ABSTRACT

FRENCH, ADAM JAMES. Squall Line Evolution in Response to a Developing Nocturnal Low-Level Jet and Mergers with Isolated Supercell Thunderstorms. (Under the direction of Matthew D. Parker.)

Squall lines are a fairly ubiquitous feature around the globe, that can significantly impact society both by bringing beneficial rainfall, but also by producing a wide variety of hazardous weather. Given these potentially significant societal impacts, it is important to understand not only when and where squall lines may form, but also how squall lines may evolve. The present study addresses a portion of this problem by investigating how squall lines evolve in two complex, yet commonly observed scenarios: in the presence of a developing nocturnal low-level jet, and following mergers with isolated supercell thunderstorms.

In the first part of this study, the impacts of a developing low-level jet on a mature squall line are investigated using idealized numerical simulations. These simulations are designed to mimic the environmental transition that occurs as night falls and the boundary layer stabilizes, while also including a gradually developing low-level wind maximum. The characteristics of the simulated LLJ atop a simulated stable boundary layer are based on past climatological studies of the LLJ in the central United States. A variety of jet orientations are tested, and sensitivities to jet height and the presence of low-level cooling are explored. The primary impacts of adding the LLJ are that it alters the wind shear in the layers just above and below the jet, and that it alters the magnitude of the storm-relative inflow in the jet layer. The changes to wind shear have an attendant impact on low-level lifting, in keeping with current theories for gust front lifting in squall lines. The changes to the system-relative inflow, in turn, impact total upward mass flux and precipitation output. Both are sensitive to the squall line-relative orientation of the LLJ, and change in time as the low-level cooling progresses.
The second part of this study is focused on identifying features and storm evolutions common in cases of mergers between squall lines and isolated supercells. A set of 21 cases wherein an isolated supercell merged with a squall line were identified and investigated using analyses from the Rapid Update Cycle (RUC) model, archived data from the WSR-88D network, and severe storm reports. This analysis revealed two primary environments in which these mergers occur: a weak synoptic forcing, weak-moderate shear environment (WF) and a strong synoptic forcing, strong shear environment (SF). Radar reflectivity data revealed a spectrum of storm evolutions across these two environments that generally lead to the merged system organizing as a bow echo. In both environments, storm rotation generally weakened and became concentrated in low-levels following the merger, although the details of this evolution varied. Finally, storm reports from both cases showed a peak in tornado production coincident with merger time in the WF environment, and immediately prior to the merger in the SF environment. Severe wind reports were most prevalent post-merger in both environments. This suggests that as storm organization changes following a squall line-supercell merger, so too does the severe weather threat.

The third part of this study builds on the observations of squall line-supercell mergers by using idealized numerical simulations to investigate the storm-scale processes that drive these events. A pair of idealized simulations run using the same background model configuration and environment are compared. The first consists of a squall line evolving in isolation, and the second consists of a squall line-supercell merger. The merger simulation is shown to capture features representative of one of the common evolutionary archetypes identified in observations. The merger and no-merger squall line are compared and it is found that while both systems evolve into bow echoes, the bowing structure associated with the merger squall line is more compact and develops further south. Additionally, the merger squall line is found to produce stronger surface winds, heavier rainfall and stronger low-level vertical vorticity than
either the no-merger squall line or the isolated supercell. A detailed examination of the cold pool evolution in the merger simulation revealed that interactions between the squall line’s gust front and outflow associated with the isolated supercell result in the squall line weakening north of the merger, as was frequently observed. Post-merger, strong downdrafts driven by enhanced rainfall associated with the merged supercell strengthen the cold pool south of the merger, and enhance surface winds through vertical momentum transport. The strong low-level vertical vorticity found in the merger simulation appears driven by tilting of horizontal vorticity from behind the squall line’s gust front. This enhanced low-level vorticity is consistent with observations of strong low-level rotation following these types of mergers.
© Copyright 2011 by Adam James French

All Rights Reserved
Squall Line Evolution in Response to a Developing Nocturnal Low-Level Jet and Mergers with Isolated Supercell Thunderstorms

by

Adam James French

A dissertation submitted to the Graduate Faculty of North Carolina State University in partial fulfillment of the requirements for the Degree of Doctor of Philosophy

Marine, Earth and Atmospheric Sciences

Raleigh, North Carolina

2011

APPROVED BY:

Anantha Aiyyer
Gary Lackmann

Sandra Yuter
Matthew D. Parker
Chair of Advisory Committee
DEDICATION

To my wife, Miranda.
BIOGRAPHY

Adam French was born and raised in Manchester, Connecticut where he got to experience a wide variety of weather, from summer thunderstorms to winter nor’easters. This sparked an interest in the weather at an early age, however it was not until he wrote his 11th grade honors thesis on severe weather in Connecticut that Adam decided he wanted to become a meteorologist.

Adam attended Valparaiso University in northern Indiana from 2001-2005, where he received his B.S. in meteorology. While at Valpo, Adam was also a member of Christ College, an interdisciplinary honors college, from which he earned a minor in humanities, and graduated as a Christ College scholar. In the summer of 2004, Adam participated in the National Weather Center Research Experience for Undergraduates in Norman, Oklahoma where he analyzed data from the SPC/NSSL Spring Program. During his senior year at Valpo, Adam served as president of the Northwest Indiana Chapter of the National Weather Association and played a main role in the organization and running of the 3rd Annual Great Lakes Meteorology Conference. Following graduation from Valparaiso in 2005, Adam came to North Carolina State University and began pursuing a Masters degree as one of the inaugural members of the Convective Storms Group under Dr. Matthew Parker. He completed his Masters degree in the fall of 2007, and promptly began working on his Ph. D. Adam participated in the VORTEX2 field experiment in the springs of 2009 and 2010, and plans to begin a position as an Assistant Professor at the South Dakota School of Mines and Technology following the completion of his Ph. D. in August 2011.

In his spare time, Adam enjoys hiking and camping with his wife Miranda and hound dog Dixie. He is also an avid photographer, and enjoys getting out to take pictures whenever his schedule allows.
ACKNOWLEDGEMENTS

I would first like to thank my advisor and committee chair, Dr. Matt Parker for taking me on as a Masters student back in 2005, and for his continued guidance, support, and encouragement all along the way. I’d also like to thank my committee, Drs. Sandra Yuter, Gary Lackmann, and Anantha Aiyer for all of their input and help through the course of this project. To the members of the Convective Storms Group, past and present, including Jerilyn Billings, Ben Baranowski, Casey Letkewicz, Mike Kiefer, Billy Booth, Matt Morin and Johannes Dahl; thank you all for your helpful input and discussions, not to mention friendship throughout my time here at NC State. Constructive comments from Stan Trier, Russ Schumacher and Alexandre Fierro helped to greatly improve the material presented in Chapter 2. This work would not have been possible without the excellent cloud model (CM1) created and maintained by George Bryan at NCAR, or without the high performance computing resources provided by the Office of Information Technology High Performance Computing at North Carolina State University, the Renaissance Computing Institute in Chapel Hill, NC, and the Computational and Information Systems Laboratory at NCAR. I would also like to thank the developers of the WDSS-II software for maintaining and providing this software free of charge for research purposes. Thanks too, to the National Science Foundation for funding this research, and my graduate education, through grants ATM-0552154 and ATM-0758509.

Finally, I need to thank my family. Thank you to my parents for your un-yielding love, support and encouragement in all of my endeavors since before I was in grade school through now. It means so much to know I always had your support. Thank you too, to my sister Molly, for your encouragement, and for inviting us down to visit in Savannah whenever I needed a break. It is also safe to say that that I would not have made it this far without the love and support of my amazing wife, Miranda. Thank you baby, I love you. And last, but certainly
not least, I have to thank Dixie the hound dog, whose snarl-faced, tail wagging, exuberance greeting me at the door every night always managed to make up for a lousy day, no matter how many times the model crashed or my codes didn’t work.
# TABLE OF CONTENTS

List of Tables .................................................................................. viii
List of Figures ............................................................................... ix

**Chapter 1 Introduction** ................................................................. 1
1 Background ................................................................................. 3
2 Organization of the following chapters ........................................... 6

**Chapter 2 The response of simulated nocturnal convective systems to a developing low-level jet** ................................................... 9
1 Introduction .................................................................................. 9
  a Background ............................................................................... 11
2 Methods .................................................................................... 13
3 Benchmark simulations ................................................................. 18
  a Overview of simulations ............................................................. 18
  b Role of changes to the low-level wind shear .............................. 20
  c Role of changes to the storm relative inflow ............................. 26
4 Sensitivity tests ........................................................................... 29
  a Simulations without low-level cooling ...................................... 29
  b 500m jet height ........................................................................ 30
5 Non-periodic 3D simulations ........................................................ 32
6 Discussion ................................................................................... 34
7 Conclusions ................................................................................ 37

**Chapter 3 Observations of mergers between squall lines and isolated supercell thunderstorms** ................................................. 57
1 Introduction .................................................................................. 57
  a Background ............................................................................... 58
2 Data and Methods ....................................................................... 60
3 Results ....................................................................................... 63
  a Overview .................................................................................. 63
  b Background environment .......................................................... 65
  c Reflectivity Analysis ................................................................. 68
  d Velocity analysis ....................................................................... 72
  e Storm Reports .......................................................................... 76
4 Conclusions and future work ........................................................ 79
## Chapter 4  Idealized numerical simulations of a squall line-supercell merger  ...  100

1  Introduction ................................................. 100
2  Idealized simulation set-up  ................................ 102
3  Overview and comparison of the basic simulations  ........ 106
   a  Overview of the NOMERGER, SUPE, and MERGER simulations ... 106
   b  Impacts of merger on storm-scale structure  ................ 108
4  Cold pool and and low-level vorticity evolution ........... 113
   a  Cold pool evolution  ....................................... 113
   b  Low-level vorticity evolution  ............................ 118
5  Sensitivity tests ............................................. 123
   a  Sensitivity to merger location  ............................ 123
   b  Additional sensitivity tests  ............................... 126
6  Conclusions and future work  ............................... 128

