figure 4.19 shows positive vertical vorticity at the surface for the strong two-cell case, WRF_04. Several discrete vertical rotation centers are generated along the forward flank gust front of the easternmost storm and merge with the mesocyclone. The merging of discrete vorticity maxima at the mesocyclone is coincident with peaks in surface vertical vorticity (Figure 4.20). The second-to-last peak in Figure 4.20 occurs as the discrete rotation center in panel B of Figure 4.19 merges with the mesocyclone. The last peak in Figure 4.20 is associated with the
merging of the discrete rotation center shown just prior to merging with the mesocyclone in panel D of Figure 4.19.

Figure 4.21 reveals these positive vertical vorticity rotation centers are part of a couplet generated along the forward flank gust front. The forward flank gust front is characterized by temperature gradients between the ambient air found ahead of the storm and air cooled by evaporation and melting. Along this temperature gradient, vertical vorticity usually exists and baroclinically-generated horizontal vorticity (associated with horizontal buoyancy gradients) is certain to be found. The development of *discrete* vertical vorticity maxima along the forward flank is now considered.

Past studies have noted shearing instability as a potential source of positive vertical vorticity maxima along a storm’s gust front (Lee and Wilhelmson 1997). Figure 4.22 is a depiction of shearing instability from Batchelor (1967). Perturbations in the wind field along a vortex sheet cause vorticity to be depleted from point C and added to points B and D. Generally, shearing instability produces vorticity maxima of one sign. In the simulations in this study, vertical vorticity along the forward flank gust front is associated with a pair of counter-rotating vortices. This observation suggests horizontal vorticity is becoming vertically oriented through tilting, not as a result of shearing instability.

Tilting requires vertical motion, e.g. a downdraft originating above or convergence leading to an updraft acting upon the pre-existing horizontal vorticity along the forward flank gust front. For the forward flank, horizontal vorticity is directed toward the mesocyclone (to the south or southwest). A downdraft immediately above the forward flank would tilt horizontal vorticity into the vertical such that the vorticity couplet would exhibit cyclonic vorticity to the south and anticyclonic vorticity to the north (Figure 4.23, panel A). Tilting by an updraft would
produce cyclonic rotation to the north, and anticyclonic rotation to the south (Figure 4.23, panel B). The latter orientation is consistent with that observed in the couplet in Figure 4.21, suggesting pre-existing horizontal baroclinic vorticity is being tilted upward by the positive vertical velocity along with the forward flank gust front. It is hypothesized that the vorticity is tilted upward by a downburst-type mechanism from behind (west) of the forward flank gust front. Momentum associated with this surface divergence reaches the forward flank, increasing convergence and ascent there, and tilting horizontal vorticity upward. The positive vertical vorticity is preferentially stretched by the positive vertical motion found along the forward flank gust front.

The discrete surface vertical rotation centers are initially < 1 km deep and approximately 1-3 km in diameter. They propagate down the forward flank gust front and are stretched by the main updraft prior to being ingested into the mesocyclone. The degree to which these discrete rotation centers along the forward flank gust front are stretched prior to becoming ingested into the storm’s mesocyclone are 1) the proximity of the downburst to the forward flank gust front 2) the length of time the rotation center resides along the forward flank gust front prior to being ingested by the mesocyclone 3) the strength of the temperature gradient along the forward flank gust front and 4) the orientation of the forward flank gust front boundary to the mesocyclone. Downbursts closer to the forward flank gust front boundary tilt horizontal vorticity more effectively than downbursts that are further removed. In the strong two-cell case, WRF_04, a downburst occurs approximately 3 km to the west of the forward flank gust front boundary (Figure 4.24, panel A). This downburst is evidenced by the divergence in the surface wind field to the west of the gust front and the pocket of relatively warmer air coincident with the divergence signature. The relatively warmer air associated with these downbursts will be
discussed in section 3. This downburst is coincident with the generation of two discrete cyclonic rotation centers to the northeast of the mesocyclone. The mesocyclone in Figure 4.24 is the southwestern-most and strongest rotation center. Between the times shown in Figure 4.24, the mesocyclone ingested the discrete rotation center immediately to the northeast and increased in intensity from 0.025 s$^{-1}$ to 0.035 s$^{-1}$ as a result. The northeastern-most rotation center intensified from 0.015 s$^{-1}$ in panel A at 3 hours 24 minutes to 0.045 s$^{-1}$ by 3 hours 41 minutes as it propagated along the forward flank gust front and was stretched by the stronger main updraft. WRF_04 reached its peak intensity at 3 hours 45 minutes after the mesocyclone ingested the strengthening discrete rotation center.