## Chapter 5  Conclusions and future work  ........................ 167

References ......................................................... 173

Appendix .......................................................... 185
   Appendix A .................................................... 186
LIST OF TABLES

Table 2.1  Vector wind differences between 0 - 2.5 km, 0 - 1 km, 1 - 2.5 km, and 0.5 - 1.5 km AGL for CTL, RTF and FTR simulations at t = 8:00. . . . . 56

Table 3.1  Details for each merger event, letters following the date denote cases with multiple mergers. Time, lat. and lon. refer to the approximate time and latitude/longitude that the merger occurred, environ. refers to the two synoptic environments discussed in 3b, and type refers to the observed post-merger evolutions discussed in 3c with SSB, EMB, HYB, and OTR corresponding to the ”system-scale bowing”, ”embedded-bowing”, ”hybrid” and ”other” evolutions, respectively. . . . . . . . . . . . . . . . 98

Table 3.2  Number of merger events categorized by convective organization (SSB, hybrid, EMB and other) and background environment. The total number of merger events is larger than the total number of cases owing to some cases producing multiple merger events. . . . . . . . . . . . . . . . . . . . . . . . . . . . . 99
LIST OF FIGURES

Figure 1.1  Schematic vertical cross section of a squall line with trailing stratiform precipitation (from Houze et al. 1989). 8

Figure 1.2  Cross section through a multicell storm over the course of 20 minutes. Note the new cells developing on the left flank of the storm (cell 5), while old cells are dissipating to the right (cell 1) (from Doswell 1985). 8

Figure 2.1  Skew-T ln-p diagram of the base state temperature and dew point (thick black lines) and wind (barbs, half and full barbs = 2.5 and 5 ms$^{-1}$, respectively) profiles for all simulations. 38

Figure 2.2  (a) Time-series of perturbation u-wind profiles illustrating the development of the simulated LLJ. (b) Schematic illustrating the different LLJ orientations. The wind profiles in (a) are from the rear-to-front (RTF) simulation. The other jet orientations have an identical shape and magnitude, but a different direction. 39

Figure 2.3  Vertical cross sections of simulated radar reflectivity (dBZ, shaded as shown) for the CTL (a,b,c), RTF (d,e,f), and FTR (g,h,i) simulations at (from left to right) t= 6:00, 8:00, and 10:00. 40

Figure 2.4  Depiction of the CTL simulation over time: (left) values vs time for the reference environmental temperature (Tref) and the minimal surface temperature on the domain (Tmin), using line styles as shown; (center) Hovmöller diagram for 5 km AGL, with along-line maxima in tracer concentration (shaded as shown) and vertical velocity (contoured at 10, 15, 20, and 25 m s$^{-1}$); (right) vertical profile of environmental CIN (shaded as shown) and CAPE (contoured at 500, 1000, and 2000 J kg$^{-1}$, with bold contour at 0) vs time. The ordinate for all three panels is the same (time). 41

Figure 2.5  Time series of (a) total upward mass flux (TUMF, kg), and (b) maximum of along-line averaged vertical velocity ([w]$_{max}$ ms$^{-1}$) from t = 3:00-10:00 for the RTF (gray solid), FTR (dashed) and CTL (black solid) simulations. 42
Figure 2.6 Assorted fields for (a) CTL, (b) RTF and (c) FTR simulations at t = 5:30. Center panel contains vertical cross section of along-line averaged vertical velocity ($w$, contoured, dashed contours $> 5 \text{ m s}^{-1}$) and potential temperature perturbation ($\theta'$, alternately shaded every 2 K below -2 K) and wind vectors within the developing stable layer and cold pool ($\text{m s}^{-1}$, scale vector at upper right). Right hand panel is the along-line averaged u-wind profile at $x = 390 \text{ km}$ (well ahead of the squall line). Left hand panel is an along-line averaged buoyancy perturbation (relative to the evolving pre-line environment) 1 km behind the system gust front.

Figure 2.7 As in Fig. 2.6, but at t = 7:30.

Figure 2.8 As in Fig. 2.6, but at t = 8:30.

Figure 2.9 Time-height plots of along-line averaged (a) CAPE ($\text{J kg}^{-1}$), (b) CIN, and (c) cold pool buoyancy perturbation ($B'$, $\text{m s}^{-2}$) between t = 3:00 and 10:00 for CTL simulation. CAPE and CIN are measured at $x = 398 \text{ km}$ (well ahead of the squall line), while $B'$ is calculated within the cold pool (relative to the evolving pre-line environment), 1 km behind the gust front. The gray shaded area represents the effective inflow layer discussed in the text.

Figure 2.10 Time series of cold pool strength ($C$, $\text{ms}^{-1}$ heavy dashed line, control simulations only) and square root of the vertically integrated horizontal vorticity flux ($u|\eta|_z$, thin solid line for RTF jet configurations and thin dashed line for FTR jet configurations) for (a) benchmark, (b) NOCOOL, and (c) 513m simulations. The square root of $u|\eta|_z$ is shown to make the units comparable to $C$ and both variables are integrated over the effective inflow layer shaded in Fig. 2.9.

Figure 2.11 Vertical cross section of along-line averaged potential temperature ($\theta$, contours every 2 K starting at 294 K), wind vectors (m s$^{-1}$, scale vector in lower right corner) and CAPE ($\text{J kg}^{-1}$ shaded as shown) and vertical profiles of LFC height (m, left-hand panel) for (a) RTF and (b) FTR simulations at t = 8:00. Heavy black lines denote the top and bottom of the applied LLJ.

Figure 2.12 Time series of vertically integrated system-relative horizontal mass flux for (a) benchmark, (b) NOCOOL, and (c) 513m simulations. The integration is performed from the bottom of the effective inflow layer shaded in Fig. 2.9 to 3 km AGL in (a) and (c) and from the surface to 3 km AGL in (b). An upper bound of 3 km is chosen as it generally represents the vertical extent of air parcels containing CAPE.
Figure 2.13  Vertical cross section of parcels trajectories plotted over 1 hour between $t = 8:00$ and $9:00$ for the (a) CTL, (b) RTF and (c) FTR simulations. The values reported below the figure label refer to the total number of parcels that passed through the updraft region of the respective squall lines during this time interval. Only a subset of these totals are plotted in the interest of clarity. ...................................................... 50

Figure 2.14  As in Fig. 2.5, but for NOCOOL simulations. ...................................................... 51

Figure 2.15  As in Fig. 2.5, but for FTR513, RTF513 and CTL simulations. .................. 52

Figure 2.16  As in Fig. 2.6, but for CTL, RTF513 and FTR513 simulations at $t = 8:30$. ...................................................... 53

Figure 2.17  Difference plot of across-line (east/west) averaged accumulated rainfall (cm) comparing PARNP and CTLNP simulations (PARNP-CTLNP). The data are grouped into 30 km wide (north/south) bins that are centered on the points shown along the horizontal axis. The dashed vertical line denotes the center of the squall line at $y = 300$ km. .................. 54

Figure 2.18  Schematic diagram illustrating the key layers of vertical wind shear participating in the cold pool/shear balance relative to the height of the maximum LLJ winds ($Z_{LLJ}$ heavy dashed line) and cold pool depth ($Z_{cp}$) over time. Thin lines are representative isentropes that define the cold pool/bore and the shaded area represents the layer of vertical shear. The evolution is broken into three periods corresponding to different points in the squall line’s evolution. .................. 55

Figure 3.1  Scatter plot of RUC analysis maximum sea-level pressure gradient (Pa/km) compared to maximum 500 mb height gradient (m/km) for (blue dots) weakly forced cases and (red triangles) strongly forced cases. Maximum values were computed over a 1400 x 1400 km domain centered on the merger. ...................................................... 82

Figure 3.2  Distributions of mean values of RUC analysis (a) sea-level pressure gradient (Pa km$^{-1}$), (b) mean 500 hPa height gradient (m km$^{-1}$), (c) mean 0-6 km shear (m s$^{-1}$), and (d) mean CAPE (j kg$^{-1}$). Means were computed over a 1400 x 1400 km area centered on the merger. Dashed red and blue boxes denote the portions of the distributions used to compute the means in Figs. 3.3-3.6. ...................................................... 83

Figure 3.3  Mean RUC analysis 500 hPa height (contours, m), wind barbs (flag = 25 m s$^{-1}$, full barb=5 m s$^{-1}$, half barb= 2.5 m s$^{-1}$), and wind speed (shaded, m s$^{-1}$) for (a) WF and (b) SF environments. The black “X” marks the mean merger location. ...................................................... 84
Figure 3.4 Mean RUC analysis sea level pressure (hPa, solid contours), surface temperature (C, shaded as shown), dew point temperature (C, dashed contours) and wind barbs (flag = 25 m s$^{-1}$, full barb=5 m s$^{-1}$, half barb= 2.5 m s$^{-1}$), for (a) WF and (b) SF environments at the time of merger. Approximate locations of the surface warm front, cold front, and dry line are annotated with traditional symbols. The black “X” marks the mean merger location.

Figure 3.5 Mean RUC analysis surface-based CAPE (shaded as shown, J kg$^{-1}$) and CIN (contours, J kg$^{-1}$), for (a) WF and (b) SF evolutions at the time of merger. The black “X” marks the mean merger location.

Figure 3.6 Mean RUC analysis 0-1 km AGL storm-relative helicity (shaded as shown, m$^2$s$^{-2}$), 0-6 km AGL bulk shear magnitude (contours, m s$^{-1}$) and bulk shear (wind barbs, flag = 50 m s$^{-1}$, full barb = 10 m s$^{-1}$, half barb = 5 m s$^{-1}$), for (a) WF and (b) SF evolutions at the time of merger. The black “X” marks the mean merger location.

Figure 3.7 Schematic diagrams illustrating the (a) SSB evolution (b) EMB and (c) hybrid evolutions as they would appear on radar (gray shading denotes higher radar reflectivity values). The dashed arrows at T=1 represent initial supercell motion vectors.

Figure 3.8 Examples of (a-e) the system-scale bowing evolution in a WF environment and (f-j) the embedded bowing evolution in a SF environment. Data are WSR-88D 0.5° tilt radar reflectivity from (a)-(e) Fort Worth, Texas (KFWS) between 2333 5 May 1995 and 0233 6 May 1995 and (f)-(j) Little Rock, Arkansas (KLZK) between 0205 and 0339 UTC 5 February 2008.

Figure 3.9 Examples of the hybrid evolution in (a-e) a WF environment and (f-j) a SF environment. Data are WSR-88D 0.5° tilt radar reflectivity from (a) Amarillo, Texas (KAMA) at 0302 UTC 16 May 2003, (b-e) Frederick, Oklahoma (KFDR) from 0335-0533 UTC 16 May 2003, and (f-j) Topeka, Kansas (KTWX) between 2203 UTC 23 March 2009 and 0008 UTC 24 March 2009.

Figure 3.10 Maximum azimuthal shear (s$^{-1}$ shaded as shown) accumulated over time to produce rotation tracks associated with the supercell and merged system from (a) the 18 April 2009 WF case, (b) supercell 1 in the 10 November 2002 SF case, (c) the 30 May 2008 WF case, (d) supercell 4 in the 6 February 2008 SF case. The vertical dashed black lines indicate the longitude of the merger in each case.
Figure 3.11 Time versus height plots of maximum azimuthal shear (contoured, (s\(^{-1}\)), color scheme on right side of figure) associated with the isolated supercell (pre-merger) and merged system (post-merger) for (a) 6 May 1995, (b) 21 April 2007, (c) 9 May 2008, (d) 29 May 2008 (merger 1), (e) 18 April 2009, and (f) 6 May 2009 WF cases. Time is in a merger-relative framework, with t=0 corresponding to the merger time, which is annotated with a vertical black line.

Figure 3.12 Constant-height, ground-relative WSR-88D velocity data (shaded as shown m s\(^{-1}\)) and 45 dBZ radar reflectivity contour from: (a-d) Frederick, Oklahoma 15 May 2003 at 3 km AGL, (e-h) Vance Air Force Base, Oklahoma, 18 April 2009 at 2 km AGL, (i-l) Wichita, Kansas, 25 May 2008 at 3 km AGL and (m-p) Topeka, Kansas 23 March 2009 at 2 km AGL. In all panels dashed circles represent approximate diameters of the circulation features initially associated with the pre-merger supercell. Black arrows point toward the radar location in panels where it is outside the plotting area, otherwise the radar location is denoted by a black “X”. Radar reflectivity from the cases in (a-d) and (m-p) is also presented in Figs. 3.9a-e and 3.9f-j respectively.

Figure 3.13 (a-d) Constant-height, ground-relative WSR-88D velocity data (m s\(^{-1}\), positive (outbound) values shaded as shown, negative (inbound) values contoured using the same color scheme) and (e-f) radar reflectivity from Fort Worth, Texas (KFWS) at 3 km AGL on from 0054-0227 UTC 6 May 1995. In all panels dashed circles represent the approximate diameters of the circulation initially associated with the pre-merger supercell (labeled SC) and subsequent mesovortices (labeled MV). White (black) arrows point toward the radar location in panels b-d (f-h) where it is outside the plotting area, otherwise the radar location is denoted by a black “X”.

Figure 3.14 As in Fig. 3.11, but for the (a) 10 November 2002 (merger 3), (b) 5 May 2007, (c) 5 February 2008 (merger 3), (d) 10 February 2009, (e) 23 March 2009, and (f) 26 April 2009 SF cases.

Figure 3.15 Storm reports vs time for (a) WF and (b) SF environments. Black dots represent all reports, red diamonds tornado reports, green triangles hail reports and blue asterisks wind reports. Colored lines correspond to a 6th-order polynomial trend line fit to the respective data. The vertical gray line denotes the merger time.

Figure 3.16 Enhanced Fujita (EF) scale rating (triangles) and path length (squares) vs time for tornadoes in all merger cases (WF and SF). The black and gray lines represent a 4th-order polynomial trend line fit to the EF scale and path length reports, respectively. The vertical gray line denotes the merger time.
Figure 4.1  Skew-T log-P diagram of temperature, dew point temperature and lifted parcel path, wind profiles and hodographs (kts) for (a) base-state squall line environment and (b) the supercell environment added 3 hours into the simulation. Color coding on the hodograph denotes various height layers as follows: green: 0-1 km, blue: 1-2 km, red: 2-3 km, yellow 3-6 km, and black 6km- model top. Wind barbs are in knots with a half barb = 5 knots, full barb = 10 knots, and flag = 50 knots.

Figure 4.2  Simulated radar reflectivity (dBZ), 1 km AGL winds (m s$^{-1}$, representative vector at the lower-right), and -1 K surface potential temperature perturbation between t = 125 and 245 minutes for (a)-(e) a squall line simulation using the initial base-state environment and (f)-(j) a squall line simulation wherein the environment is modified after 3 hours, as discussed in the text.

Figure 4.3  Simulated radar reflectivity (dBZ, shaded as shown) and 1 km AGL winds (m s$^{-1}$, representative vector at lower right of panels e and f) at (a) 200 (b) 245 (c) 265 (d) 280 (e) 300 and (f) 320 minutes into the NOMERGER simulation.

Figure 4.4  As in Fig. 4.3, but for the SUPE simulation. Note that the times labels reflect storm initiation at t = 180 min., to ease comparison with the MERGER simulation. Also, a smaller plotting window is used to zoom in on the supercell.

Figure 4.5  As in Fig. 4.3, but for the MERGER simulation.

Figure 4.6  As in Fig. 4.5, but zoomed in on the merger with data every 10 minutes between t = 255 and t = 310 min.

Figure 4.7  Swaths of maximum vertical vorticity (s$^{-1}$, shaded as shown) at 1 km AGL (top panels), 3 km AGL (middle panels) and 6 km AGL (bottom panels) accumulated between 3 and 6 hours into the simulation for the a, c, e MERGER and b, d, f NOMERGER simulations. The fields are accumulated by taking the maximum value over time across the subset of the model domain shown. The vertical dashed line in each panel denotes the approximate X-position of the merger. Dashed ovals highlight features of interest discussed in the text.

Figure 4.8  As in Fig. 4.7 but for top panels: minimum surface potential temperature perturbation (K, shaded as shown). Middle panels: 1 km AGL vertical velocity (m s$^{-1}$, shaded as shown). Bottom panels: 5 km AGL vertical velocity (m s$^{-1}$, shaded as shown).
Figure 4.9 Difference plot comparing the surface potential temperature perturbation (K, shaded as shown) in the MERGER and NOMERGER simulations (MERGER-NOMERGER). The sold and dashed black contours indicate the positions of the -1 K potential temperature contours, representing the gust front location for the MERGER and NOMERGER simulations, respectively.

Figure 4.10 As in Fig. 4.7 but for maximum wind speed (top panels, m s\(^{-1}\), shaded as shown) and vertical vorticity (middle panels, s\(^{-1}\), shaded as shown) at the lowest model level, and rainfall rate (bottom panels, mm hr\(^{-1}\), shaded as shown).

Figure 4.11 Evolution of gust front lifting in the MERGER simulation. (a-e) magnitude of the surface potential temperature gradient (K km\(^{-1}\), shaded as shown) and -1 K theta perturbation (dashed contour). (j-g) 1 km AGL vertical velocity (m s\(^{-1}\), shaded as shown and -1 K surface theta perturbation (dashed contour). Arrows and dashed ovals denote features of interest discussed in the text.

Figure 4.12 Evolution of storm-relative inflow at between t = 230 minutes and t= 260 minutes (left-to-right) in the MERGER simulation. Top panels: plan-view of surface equivalent potential temperature (K, shaded as shown), surface potential temperature perturbation (contoured at -2 (black), -4 (dark blue), -6 (medium blue) and -8 (purple) K, and squall line-relative surface winds. Bottom panels: x-z cross sections of the same fields in as in (a-c) taken along the black lines shown in (a-c).

Figure 4.13 Surface potential temperature perturbation (K, shaded as shown) and surface winds greater than 35 m s\(^{-1}\) (representative vector below color bar) for the (top panels) MERGER simulation and (bottom panels) NOMERGER simulation at t = 265, 275 and 295 minutes (left to right).

Figure 4.14 Time series of maximum cold pool strength, c (m s\(^{-1}\)) for the MERGER (solid black) and NOMERGER (dashed red) simulations between t = 190 and 360 minutes. The approximate merger time is denoted by the vertical dashed line. The maximum value is calculated between y = 120 and 180 km, which encompasses the region of maximum cold pool intensity in associated with the post-merger bow echo in the MERGER simulation.

Figure 4.15 As in Fig. 4.13, but for downward momentum flux (shaded as shown, m\(^2\) s\(^{-2}\)), surface rain rate (contoured every 50 mm hr\(^{-1}\) starting at 50), surface winds greater than 35 m s\(^{-1}\) (representative vector below color bar), and -1 K surface potential temperature perturbation.
Figure 4.16  Vertical (x-z) cross sections of horizontal wind speed (m s\(^{-1}\), shaded as shown) and w (contoured at -1, -5 and -10 m s\(^{-1}\)) along y = 165 km for the (top panels) MERGER and (bottom panels) NOMERGER simulations at (left-right) t = 270, 280, 290 minutes.