Figure 4.24 also illustrates that the longevity of the discrete rotation center along the forward flank gust front modulates the degree to which it is stretched prior to merging with the mesocyclone. In panel A, the discrete rotation center to the northeast closest to the mesocyclone and was stretched further while traveling 5 km along the forward flank gust front before merging with the mesocyclone. The rotation center furthest to the northeast in panel A traveled 20 km along the forward flank gust front and was stretched by a factor of 13 before merging with the mesocyclone. Longer residence times along the forward flank gust front allow discrete rotation centers to be stretched more substantially. Stronger discrete rotation centers are associated with stronger low-level mesocyclone intensification upon merging.

In addition to the length of time the discrete rotation centers persist along the forward flank gust front, their maintenance and intensity are modulated by the temperature gradient across the boundary. Despite the relatively warm pocket of air associated with the difference 3 km to the west of the forward flank gust front in panel A of Figure 4.24, the temperature gradient across the boundary strengthened from 2.0°C at 3 hours 24 minutes (21 minutes prior to
maximum vorticity time, hereafter MVT) to 6.0°C at 3 hours 34 minutes, 11 minutes before MVT (Figure 4.25, panel A). An examination of discrete rotation centers in the isolated cell control case revealed how they can be weakened by a relatively weak temperature gradient along the forward flank gust front boundary. Similar to WRF_04, vorticity is tilted vertically along the gust front in the control simulation. In panel A of Figure 4.26, a discrete rotation center is just about to merge with the mesocyclone from the northeast. An isolated rotation center is present further to the northeast with an intensity of 0.01 s⁻¹. The temperature gradient across the forward flank gust front boundary between the five-minute time span depicted in panels A and B of Figure 4.26 is 1.0°C.

It is important to note the circulation associated with the discrete rotation center in panel A decreases by a factor of 2 in Panel B and becomes more elongated. This discrete rotation center never merges with the mesocyclone. A temperature difference of 5.0°C as seen in WRF_04 is sufficient for maintaining discrete rotation centers along the forward flank gust front but a 1.0°C temperature difference (as noted in the isolated control case) is too weak and the rotation decays before it can reach the mesocyclone.

The orientation of the forward flank gust front boundary also plays a role in the extent to which the low-level mesocyclone intensifies owing to mergers with discrete vorticity maxima. In the suite of 52 simulations, the tilting and stretching of discrete vorticity maxima along the forward flank occurs to the largest extent when the boundary is oriented southwest to the northeast relative to the mesocyclone (Figures 4.28, 4.32 and 4.33). The 8 km difference in horizontal distance between in the cell pair used to initialize WRF_03 and WRF_04 impacted storm morphology to the extent that WRF_04 obtained an evidently optimum boundary orientation while the structure of WRF_03’s surface boundaries were markedly different. In
WRF_03, the two-cell case found to be half the intensity of WRF_04, weaker vertical vorticity centers were ingested from behind the mesocyclone (Figures 4.29, 4.30 and 4.31) as opposed to along a 45°-oriented forward flank gust front boundary. There was one identifiable discrete rotation center generated along the boundary behind the mesocyclone and it was only slightly stretched before it merged with the mesocyclone and the storm reached its peak intensity at 4 hours 11 minutes.