Figure 4.17  As in Fig. 4.13, but showing w (values < 0 shaded as shown, m s\(^{-1}\)), surface rain rate (contoured every 50 mm hr\(^{-1}\) starting at 50) and -1 K surface potential temperature perturbation.

Figure 4.18  Time vs. height plot of maximum vertical vorticity (s\(^{-1}\)) associated with the supercell pre-merger and merged system post-merger. The vertical dashed line denotes the merger time.

Figure 4.19  500 m AGL vertical vorticity (s\(^{-1}\), shaded as shown) and 40 dBZ simulated radar reflectivity contour for the MERGER simulation at the times shown in the upper right of each panel. Black arrows in each panel identify the vorticity associated with the isolated supercell and subsequent merged system.

Figure 4.20  (a) 500 m AGL maximum vertical vorticity (s\(^{-1}\)) and (b) 1 km AGL maximum vertical velocity for (solid black) MERGER and NOMERGER (dashed red) simulations between t = 240 and t = 325 minutes.

Figure 4.21  Horizontal cross-sections of vertical vorticity (s\(^{-1}\), shaded as shown) and 40 dBZ radar reflectivity contour at 1 km AGL for the SUPE simulation at t = (a) 250, (b) 260, (c) 270, (d) 280, (e) 290, and (f) 300 min.

Figure 4.22  Time series of (a) vertical vorticity and (b) vertical vorticity tendency due to tilting (solid black), stretching (dashed red), east-west flux divergence (dotted green), and north-south flux divergence (alternating dashed blue) integrated over the 12 km x 12 km area described in the text.

Figure 4.23  Cross section of along-line averaged vertical velocity (contoured every -2.5, 2.5, 5,10 m s\(^{-1}\)) and along-line maximum vertical vorticity (shaded as shown, s\(^{-1}\)) at t = (a) 250 (b) 265 (c) 275 (d) 295 min. Values are averaged/maxed between y = 150 and 180 km.

Figure 4.24  Evolution of tilting term in vorticity equation at t = (a,b) 265, (c,d) 280 and (e,f) 295 min. Left panels (a,c,e): horizontal vorticity vectors and magnitude (shaded as shown, s\(^{-1}\) and vertical velocity (contoured at 5, 10 m s\(^{-1}\)). Right panels: tilting term (shaded as shown, s\(^{-2}\) and vertical velocity (contoured at 5 m s\(^{-1}\)).

Figure 4.25  Horizontal cross-sections of vertical vorticity (s\(^{-1}\), shaded as shown) and storm-relative wind vectors at 3 km AGL for the MERGER simulation at t = (a) 275, (b) 290, (c) 305 and (d) 320 min.
Figure 4.26 Plan-view plots of simulated radar reflectivity (shaded as shown, dBZ) and integrated updraft helicity (contoured every 250 m² s⁻²) at t = 235 min. for (a) MERGER simulation, (b) Y150 simulation, (c) Y130 simulation, (d) Y90 simulation. Black arrows identify the supercell in each panel. .................................................. 157

Figure 4.27 As in Fig. 4.26 but at t = 330 min. Black arrows identify the location of the merged supercell in each panel. .................................................. 158

Figure 4.28 As in Fig. 4.27 but showing the surface potential temperature perturbation (shaded as shown, K) and the 40 dBZ simulated radar reflectivity contour. Black arrows denote areas of comparatively weaker cold pool in the vicinity of the merger in each simulation. ......................... 159

Figure 4.29 As in Fig. 4.10 but comparing the MERGER and Y150 simulations. .......................... 160

Figure 4.30 As in Fig. 4.10 but comparing the MERGER and Y130 simulations. ......................... 161

Figure 4.31 As in Fig. 4.10 but comparing the MERGER and Y90 simulations. ......................... 162

Figure 4.32 Time-series of maximum vertical vorticity s⁻¹ calculated at 1 km AGL over 250 km (east-west) by 100 km (north-south) box centered on the merger for (a) MERGER simulation (b) Y150 simulation, (c) Y130 simulation, and (d) Y90 simulation (solid black line). The dashed red line in each panel represents the maximum vorticity calculated over the same region in the NOMERGER simulation. ........................................ 163

Figure 4.33 As in Fig. 4.3, but for the X220 simulation at t = (a) 290 (b) 305 (c) 315 (d) 330 (e) 345 and (f) 360 min. ................................................................. 164

Figure 4.34 As in Fig. 4.33, but showing 500 m AGL vertical vorticity (shaded as shown, s⁻¹) and the 40 dBZ simulated reflectivity contour. Black arrows denote the isolated and eventual merged supercell and red arrows identify squall line meso-vortices. ........................................ 165

Figure 4.35 Simulated radar reflectivity (dBZ, shaded as shown) and -1 K surface potential temperature perturbation for the reverse base-state substitution simulation at t = (a) 190 (b) 210 (c) 230 (d) 250 (e) 270 and (f) 290 min. ................................................................. 166

Figure A.1 As in Fig. 2.4, except for our rerun of the PJ04-289K simulation. This figure is directly comparable to Fig. 11 of P08. .................................................. 189

Figure A.2 As in Fig. 2.4, except for our rerun of the DEEP-unif simulation. This figure is directly comparable to Fig. 15 of P08. .................................................. 190
Chapter 1

Introduction

Deep, moist convection in the atmosphere organizes on scales that range from a single “ordinary” cell (Byers and Braham 1949) up to very large mesoscale convective systems (MCSs, e.g. Zipser 1982) that can include many convective cells and encompass 100s of km$^2$. The present study focuses on a subset of MCSs that are characterized by convective cells organized into a line. These quasi-linear convective systems, or “squall lines” can range in size from 10s to 100s of km, and are a fairly ubiquitous feature around the globe (e.g. Fritsch and Forbes 2001). They can produce both positive and negative impacts for society, bringing vital growing-season rainfall to areas such as the Central United States (Fritsch et al. 1986), as well as hazardous and destructive weather including tornadoes, damaging winds, and flooding (e.g. Fujita 1978; Johns and Hirt 1987;Doswell et al. 1996; Schumacher and Johnson 2005; Gallus et al. 2008).

While much is known about the basic structures and mechanics to squall lines (e.g. the thorough reviews of Fritsch and Forbes 2001 and Houze 2004), the present work is interested in improving the understanding of how these systems evolve in more complex situations. Recent studies have focused on a number of pertinent issues in MCS dissipation (Gale et al. 2002), how squall lines evolve as night falls (Parker 2008), or day breaks (Marsham et al. 2011), and
how rapidly squall lines develop in varying environments (Coniglio et al. 2010). The present work examines two other novel aspects of squall line evolution: how squall lines evolve in the presence of a temporally-varying background environment with a developing low-level jet; and, how squall lines evolve following mergers with isolated supercells.

The first of these topics is of great interest, as it is a fairly common occurrence over the Central United States during the warm season. A number past studies have observed MCSs that form during the afternoon and persist during the overnight hours (e.g. Wetzel et al. 1983; Cotton et al. 1989), and recent work by Parker (2008) has investigated how squall lines evolve in response to a stabilizing boundary layer. However, past work regarding the role of the low-level jet as it pertains to squall lines has generally either focused on the role it plays in priming the convective environment (e.g. Helfand and Schubert 1995; Higgins et al. 1997; Tollerud et al. 2008) through the transport of warm, moist air, or as a potential forcing mechanism through the mesoscale lift the ensues when the jet impacts a frontal zone (e.g. Trier and Parsons 1993; Tuttle and Davis 2006; Trier et al. 2006). To date, and to the author’s knowledge, the role of a developing low-level jet, in altering the *convective-scale evolution* of an on-going squall line has not been investigated.

Similarly, a number of studies have documented cases of mergers between squall lines and isolated supercells (e.g. Goodman and Knupp 1993; Sabones et al. 1996; Wolf et al. 1996; Wolf 1998; Sieveking and Przybylinski 2004), however there has yet to be an investigation of a large number of cases in order to determine the common storm evolutions that can be expected following a merger. Even less clear are the dynamics involved when these two modes interact; very few such studies appear in the literature (e.g. Goodman and Knupp 1993). This problem is important because these types of interactions are often associated with severe weather, including tornadoes (e.g. Goodman and Knupp 1993; Sabones et al. 1996 Sieveking and Przybylinski 2004).
1. Background

This introduction will provide some general background on squall line structures and dynamics. Because there is minimal overlap between the background information on low-level jet interactions and supercell interactions, a detailed discussion of pertinent past research on each will be presented in their respective chapters.

The basic structure of the squall line thunderstorm was first examined by Newton (1950) using data collected during the Thunderstorm Project (Byers and Braham 1949) in the late 1940s. Using upper air and surface data from two squall line cases, Newton (1950) created a cross-sectional schematic that captured many of the key features now known to be common to squall lines. These include a large pool of cold air beneath the system, distinct from any nearby synoptic-scale boundaries, and an apparent inflow of cooler, drier mid-level air descending from behind the line. This study was also one of the first to note the apparent importance of vertical wind shear within the squall line environment, and to hypothesize that squall line longevity is the result of periodic re-triggering of new storms rather than a single persistent cell.

The results of numerous additional observation-based studies since the 1950s (e.g. Fujita 1955; Zipser 1977; Smull and Houze 1985; Smull and Houze 1987a; Johnson and Hamilton 1988) are synthesized into an updated squall line model presented by Houze et al. (1989), included here as Fig 1.1. This schematic distinguishes between the convective precipitation associated with the leading line and the broad region of stratiform precipitation towards the rear of the system (e.g. Smull and Houze 1985). It also includes streamlines representing the basic flow field within the squall line, including an ascending front-to-rear current (Smull and Houze 1987a) and a descending rear-to-front current, or “rear inflow jet” (Smull and Houze 1987b). Finally, the updated schematic includes basic representations of the pressure field.
throughout the storm. This includes surface pressure features such as the meso-high induced hydrostatically within the squall line’s cold pool and a trailing wake-low that results from warming aloft in the stratiform region (e.g. Fujita 1955; Johnson and Hamilton 1988), along with a mid-level low pressure centers located near the rear of the convective region and just above the melting layer that owe their existence the effects of diabatic heating (e.g. Smull and Houze 1987b; Lafore and Moncrieff 1989).

Squall lines are sustained by what is referred to as a “multi-cell” process wherein new convective cells are periodically triggered along the leading edge of the system (e.g. Browning 1977; Fovell and Dailey 1995; Fovell and Tan 1998). More specifically, these cells develop at the leading edge of the expanding pool of cold air that develops as a result of rainy downdrafts associated with the squall line (e.g. the “cold pool”). At any given stage there are likely multiple cells present at different stages in their development, as shown in the schematic in Fig. 1.2. Provided an environment favorable for the continued redevelopment of new cells, squall lines can thus persist for many hours, and impact areas covering 1000-10,000s of km².

For cold pool driven systems, such as squall lines, this periodic regeneration tends to occur along the downshear side of the cold pool. Rotunno et al. (1988) presented a theory explaining this process, demonstrating that favorable lifting occurs on the downshear side of the system-generated cold pool as the shear offsets the horizontal vorticity generated by the squall line’s cold pool. The more nearly the wind shear balances this horizontal vorticity, the stronger the low-level lifting. This process will be discussed in greater detail in Chapter 2.

While the observational studies cited above have provided a wealth of knowledge in terms of identifying the key structures and common behaviors that characterize squall lines, often times authors were left to speculate about the mechanisms or processes responsible. Early numerical simulations using simple, 2-D models helped to shed light on some of the basic processes at work within squall lines (e.g. Hane 1973; Thorpe et al. 1982), however as advances in
computing made complex, 3-D simulations feasible, the utility of numerical models expanded greatly. For instance, idealized simulations have helped to identify the governing dynamics associated with a number of commonly observed squall line features and the roles they play in squall line evolution. These include the processes that drive: bow echoes (e.g. Weisman 1993), rear-inflow jets (e.g. Weisman 1992; 1993); “bookend” vortices (e.g. Weisman 1993; Skamarock et al. 1994), mesoscale convective vortices (e.g. Davis and Weisman 1994; Cram et al. 2002), and squall line mesovortices (e.g. Trapp and Weisman 2003; Weisman and Trapp 2003; Atkins and St. Laurent 2009a; Atkins and St. Laurent 2009b). Idealized simulations have also proven fruitful in elucidating the dynamical features of various “non-traditional” squall line archetypes including squall lines with leading- stratiform precipitation (Parker and Johnson 2004a; Parker and Johnson 2004b) and parallel-stratiform precipitation, (Parker 2007a; Parker 2007b) as well as quasi-stationary MCSs that produce heavy rainfall (Schumacher 2009). Recent studies have also begun employing full-physics mesoscale models, such as the Weather Research and Forecasting model (Skamarock et al. 2008) for studies of squall line phenomena in realistic background environments. These types of studies open up a whole new realm of possibilities in terms of evaluating the predictability of squall lines in realistic environments (e.g. Trier et al. 2006), while also improving our understanding of observed cases through the detailed analysis of realistic simulations (e.g. Wheatley et al. 2006; Schumacher and Johnson 2008; Trier and Sharman 2009). In short, as computing power has advanced over the last two decades, increasingly realistic squall line simulations have provided a useful tool in expanding our knowledge base. These simulations effectively provide a means of isolating the processes responsible for observed squall line behaviors, producing a more well-rounded picture of how squall lines work.
2. Organization of the following chapters

This marriage of high-resolution observations and numerical simulations has provided us with a clearer picture of squall line dynamics than ever before. It is in this vein that the present study is introduced, as it follows closely in the tradition of using observed data to identify key behaviors characteristic of real-world squall lines, and then running targeted numerical simulations to identify the processes responsible for these behaviors. In Chapter 2, this consists of creating idealized simulations set up to mimic the often-observed environmental condition of a developing low-level jet associated with a stabilizing boundary layer. This condition is applied to a mature squall line in order to understand how the direct impacts upon a squall line’s convective-scale processes. The work presented Chapter 2 has been recently been published in the *Journal of the Atmospheric Sciences* (see French and Parker 2010).

Chapter 3 focuses on identifying the key features found in observed merger cases. This consists of reviewing archived radar data and Rapid Update Cycle (RUC, Benjamin et al. 2004) model analyses to find features common to a number of squall line-supercell merger events. The goal of this portion of the study is to determine how common some of the past observations of squall line-supercell interactions are, and perhaps provide a more general sense of the types of storm organizations that can be expected in these scenarios. The results presented in Chapter 3 have been submitted to *Weather and Forecasting* and are currently in review.

Chapter 4 builds on the observational results presented in Chapter 3, as a series of numerical simulations are presented to explore the storm-scale processes related to squall line-supercell mergers. By using an idealized model configuration, the impact of the merger can be isolated from other controls on storm evolution such as the background environment. Furthermore, the high-resolution simulation of a representative merger case provides a unique opportunity to investigate the dynamical processes a work in these cases, and test some of the hypotheses
generated based on the observational results in Chapter 3.
FIGURE 1.1: Schematic vertical cross section of a squall line with trailing stratiform precipitation (from Houze et al. 1989).

FIGURE 1.2: Cross section through a multicell storm over the course of 20 minutes. Note the new cells developing on the left flank of the storm (cell 5), while old cells are dissipating to the right (cell 1) (from Doswell 1985).
Chapter 2

The response of simulated nocturnal convective systems to a developing low-level jet

1. Introduction

Warm season precipitation has long been observed to exhibit a nocturnal maximum over the central United States (e.g. Kincer 1916), which has been attributed to a preponderance of nighttime thunderstorms and mesoscale convective systems (MCSs) that cross the region (e.g. Wallace 1975; Maddox 1980). While these MCSs benefit agriculture by supplying a significant portion of the growing-season rainfall in this region (Fritsch et al. 1986), they can also cause flooding (Doswell et al. 1996) and severe weather (Gallus et al. 2008) with associated concerns for life and property. In light of this, nocturnal MCSs have been the focus of a variety of studies, using both observations (e.g. Maddox 1980; Maddox 1983; Wetzel et al. 1983; Goodman and

---

The material presented in this chapter has been published as French and Parker (2010), (c)American Meteorological Society. Reprinted with permission.
Determining the processes that sustain these convective systems in a nocturnal environment featuring a statically stable lower troposphere has been of particular interest. To this end, Parker (2008, hereafter P08) examined the effect of low-level (i.e. the lowest 1 km) cooling on simulated squall lines, focusing on the convective-scale dynamical processes. He found that as the convective system transitioned from surface-based (ingesting parcels from approximately the lowest 500 m) to elevated (ingesting parcels from above approximately 500 m), the mechanism responsible for lifting inflowing parcels evolved from a cold pool to a bore. In the interest of simplicity, the simulations of P08 utilized a homogeneous background wind profile that did not include a low-level wind maximum, or low-level jet. Past observational research has identified a number of low-level jet structures that are common in MCS environments. Some of these are generated in response to the MCS itself such as squall line rear-inflow jets (e.g. Smull and Houze 1987b; Weisman 1992), or line-parallel jets that develop in response to convective heating such as that observed along cold frontal systems by Lackmann (2002). The present study is interested in low-level jet structures that are driven by boundary layer and terrain processes (e.g. Blackadar 1957; Buajitti and Blackadar 1957; Holton 1977), such as the "Great Plains" low-level jet found in the central United States (e.g. Means 1952; Means 1954; Bonner 1968). This low-level jet develops independently of the MCS, and has been commonly observed as a background environmental feature in nocturnal MCS environments (e.g Maddox 1983; Cotton et al. 1989). Thus, hereinafter "low-level jet" (LLJ) will refer to this environmental phenomena rather than any MCS-induced jet structure. While numerous previous studies have highlighted the importance of the LLJ as a forcing mechanism and as a means of transporting high-$\theta_e$ air into the region where storms are occurring, few have examined how the development of the
jet impacts the dynamics of pre-existing convection. The present work looks to fill in this
gap in the knowledge base by focusing on how a developing LLJ effects simulated nocturnal
convective systems, namely those with a linear organization such as squall lines.

a. Background

A primary reason that the LLJ is of interest is its long-standing association with warm-
season precipitation in the central United States, particularly thunderstorms (e.g. Means 1952;
Pitchford and London 1962; Higgins et al. 1997). Given its location and typical southerly di-
rection, the LLJ is a significant source of instability for the central United States because it
advects warm, moist air from the Gulf of Mexico northward into the Great Plains and Mid-
western regions (Helfand and Schubert 1995; Higgins et al. 1997; Tollerud et al. 2008). This
warm, moist air is often potentially buoyant and can provide the sustaining inflow for convec-
tive storms (Maddox 1983; Cotton et al. 1989). The LLJ can also destabilize a region when
its associated ribbon of high-$\theta_e$ air under-runs colder air aloft, leading to either potential or
conditional instability that, when released, fuels storm development (Trier and Parsons 1993).
Additionally, the jet can locally deepen the moist layer via strengthening isentropic lift and
frontogenetic circulations in cases where the jet interacts with deep thermal boundaries (Trier
et al. 2006). These effects can be especially significant for nocturnal convection or storms that
form on the cool side of frontal boundaries, as the LLJ provides an elevated source of unstable
air upon which storms can be sustained despite the stable boundary layer (Trier et al. 2006).