The merging of discrete rotation centers along the forward flank gust front is coincident with low-level mesocyclogenesis events. Figures 4.34 and 4.35 summarize the mechanisms through which low-level mesocyclogenesis occurs. Figure 4.34 is a vertical cross section oriented east-west across the southwest-northeast trending forward flank gust front of the easternmost storm in WRF_04. The surface is colored by divergence with surface winds. At the time of this figure, 3 hours 31 minutes, or 14 minutes prior to the time of peak intensity, a 2 ms$^{-1}$ downburst (shaded with cool colors in the vertical cross section) occurs approximately 5 km to the west of the forward flank gust front boundary. The downburst spreads out laterally at the surface, leading to enhanced convergence and rising motion along the forward flank gust front to its east. This downburst is responsible for tilting horizontal vorticity into the vertical (Figures 4.27 and 4.35), where it is preferentially stretched by the updrafts along the convergence zone associated with the gust front. Figure 4.35 shows the vorticity arch feature responsible for generating the vertical vorticity couplet along the forward flank gust front. Vortex arching has been studied previously in relation to a supercell’s rear-flank downdraft and its role in tornadogenesis (Markowski 2008). Results from this study highlight the significance of vortex arching occurring along the forward flank gust front and the necessary role for low-level mesocyclogenesis the positive member has upon merging with the primary mesocyclone.
4.4 Unsteady forward flank downdraft: Downbursts

Vortex arching and the merging of discrete rotation centers along the forward flank gust front with the mesocyclone are tied to the location and timing of downbursts. This relationship was observed in all 52 simulations completed for this study. Because these downbursts appeared to be the driving mechanism for tilting horizontal vorticity vertically, identifying their source was critical.

An unsteady, pulse-like behavior was attributed to all forward flank downdrafts in the suite of simulations (Figure 4.36). They were characterized by bursts of relatively warm air at the surface relative to surrounding cold pool air. These bursts were between 1-4 km in diameter and approximately 2.0°C warmer than the surrounding cold pool air and occurred frequently but in various locations.

Figure 4.37 is a vertical cross-section of reflectivity for the strong two-cell case, WRF_04, oriented southwest to northeast just west of the forward flank gust front. It transects the downburst 5 km west of the forward flank gust front at 3 hours 21 min associated with tilting the vorticity show in Figure 4.19. There is a 4 gkg\(^{-1}\) maximum in the rain-water mixing ratio field at 3 km AGL. This maximum is co-located with a region of melting hail depicted by the relative maximum in hail mixing ratio and absence at lower elevations. Melting hail at 3 km AGL generated a maximum in reflectivity (Figure 4.38) aloft twenty-four minutes prior to the time of peak storm intensity, and was associated with a 48 dBz reflectivity signature. Melting and evaporation created the downburst that descended to the surface in close proximity to the forward flank gust front and caused extensive tilting along the forward flank gust front.

All simulations in the study herein produced unsteady forward flank downdraft behavior. This behavior was also observed in microphysical tests using the Milbrandt-Yau and Thompson
schemes in WRF, suggesting it is not a unique consequence of one microphysics scheme. Such unsteadiness has been noted in a separate tornado modeling effort (Greg Tripoli, personal communication). The surface bursts produced by the melting hail at 3 km AGL and a corresponding 48 dBz reflectivity signature were warmer than the surrounding cold pool air, but this temperature deviation persisted to <1 km AGL. The horizontal winds at the surface associated with these bursts ranged from 5-20 ms$^{-1}$.

Based on these observations, a wet-microburst type mechanism was proposed to explain the occurrence of the downbursts and the overall unsteady behavior of the forward flank downdraft. Melting hail, precipitation loading, relatively warmer temperatures up to 1 km AGL, and significant precipitation at the surface are key identifiers for wet microbursts (Wakimoto and Bringi 1988). Srivastava (1985) used data from the JAWS field campaign to identify maximum reflectivity signatures observed with wet microbursts as a function of the mid-level lapse rate (Figure 4.39) and found an environmental lapse rate of approximately $-7.8$ Kkm$^{-1}$ produced microbursts within the 50 dBz range. In this study, the initial environmental lapse rate was $-7.8$ Kkm$^{-1}$ and the microbursts were associated with 48 dBz reflectivity signatures$^1$.

A wet-microburst-type mechanism is responsible for tilting baroclinically-generated horizontal vorticity into the vertical in all 52 simulations, but the location and timing of these downbursts are modulated by orientation of the initial cell pair. The downbursts are associated with the storm-internal process of melting hail at 3 km AGL, but the spatial distribution of hail aloft is modulated by the orientation of the initial cell pair (Figure 4.40).

The isolated cell control case and WRF_04, the relatively strong two-cell case, both exhibit supercell structure and qualitatively comparable hail distributions at 3 km. They had

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$^1$ The reflectivity calculations in this study were not tuned directly for the Morrison microphysics scheme, and so can only be used for qualitative comparisons.
downbursts occur immediately west of their forward flank gust fronts, both with the 45° optimum boundary orientation (However, the control case had a considerably weaker temperature gradient across its forward flank gust front boundary than WRF_04, which led to weakened the discrete rotation centers). At 3 hours, WRF_03 appeared less organized and more multi-cellular compared to WRF_04 and the control. WRF_03 did not achieve the optimum forward flank boundary orientation associated with simulations containing organized, steady supercells. All three cases at this time exhibit maximum hail mixing ratios on the order of 5 gkg⁻¹, but the variation in spatial distribution was a critical factor for determining the location and timing of downbursts relative to the forward flank gust front boundary.

4.5 Forecasting implications

This study has revealed that even under identical buoyancy and shear conditions, a wide variety of convective scenarios are possible, including an enhanced possibility for mesocyclogenesis due to interaction between nearby storms. This is a forecasting challenge since it extends the range of possible severe outcomes compared to that expected for a single, discrete thunderstorm. An improved understanding of how storm type may be influenced by nearby storms could directly benefit the skill and uncertainty in operational forecasting of severe weather events.

Nowcasting requires diagnosing and predicting the likely short-term trends of severe weather. The discrete rotation centers found in this study along the forward flank gust front are shallow features, initially less than 1 km deep (Figure 4.41). They increase in depth and intensity as they propagate along the forward flank gust front and are stretched by the main updraft
immediately prior to merging with the mesocyclone. Unless storms are in close proximity to radar, a 0.5° elevation scan would not detect these features. Given their maximum lifespan of 20-30 minutes prior to merging with the mesocyclone and instigating low-level mesocyclogenesis, awareness of these features and better detection could aid nowcasters in anticipating a storm intensification episode.

In terms of detecting wet-microburst-type features that generate the discrete rotation centers sustained along the forward flank gust front, dual-polarization radar (Atlas et al. 2004) and the use of spectral width (Melnikov and Doviak 2002) are potential tools that could improve the detection of these features.
4.6 Chapter 4 figures and tables

Maximum and Minimum Vertical Velocity

Figure 4.1: Maximum and minimum vertical velocity over the entire domain at each minute of simulation time for the control case.
Surface Area with Reflectivity > 40 dBz

Figure 4.2: Number of grid points in the domain with surface reflectivity exceeding 40 dBz.
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Figure 4.7: Same as 4.6. T=120 minutes
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Figure 4.9: Color shading is surface vertical vorticity ($10^{-5}$ s$^{-1}$). Light and dark shading are 10 and 40 dBz reflectivity contours, respectively. Arrows show surface winds.
Maximum Vertical Velocity