In addition to helping prime the convective environment, the LLJ can also provide a signif-
icant forcing mechanism for long-lived convective systems. This is especially true when the
jet intersects a frontal boundary (Pitchford and London 1962; Augustine and Caracena 1994;
Anderson and Arritt 1998). As the jet impinges upon the frontal boundary, isentropic upglide
and frontogenetic forcing are enhanced, leading to a local increase in vertical motion, and providing a lifting mechanism for storms (Trier and Parsons 1993; Tuttle and Davis 2006; Trier et al. 2006). This forcing can be important to storm longevity during the overnight hours. In a study examining MCS dissipation, Gale et al. (2002) noted that not only was an LLJ present in a majority of their cases, but that storm dissipation often corresponded to the removal of the LLJ forcing. This was often the result of the jet becoming oriented more parallel to the front, resulting in a decrease in isentropic upglide and frontogenetic forcing for vertical motion. The authors attributed this change in direction to the diurnal inertial oscillation. In the absence of a frontal boundary, upward vertical motion associated with speed convergence at the terminus of the jet (Trier and Parsons 1993), or interactions between the LLJ and a pre-existing mesoscale convective vortex (Schumacher and Johnson 2009) are also capable of initiating and sustaining MCSs. As a forcing mechanism, the strength of the LLJ also has significant implications for MCS intensity. Tuttle and Davis (2006) found that stronger jets were associated with heavier rainfall. This corresponds with the findings of Arritt et al. (1997), who concluded that an increased frequency of strong LLJs may have played a role in the significant flooding that occurred across the Midwestern United States in 1993.

Finally, given this role as a forcing mechanism for MCSs, Corfidi et al. (1996) developed a method for forecasting the motion of mesoscale convective complexes (MCCs) that included using the LLJ as a proxy for the propagation component of storm motion. Quite simply, storm propagation is represented by a vector equal in magnitude, but opposite in direction to the LLJ. A refinement of this study (Corfidi 2003) noted that this proxy does not work as well for cold pool driven systems that move downwind, and is instead more applicable to systems that propagate upwind, such as the “backbuilding” mode of Bluestein and Jain (1985) or the quasi-stationary/backbuilding mode of Schumacher and Johnson (2005).

In short, a significant body of work has been compiled outlining the role of the LLJ as an
environmental feature associated with large, long-lived MCSs. However, much of this work has focused on either the role that the jet plays in priming the convective environment (i.e. transporting high-\(\theta_c\) air to the region where storms are occurring), or on storms that develop as a direct result of forcing by the jet. Missing from this collection of previous work is an explanation of how the development of the nocturnal jet may affect pre-existing convection, i.e. storms that have formed during the afternoon and continue into the overnight hours as the boundary layer stabilizes and the LLJ develops. Of particular interest is how the LLJ impacts the convective-scale dynamics of these systems. One way that the development of the nocturnal LLJ may impact ongoing storms, particularly squall lines, is through changing the vertical wind shear profile. Given that vertical shear is important for squall line organization and longevity (e.g. Rotunno et al. 1988, hereafter RKW), changes to the vertical shear due to the developing LLJ are hypothesized to be important as a means of modulating updraft intensity. In addition, the presence of an LLJ may also govern the intensity of the storm relative inflow that sustains a squall line, which can directly impact storm intensity (Gale et al. 2002).

The present work looks to examine these effects using idealized numerical simulations. The following section provides an overview of the methods used in this study. This is followed in Section 3 by a discussion of the results of a set of simulations that utilize periodic-\(y\) boundary conditions. Section 4 details the results of a variety of sensitivity tests, and Section 5 examines the impact of adding the jet to a non-periodic squall line. These results are synthesized in Section 6, leading up to a summary and conclusions in Section 7.

2. Methods

This work utilized 3D idealized numerical model simulations using version 1.10 of the Bryan cloud model (CM1) described by Bryan and Fritsch (2002). In order to compare our results to previous work, our model configuration followed that of P08. This included using
the Lin et al. (1983) ice microphysics scheme (with the modifications of Braun and Tao (2000) included), and a 250 m horizontal grid spacing in order to explicitly represent convective processes. The vertical grid spacing was stretched from 100 m at the surface to 250 m above \( z=2500 \) m. This provided higher resolution in the lower troposphere, where both a nocturnal stable layer and the simulated LLJ were located. The 3D simulations used a 400 x 60 x 20 km domain with periodic y- (along-line) and open x-lateral boundary conditions. Finally, we employed free-slip upper and lower boundary conditions, with a Rayleigh damping layer above 14 km. In the interest of simplicity, accelerations due to the Coriolis force, as well as any radiative effects, were neglected. This basic model configuration was used for the majority of our simulations; any changes for subsequent tests will be noted where applicable. All of the simulations were run for 10 hours.

As in P08, the guiding philosophy is to mimic the transition of afternoon storms as night falls. The base-state environment employed for these simulations was the same horizontally homogeneous environment used in P08 (Fig. 2.1), utilizing the mean mid-latitude MCS sounding with a deepened moist layer from Parker and Johnson (2004c). This deeper moist layer is representative of an environment that supports surface-based storms, but also contains an elevated layer of high-\( \theta_e \) air, which is necessary to sustain an MCS in an environment with a stable boundary layer. In the central United States, this air is commonly advected into place by the LLJ; for simplicity, we assume that the higher \( \theta_e \) layer is already present, so that the dynamical impact of the LLJ is isolated to the changing wind profile. The rapid decrease in relative humidity above this moist layer contributes to a potentially unstable layer (\( \frac{d\theta_e}{dz} < 0 \)) between approximately 1-4 km AGL. This may contribute to increased destabilization through the formation of moist absolutely unstable layers in and near the region of strong gust-front lifting that develops with the simulated squall line (e.g. Bryan and Fritsch 2000, serving to enhance convective motions. Convection was initiated within this environment using a 10 km
wide, 3 km deep, north-south line thermal that spanned the domain in the y-direction. This initial perturbation was 2 K warmer than the background environment, had a relative humidity of 85%, and featured random temperature perturbations of up to 0.1 K to help facilitate 3D structures.

Artificial low-level cooling was introduced in the simulations using much the same method as that used by P08. In P08, the simulated squall lines were allowed to mature for 3 hours after which a simplified low-level cooling term was added in the lowest 1 km of the model. The net effect of this cooling was to create an isothermal layer whose temperature decreased from the initial surface temperature at a rate of $-3 \text{ K h}^{-1}$. This approach is again adapted in the present study as a simple way of enabling convection to mature in an afternoon-like sounding and then to evolve as a nocturnal-like sounding develops. The cooling is applied only at those grid points whose temperature exceeds the temporally declining isothermal reference temperature. In the present study the isothermal reference temperature decreases indefinitely, making the cooling configuration the same as the “-unlim” experiments of P08.

When such cooling is applied indefinitely, a fog will eventually form in the low-level environment, a situation that is not handled well by the model’s idealized microphysical parameterization. Therefore, P08 reported that at grid points where the artificial cooling was applied, the relative humidity (RH) was reset to be no larger than 0.98 through the instantaneous removal of water vapor (without any latent heat release). In the course of extending the work of P08 via some new experiments, we discovered: a) that a number of the simulations described by P08 actually used an older version of the artificial cooling scheme that did not include the RH treatment; and, b) that the RH treatment, as it was implemented in the remaining simulations, appears to have some undesirable effects. These problems were attributable to a coding error introduced by P08 and are not general problems with the standard distribution of the Bryan cloud model. The ramifications of these deficiencies are discussed in more detail in Appendix
A. In revisiting the original P08 simulations to determine the sensitivity of the results to the RH correction, we have developed a revised technique that is much preferred to the one described by P08.

In the revised version, the artificial cooling routine has been moved so that it occurs at the end of the model microphysical parameterization instead of during the model’s main prediction of potential temperature computations. This allows the model to treat moist processes associated with the model’s simulated physical processes first, before the artificial cooling is introduced. As a result of this improvement, it was then possible simply to add the artificial cooling and remove any induced supersaturation, resetting RH to be no larger than 1.00 and allowing no condensation during the artificial cooling step. This revised technique is preferable to the original, and has been applied uniformly to all of the present simulations. The modest impacts of this revision upon the results of P08 are addressed in the appendix.

In order to study the effects of an LLJ on nocturnal convective systems, it was of interest to include a developing low-level wind maximum within these simulations. A review of past LLJ climatologies (Bonner 1968; Mitchell et al. 1995; Arritt et al. 1997; Whiteman et al. 1997) provided the basis for our simulated LLJ. This included using a jet that was 5 m s$^{-1}$ stronger than the background winds$^1$, approximately 1 km deep, and centered just above the developing stable layer (in our case, just above 1 km AGL). As is often observed (Mitchell et al. 1995) our jet gradually develops over the course of 5 hours (beginning 3 hours into the simulation, the same time as the onset of the low-level cooling) resulting in the perturbation wind profile shown in Fig. 2.2a. This profile features a rapid increase in wind with height to the maximum jet speed followed by a gradual relaxation back to the base state profile above this level, similar to the LLJ wind profiles observed by Whiteman et al. (1997) and Tollerud et al. (2008). While several

$^1$Stronger jet speeds were tested as well, and did not fundamentally change our results. Jet-induced changes were of the same sign, just larger in magnitude for a stronger jet.
climatologies (e.g. Bonner 1968; Whiteman et al. 1997; Song et al. 2005) place the level of maximum winds within the jet below the 1 km AGL level used in our simulations, they also point out that the jet tends to be located just above the top of the nocturnal stable layer. Since the latter of these would appear more dynamically relevant, for our primary simulations we chose to place our jet relative to the top of our imposed stable layer rather than at a specific height. Finally, three different jet orientations relative to the squall line’s leading line were tested; rear-to-front (RTF), front-to-rear (FTR), and parallel (PAR) (Fig. 2.2b). Of these orientations the FTR and PAR are likely the most representative of typical jet orientations seen in nature, with the FTR representing cases where squall lines propagate into a low-level jet (e.g. Corfidi et al. 1996; Corfidi 2003) and the PAR representing squall lines moving along a stationary boundary oriented perpendicular to a low-level jet (e.g. Johns 1993; Trier and Parsons 1993). However, the RTF configuration is of interest is it provides a means of more completely examining the parameter space by representing the extreme at the opposite end of the spectrum from the FTR jet. Namely, comparisons between these two jet configurations produce opposing shear and storm-relative inflow profiles, helping to isolate the impacts of both of these features on the simulated squall lines. These simulations were compared to a control simulation (CTL) which did not include any jet development, and was therefore identical to the “deep-unlim” simulation of P08.