Figure 4.10: Peak updraft intensity time series (m\(s^{-1}\)) for all 52 simulations.
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Figure 4.11: Minimum surface temperature perturbation time series for all 52 simulations
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Figure 4.12: Number of runs achieving a maximum intensity within a given surface vertical vorticity threshold.
Figure 4.13: Maximum surface vertical vorticity for each simulation. Single-cell control case highlight in red.
Figure 4.14: Strong case – WRF_04, at 0.5 hours. 2-km simulated radar reflectivity is shaded, and surface wind vectors are also shown. The second cell is 28 km to the west and 9 km to the south of the control cell.
Figure 4.15: Second strong case – WRF_20, at 0.5 hours. 2-km simulated radar reflectivity is shaded, and surface wind vectors are also shown. The second cell is 32 km to the west and 5 km to the south of the control cell.
Figure 4.16: Strong case WRF_04, at 3 hours 45 minutes, time of peak surface rotation. 2-km simulated radar reflectivity is shaded, and surface wind vectors are also shown.
Figure 4.17: Second strong case – WRF_20, at 3 hours 18 minutes, time of peak surface rotation. 2-km simulated radar reflectivity is shaded, and surface wind vectors are also shown.
Figure 4.18: Time evolution of the control case, WRF_03, and WRF_04 from 0.5 hrs to 4 hrs. Plots are simulated reflectivity at 2 km above ground. Peak surface rotation is noted at the top of the diagram for each simulation. For 2-4 hours, an approximately 95 x 55 km sub-domain is shown.
Figure 4.19: WRF_04. Color shading is vertical vorticity at the surface and red lines are 500 s\(^{-1}\) vertical vorticity contours. The mesocyclone is the rotation center furthest to the southwest. Plots at A) 3 hours 23 min. B) At 3 hours 28 min. C) 3 hours 40 min. D) 3 hours 45 min – time of peak intensity. Features are within a 19.5 km x 27 km subdomain.
Figure 4.19 (cont).
Figure 4.20: Mesocyclone intensity time series for WRF_04 leading up to the time of peak surface rotation.
Figure 4.21: Plan view of surface vertical vorticity couplet along forward flank gust front in WRF_04. Color shading is surface vertical vorticity, positive vorticity in warm colors negative vorticity in cold colors. Red lines are 0.05 s$^{-1}$ vertical vorticity contours. Arrows are winds at the surface. Feature are within a 10.8 km x 13.0 km subdomain.

Figure 4.22: Shearing instability. Batchelor (1967)
Figure 4.23: Yellow bar represents pre-existing horizontal vorticity along the forward flank gust front. A) Horizontal vorticity is tilted upward by a directly-hitting downdraft, producing anticyclonic rotation to the north and cyclonic rotation to the south. B) Horizontal vorticity is tilted upward by an updraft along the forward flank gust front, producing cyclonic vorticity to the north and anticyclonic vorticity to the south.
Figure 4.24: Forward flank gust front plan view of easternmost storm in WRF_04. North is up. Color shading is surface temperature, black lines are vertical vorticity contours (0.05 s$^{-1}$) and arrows are surface winds. A) At 3 hrs 24 min. B) At 3 hrs 41 min; 4 minutes prior to WRF_04’s peak intensity. Figures are 11.8 km x 18.3 km.
Figure 4.25: Same as Figure 4.24, except panel A is at 3 hours 34 minutes. Figures are 11.8 km x 18.3 km.
Figure 4.26: Same as Figure 4.24, for isolated cell control case. White lines are surface vertical vorticity $0.05 \, \text{s}^{-1}$ contours. A) 3 hours 25 min. B) 3 hours 30 min. 13 min. and 8 min prior to MVT, respectively. The subdomain is 8.6 km x 17.2 km.
Figure 4.27: 3D subdomain of WRF_04. Pink isosurface is $0.0122 \, s^{-1}$ vertical vorticity and purple isosurface is $2.08 \, gkg^{-1}$ rain-water mixing ratio. Surface color shading is vertical vorticity. Arrows are winds at the surface. A) 3 hours 8 min. B) 3 hours 25 min. C) 3 hours 33 min. D) 3 hours 38 min. E) 3 hours 45 min. 29 km across, 4.5 km in the vertical. Looking down and northwest at the mesocyclone in the southwest corner and along the forward flank gust front.
Figure 4.28: 3D subdomain of WRF 04. Cyan isosurface is $0.0122 \text{ s}^{-1}$ vertical vorticity, and red and dark blue isosurface is $0.0335 \text{ s}^{-1}$ 3D vorticity color-shaded by vertical velocity in ms$^{-1}$ (red is positive vertical velocity, blue is negative). A) 3 hours 25 minutes. B) 3 hours 31 minutes. C) 3 hours 39 minutes. MVT is 3 hours 45 minutes. Looking down and to the northwest at the forward flank gust front. Vertical vorticity associated with mesocyclone is located in to the far southwest. Subdomain is 35 km across and extends to 4 km AGL.
Figure 4.28 (cont).

Figure 4.29: Same as figure 4.26, but for weak case WRF_03. T= 4 hours 11 min, MVT.
Figure 4.30: WRF_03 at 3 hours 47 minutes, 24 minutes prior to MVT. Color shading is plan view of simulated reflectivity at the surface and white lines are 0.002 s$^{-1}$ vertical vorticity at the surface. Subdomain is 40 km x 30 km. North is up.
Figure 4.31: Plan view of WRF_03 at 3 hours 47 minutes. The surface is color shaded by divergence (warm colors) and convergence (cool colors). Red lines are 0.002 s$^{-1}$ surface vertical vorticity contours, and arrows are winds at the surface. Subdomain is 16.2 km x 20.5 km.
Figure 4.32: WRF 04 at 3 hours 25 minutes, 20 minutes prior to MVT. Color shading is plan view of simulated reflectivity at the surface and white lines are 0.005 s$^{-1}$ vertical vorticity at the surface. Subdomain is 35 km x 40 km. North is up.
Figure 4.33: Plan view of WRF_04 at 3 hours 25 minutes, 20 minutes prior to MVT. The surface is color shaded by divergence (warm colors) and convergence (cool colors). Red lines are 0.005 s$^{-1}$ surface vertical vorticity contours, and arrows are winds at the surface. North is up. Subdomain is 18.3 x 24.8 km.
Figure 4.34: Cross section: Vertical velocity in E-W plane across the forward flank gust front of the easternmost storm in WRF_04 at 3 hours 31 minutes, 14 minutes prior to MVT. Color shading corresponds to positive and negative vertical velocities. Pink lines are 200 cm s⁻¹ contours. Surface: Ground wind vectors, with color shading depicting divergence (warm colors) and convergence (cool colors). Looking north. 13 km across x 3.5 km AGL.
Figure 4.35: Looking northwest at the leading edge of the forward flank gust front in WRF_04 at 3 hours 31 minutes, 14 minutes before MVT. SW-NE oriented vertical cross-section of 3-D vorticity is shown, 3 km west of the forward flank gust front. Surface divergence and wind as in Figure 4.33. Red lines are 0.05 s$^{-1}$ contours of surface vertical vorticity. Bulls-eye in vorticity represents the mesocyclone. Cross-section is 23 km across x 3 km AGL.
Figure 4.36: Plan view of WRF_04 at 3 hours 10 minutes, 35 minutes prior to MVT. Color shading is surface temperature and arrows are winds at the surface. North is up. Subdomain is 23.7 km x 32.4 km.
Figure 4.37: Reflectivity vertical cross section at 3 hours 21 min in WRF_04. Cross-section is oriented southwest to northeast and is located approximately 5 km west of the forward flank gust front boundary. Dark blue lines are 1.0 gkg⁻¹ hail mixing ratio contours and light blue lines are 0.5 gkg⁻¹ rain water mixing ratio contours. Looking west at the leading edges of the rear-flank and forward flank downdrafts. Cross section is 30 km x 15 km.
Figure 4.38: Same as Figure 4.36, except dark blue lines are 5 dBz reflectivity contours.
Figure 4.39: Reflectivity as a function of lapse rate for 43 occurrences of microbursts observed in the JAWS field campaign. (Srivastava 1985)
Figure 4.40: Surface reflectivity at 3 hrs and 1.0 gkg\(^{-1}\) white hail mixing ratio contours at 3 km AGL. A) WRF_04 – strong two-cell case B) Control C) WRF_03 – weak two-cell case.
Figure 4.41: WRF_04 at 3 hrs 20 minutes, 25 minutes prior to MVT. Red isosurface is 0.012 s\(^{-1}\) vertical vorticity surface is color shaded by divergence (warm colors are divergence, cool colors are convergence. Arrows are wind at the surface. Looking west northwest at the forward flank gust front. Picture scale is 24.8 km x 2.5 km in the vertical. Southern-most vertical vorticity center is the mesocyclone, minutes from ingesting a discrete rotation center undergoing significant stretching.
CHAPTER 5
CONCLUSIONS

The aim of this study was to determine how the orientation of an initial cell pair, an external forcing, modulated storm-internal forcings such as downdrafts and baroclinic vorticity generation. Results from a suite of 52 idealized numerical storm interaction simulations, one isolated storm control case and 51 two-cell simulations, revealed striking differences in storm morphology based on the initial orientation of the interacting cell pair. Fifty-one of the fifty-two storm interaction simulations produced stronger low-level mesocyclones than the single-cell control case.