It should be noted that our applied LLJ is a jet in the vertical only, as it is applied uniformly across our domain in the x and y directions. In addition, we do not limit the jet application to the undisturbed environment ahead of the squall line, but rather apply it at every point in the x and y, which includes points within our simulated squall line. This was done to avoid introducing additional convergence within the model at the jet terminus, and thus isolate the jet as a characteristic of the larger (meso-beta/alpha) scale environment rather than a forcing mechanism for our simulated squall line. There is little evidence of the applied u-perturbations within
the squall line structure, likely owing to strong vertical motions effectively mixing the narrow maximum with the surrounding wind field. Furthermore, we ran several 2D tests wherein the jet was limited to the pre-line region of the model domain, and found the primary impact was that storm motions were more uniform across the simulations. Overall, key features and storm evolutions remained unchanged, giving us confidence that our results are robust.

3. Benchmark simulations

a. Overview of simulations

In order to test the effects of adding a basic LLJ structure to the simulations of P08, we first varied the jet orientation with respect to the simulated linear MCS. An examination of the CTL, FTR, RTF, and PAR (not shown) simulations shows a fairly similar evolution through 6 hours of run time (Fig. 2.3). In each case, an initially surface-based squall line develops and becomes elevated as is illustrated in Fig. 2.4 for the CTL simulation. The shaded tracer in Fig. 2.4 denotes low-level parcels being ingested by the squall line, with a decrease in tracer denoting the transition from surface-based to elevated convection as the surface temperature decreases and low-level CAPE tends to 0 (left and right-hand panels, respectively of Fig. 2.4). As discussed in P08, the life cycle of the CTL squall line is characterized by a developing stage as the squall line intensifies between 0-3 hours, a mature/steady phase between 3-6 hours while the squall line is surface-based, a transitional stage as the squall line evolves from surface-based to elevated between 6 and 8 hours (the “stalling phase” discussed in P08), and finally the elevated stage wherein the squall line is only ingesting parcels from above the boundary layer. All of the simulations steadily diminish in intensity with time (e.g. Figs. 2.3 and 2.5) due to dwindling pre-line CAPE values in response to the low-level cooling. This evolution was discussed at length by P08, and given that all three jet simulations largely mimic the CTL
simulation in this progression, the reader is referred to that publication for more detail.

The observed similarity is maintained between CTL and PAR for the duration of the simulations (not shown), suggesting that for the case of periodic-\(y\) boundary conditions, a line-parallel LLJ has little effect on the simulated squall line. In light of this, subsequent discussions will focus solely on the line-perpendicular jet orientations (RTF and FTR), as these produced more dramatic changes from CTL. An analysis of the impacts of the PAR configuration using non-periodic boundary conditions and a finite length squall line are discussed in Section 5.

In general, the CTL, FTR, and RTF simulations are qualitatively similar through approximately 6 hours, at which point the region of higher reflectivity in the RTF simulation begins to diminish at a faster rate compared to the CTL and FTR simulations (Fig. 2.3). By the end of the simulation (10 hours), reflectivity values have decreased in all three of the simulations, but the FTR run continues to exhibit a larger area of \(>40\) dBZ simulated radar reflectivity than the CTL run, while this area is considerably smaller in the RTF simulation. A more quantitative examination of the evolution during this period is evident using the total upward mass flux (TUMF, Fig. 2.5a). Starting at approximately \(t = 3:30\) (30 minutes after the onset of low-level cooling and LLJ application), the FTR simulation begins to exhibit larger TUMF than the CTL simulation, while the TUMF becomes becomes slightly smaller in the RTF run. This pattern is maintained throughout the duration of the simulation. This suggests that the addition of the LLJ very quickly leads to an increase in total upward mass flux in the FTR simulation, which eventually manifests itself in a prolonged duration of stronger precipitation compared to the RTF simulation, as is evident in the simulated radar reflectivity (Fig. 2.3).

A time-series of the maximum along-line averaged vertical velocity \((\overline{w}_{max})\) indicates that updraft evolution is not as straight forward (Fig. 2.5b). By approximately 4 hours (1 hour after the onset of the low-level cooling and LLJ application) the RTF and FTR simulations begin to diverge from the CTL simulation. The RTF simulation develops stronger updrafts
while the FTR updrafts remain weaker (and shallower, not shown). This discrepancy increases until approximately t= 5:30 at which point the RTF simulation has along-line averaged vertical velocity ($\bar{w}$) that is 2 ms$^{-1}$ stronger than the FTR (Fig. 2.6b, c). In addition, both the RTF and CTL simulations feature unbroken regions of strong ( >5 ms$^{-1}$ ) vertical velocities extending approximately 1 km deeper than in the FTR simulation (dashed contours in Fig. 2.6). This suggests weaker low-level lifting occurring in the FTR simulation during this period compared to both the RTF and CTL simulations. This trend reverses between approximately t = 6:00 and 8:00, (Fig. 2.5b). The updrafts in the RTF simulation begin a steady decline in intensity and depth, while those in the FTR remain nearly steady in intensity and begin to deepen. The three simulations are comparable by t = 6:45 (e.g. Fig. 2.5b), and by t = 7:30 the FTR simulation has surpassed both the CTL and RTF in updraft intensity (nearly twice the intensity of the RTF) with strong $\bar{w}$ extending to a higher altitude than both CTL and RTF (Fig. 2.7). Lastly, after about t = 7:30, a final evolution takes place, as $[\bar{w}]_{max}$ values in the FTR squall line weaken rapidly, while those in the RTF slow in their decrease (Fig. 2.5b), and the simulations’ $[\bar{w}]_{max}$ values become quite similar again. However, the low-level structure varies considerably during this period, with the RTF exhibiting a narrow, erect updraft, while the FTR features a more broad, rear-ward sloping structure (Fig. 2.8). The reasons for these comparative changes are explored next.

b. Role of changes to the low-level wind shear

One way in which the development of a low-level jet can impact a squall line is by changing the vertical wind shear of the pre-line environment, which can have a significant impact on the strength of vertical motion within the squall line. As presented by RKW, lifting along a squall line’s cold pool is optimized when the cold pool strength, $C$ (a measure of integrated buoyancy
within the cold pool) is balanced by the environmental wind shear in the pre-line region, $\Delta u$. The physical interpretation of this balance is that the negative horizontal vorticity produced baroclinically by the cold pool is balanced by the positive horizontal vorticity flux associated with the vertical wind shear, resulting in vertically erect, intense updrafts. This concept has been demonstrated repeatedly in numerical simulations of squall-lines (e.g. Weisman et al. 1988; Weisman and Rotunno 2004; Bryan et al. 2006; P08), and given the changes to the vertical shear that result from our addition of an LLJ, it is of interest to examine our simulations in terms of this cold pool-shear balance. As shown by P08, in the CTL simulation the cold pool vorticity is stronger than the shear to begin with, and thus we assess increases in vertical shear as bringing the system closer to balance and decreases in shear as taking the system further from balance.

RKW and a more recent study by Weisman and Rotunno (2004) utilize a layer of fixed depth in assessing the impact of the environmental shear (e.g. 0-2.5 km AGL). The general logic behind this treatment is that a) the shear and the cold pool’s buoyancy perturbation do not vary appreciably with height in the layer, and b) that the shear in the layer containing the cold pool is of greatest importance. The combination of these two assumptions leads to an elegant approach in which the cold pool’s buoyancy and environment’s flux of horizontal vorticity can be integrated and compared via the proxies $C$ and $\Delta u$. However, in the present study taking a simple vector wind difference over the 0 - 2.5 km AGL layer, for example, does not capture the impacts of our LLJ on the pre-line shear because the jet is contained almost entirely within this layer. As a result, the 0-2.5 km vertical wind shear values in the RTF and FTR simulations are nearly identical to those found in the CTL simulation (Table 2.1).

While assessing the appropriateness of the preceding assumptions for our LLJ experiments, we noted that the layer of strongly negative and nearly constant buoyancy in our own cold pool extended up through approximately 1 km AGL (e.g. left hand panels of Fig. 2.6). This also
nearly coincides with the height of the imposed LLJ’s wind maximum. We therefore find it most instructive to separate the 0-1 km and 1-2.5 km layers (due to their differing cold pool buoyancy and environmental shears). This separation at 1 km is also convenient because, once the artificial cooling is applied, the air parcels below 1 km AGL have escalating amounts of CIN and, eventually, vanishing CAPE (Fig. 2.9a, b). In the original formulation of RKW (e.g. their p. 477), the horizontal vorticity equation is integrated vertically. It seems to us that the integrated horizontal vorticity tendencies should be determined by the relationship of the (horizontal gradients in) environmental horizontal vorticity flux to the cold pool’s buoyancy in the layer where the cold pool buoyancy (and its horizontal gradient) are maximized. It also seems to us that, once parcels no longer have CAPE (i.e. in the layer of 0-1 km artificial cooling), they should be excluded from the problem (because they no longer take part in deep convection). These hypotheses underlie our approach to assessing the impact of shear on our simulations, using the 0-1 km and 1-2.5 km layers throughout for simplicity.