Only two storm interaction simulations produced a surface rotation greater than 0.1 s\(^{-1}\), and IN both initially located nearly just south and due west of the control cell. The consistency in this configuration to produce the two strongest storms within the parameter space of this study suggests the relationship could be physical, as opposed to random. However, storm rotational intensity is also found to be highly sensitive to the initial orientation of the interacting storms. Moving the second initial cell in strong case WRF_04 just 8 km east produced a storm of half the intensity in terms of rotation at the surface.

A key storm-internal process affected by the orientation of the initial cell pair is discrete rotation centers generated along the forward flank gust front that subsequently merge with the mesocyclone, leading to low-level mesocyclogenesis. Baroclinically-generated horizontal vorticity is tilted vertically by a downburst behind the forward flank gust front. Momentum associated with the burst enhances convergence along the forward flank gust front, tilts horizontal vorticity upward, and the positive vertical vorticity member is stretched preferentially.
by updrafts associated with the wind convergence between outflow and inflow air along the forward flank. The degree to which these discrete rotation centers intensify in rotation prior to merging with the mesocyclone is modulated by their lifespan propagating along the forward flank gust front, as well as the temperature gradient across the forward flank boundary. In WRF_04, the strong two-cell case that achieved $> 0.1 \text{s}^{-1}$ rotation, a rotation center was stretched by a factor of 13 prior to merging with the mesocyclone. It traveled approximately 20 km along the forward flank gust front with a 5.0-6.0°C temperature difference. Discrete rotation centers in the weaker control case traveled less than 10 km along boundaries with only 1.0-2.0°C temperature differences between air and rain-cooled air within the forward flank.

The orientation of the boundary along which these discrete rotation centers form also plays a role in the extent to which low-level mesocyclogenesis occurs. The forward flank gust front boundary in the particularly strong case, WRF_04, had a 45° orientation relative to the mesocyclone. Tilting and stretching of these discrete rotation centers occurred to the greatest extent along forward flank gust front boundaries that achieved this orientation for the entire suite of simulations. In the weak two-cell case, WRF_03, the rotation centers merging with the mesocyclone came from behind, rather than along a southwest-northeast oriented gust front.

The unsteady nature of the forward flank downdraft is the driving mechanism for tilting the baroclinically-generated horizontal vorticity along the forward flank gust front into the vertical. Wet-microburst-type features occur frequently in the forward flank downdraft, caused by melting hail at 3 km AGL in these simulations. Precipitation loading and latent cooling leads to a microburst that descends to the surface, generating divergence in the surface wind field, air approximately 2.0°C warmer than the surrounding cold pool air, and straight-line winds ranging from 5-20 ms$^{-1}$. In WRF_04, a downburst occurred just behind the forward flank gust front and
tilting of horizontal vorticity produced several discrete rotation centers as a result. In WRF_03, a downburst did not occur within a few kilometers of a forward flank gust front, and substantially less horizontal vorticity was tilted vertically.

The timing and location of the wet-microburst pulses generating discrete rotation centers that are responsible for low-level mesocyclogenesis are modulated by the spatial distribution of hail aloft. In turn, the spatial distribution of hail aloft is modulated by the initial orientation of the interacting cell pair. Future work will focus on understanding how the initial cell pair orientation affects the distribution of hail aloft and the orientation of the forward flank prior to mesocyclogenesis.

Because these discrete rotation centers often have a 20-30 minute lifespan prior to merging with the low-level mesocyclone, increased awareness and better detection could aid nowcasters in anticipating storm intensification events. These rotation centers are initially < 1 km deep and would be passed over by a base elevation scan if the storms were not in close proximity to the radar. Dual-polarimetric radar could potentially be useful in detecting the wet microburst features in the forward flank responsible for generating the rotation centers along the gust front.
REFERENCES