Over the 0-1 km layer the RTF jet results in an increase in vertical shear from the base state (CTL) environment (Table 2.1), which better balances the cold pool (found predominantly in that layer, Fig. 2.6). This is associated with an increase in $\overline{w}_{\text{max}}$ during the surface-based phase (t = 3:00-6:30), as is seen in Fig. 2.5b. Meanwhile, the FTR jet reduces the vertical shear in the 0-1 km layer which is concurrent with the decreased $\overline{w}_{\text{max}}$ in Fig. 2.5b. In short, while the squall line is surface-based, the below-jet shear is well-correlated to the peak vertical motions in our simulations.

Shortly after 6 hours the simulated squall lines enter the “stalling phase” which is characterized by a weakening of the surface cold pool and a secondary maximum in vertical velocity as discussed by P08. With the addition of the LLJ, this phase marks the start of a period of comparatively strong vertical velocities in the FTR simulation. As the squall line enters the stalling phase, several things happen simultaneously. First, as the stable layer continues to
cool, the temperature difference across the cold pool’s gust front weakens (Fig. 2.7, 2.9c), and the strength of lifting by the cold pool diminishes. This happens first at low-levels, where the air is colder, and then extends upward (Fig. 2.9c). At the same time, continued cooling of the stable layer continues to reduce CAPE and increase CIN (Fig. 2.9a, b). Again, this effect is largest at low-levels and extends upward with time. The combined impacts of the low-level weakening of cold pool and increasing pre-line stabilization causes the layer of potentially buoyant parcels feeding the squall line to shift upward with time. Fig. 2.9a also illustrates that CAPE for parcels above approximately 1500 m tends to increase in time. We attribute this to the presence of low-frequency gravity waves generated by the squall line that were also observed and explained at length in P08 (pp 1336-1337). As a result of these combined effects, the above-jet wind profile begins to have an increasing impact on the squall line. This results in the FTR squall line producing stronger vertical velocities than the CTL and RTF runs, as it has more favorable shear above 1 km (Table 2.1). In the RTF simulation, a precipitous drop-off in $\overline{[w]}_{max}$ occurs as the 0-1 km shear wanes in importance since there is very little above-jet shear (e.g. 1-2.5 km AGL, Table 2.1).

The 0-1 km and 1-2.5 km shear layers are pedagogically useful, but it is also worthwhile to assess the actual vertically integrated horizontal vorticity flux in the LLJ simulations, because the jet changes both the shear and the system-relative flow magnitude. Fig. 2.10a compares the cold pool strength, $C$, for the CTL simulation to the vertically integrated horizontal vorticity flux, $u\eta|z$, for the RTF and FTR simulations. Both $C$ and $u\eta|z$ are integrated over a subjectively determined effective inflow layer that is characterized by the union of the strongest buoyancy perturbation\(^2\) within the cold pool ($B'$) and a pre-line thermodynamic environment characterized by large CAPE and small CIN (denoted by gray shading in Fig. 2.9). During

\(^2\)This perturbation is calculated relative to the changing pre-line environment in order to account for the cooling being applied to the inflow environment. Thus it is not a perturbation from the base state ($t=0$) buoyancy field.
the surface-based “steady phase” from t = 3:00-6:00 the below-jet shear predominates, with the RTF (FTR) configuration featuring the most (least) favorable shear (Fig. 2.10a). This corresponds well with the time evolution of vertical velocity, as the RTF simulation produces the strongest vertical motion during this time (Fig. 2.5b). From t = 6:30 - 7:30, as the effective inflow layer shifts upward, the above-jet shear becomes more important, with the FTR jet providing the most favorable conditions (Fig. 2.10a). This results in the FTR simulation producing the strongest vertical velocities (Fig. 2.5b), an outcome that is further enhanced because the cold pool has weakened a great deal by this point (Fig. 2.10a), and thus the system is closer to being in RKW balance. Meanwhile, since there is very little shear present above the jet in the RTF simulation, its \( u|_{z} \) declines dramatically, and the squall line weakens considerably during this same period.

After approximately t = 7:30, as the simulated squall lines become elevated and bore-driven (rather than cold pool-driven) a final evolution in updraft strength takes place. While all three simulations continue to diminish in updraft intensity, the RTF simulation begins to diminish more slowly, while the FTR hastens its weakening trend (Fig. 2.5b). Prior to this point, updraft intensity has been well-related to the strength of the vertical wind shear within the effective inflow layer, which causes the relevant shear layer to shift upward with time as the depth of the stable layer increases. However, this does not explain the later behavior as the FTR simulation contains the most favorable shear (Fig. 2.10a) but the weakest updraft (Fig. 2.8c). Instead, we hypothesize that once the system becomes bore-driven, the low-level vertical motions are controlled primarily by the amplitude of the gravity wave/bore. This mechanism can effectively maintain a simulated squall line largely because, once the system has become elevated, the source of parcels containing CAPE is almost entirely above 1 km (e.g. Fig. 2.11). These parcels need minimal vertical displacements to attain their level of free convection (which were generally located about 2 km AGL, left-hand panels of Fig. 2.11). Such displacements
are achieved merely by isentropic ascent at the leading edge of the bore (Fig. 2.11), and thus convection is sustained. However, vertical shear still plays an important role in modulating this mechanism.

Schmidt and Cotton (1990) demonstrated that the amplitude of low-level gravity waves that drive squall lines in a statically stable boundary layer is controlled by the vertical wind shear, with stronger shear resulting in a higher amplitude wave on the down-shear side of the system. A similar result was found by Buzzi et al. (1991) who determined that in the presence of a low-level stable layer, the shear within the stable layer played a significant role in organizing the solitary wave driving their simulated squall line. In the present simulations there is little evidence of the gravity wave/bore below approximately 0.5 km, thus we assume that the wave is propagating through the layer between 0.5 km and the top of the low-level stable layer (approximately 1.5 km AGL by t = 8:30). According to the studies cited above, the shear over this 0.5-1.5 km AGL layer should control the amplitude of the gravity wave/bore that drives the convective system, and thus govern updraft intensity. The RTF simulation features stronger shear between 0.5 and 1.5 km AGL compared to the FTR simulation (Table 2.1), and this corresponds to a higher amplitude wave and stronger low-level vertical velocities (Fig. 2.8b, c). This argument also explains differences in updraft shape that become evident between t = 8:00 and 9:00. In the FTR simulation, the lower amplitude wave produces a gradual rearward slope in the isentropes, which in turn accounts for the rearward tilted updraft evident in Fig. 2.8c. Conversely, the higher amplitude wave with vertically oriented isentropes in the RTF simulation produce the nearly vertical updraft that is observed (Fig. 2.8c).

To summarize, the LLJ-induced changes to low-level vertical shear control updraft intensity throughout the course of our simulations. While the system remains cold pool driven, the shear within the effective inflow layer appears to explain the changes in updraft strength. From approximately t = 3:00-6:30 this layer extends well below the LLJ, causing the below-jet shear
to have the largest impact, favoring stronger updrafts in the RTF simulation. As the stable layer grows and approaches the level of maximum winds in the LLJ (approximately \(t = 6:30-7:30\)), the above-jet shear plays an increasingly important role, and the FTR jet then provides the most favorable condition for strong updrafts. Finally, once the squall line has become elevated and bore-driven (\(t = 7:30-10:00\)), vertical motion is most clearly linked to the amplitude of the bore itself. This amplitude is controlled by the shear within the stable layer, with the stronger shear associated with the RTF jet providing the best conditions for strong lift. These arguments explain the evolution of vertical velocity observed over the course of our simulations, however they still do account for the persistent, increased values of TUMF found in the FTR simulation. To explain this behavior, we now examine the role of the LLJ in modulating system-relative inflow.

c. Role of changes to the storm relative inflow

A key element in the sustenance of any MCS is a source of high-\(\theta_c\) air that fuels the convective processes. In a study examining MCS dissipation (including daytime and nocturnal systems) Gale et al. (2002) found that a reduction or cessation of system-relative inflow (SRI) or, in the case of elevated storms, elevated system-relative inflow (ESRI), leads to MCS weakening and eventual dissipation. They also noted that maximum values for ESRI tended to coincide temporally with the maximum LLJ intensity (approximately 0600 UTC in their dataset), suggesting that the LLJ may play a role in modulating the intensity of the ESRI. For our purposes we define SRI as flow toward the squall line within a layer containing parcels with non-zero CAPE. This distinction is important to make because as the simulated stable layer grows, air within the stable layer still flows toward the squall line. However, due to the lack of CAPE, parcels are not actually ingested by the convective updrafts. Thus, the depth of SRI changes
with time, eventually becoming confined to a layer co-located with our simulated LLJ\(^3\) (Fig. 2.11a, b).

The changes to system-relative inflow over time (Fig. 2.12a) correspond well to the changes seen in total upward mass flux (Fig. 2.5b), with the RTF simulation seeing a reduction starting just after the implementation of the LLJ, and the FTR an increase. These changes become particularly pronounced as the system becomes elevated after approximately \(t = 7:30\) (Fig. 2.12a) due to the layer of high-\(\theta_e\) air becoming exclusively co-located with the LLJ (Fig. 2.11). These changes are directly related to the direction of the LLJ in the respective simulations. The RTF jet, directed away from the squall line, slows the background wind within the inflow layer (Fig. 2.11a), causing the observed decrease in SRI (Fig. 2.12a), while the FTR jet accelerates the background flow toward the squall line (Fig. 2.11b), increasing the SRI (Fig. 2.12a).

The variation in SRI is also evident in a trajectory analysis presented in Fig. 2.13. In the RTF simulation trajectories that move through the storm updraft have relatively short horizontal lengths (Fig. 2.13b) compared to those in the FTR simulation (Fig. 2.13c). The longer trajectories in the FTR simulation imply that during the 1-hour plotting interval the updraft ingests parcels from a region that extends well ahead of the squall line. In contrast, the shorter trajectories in the RTF simulation imply that only parcels from very near the updraft are being lifted during the plotting interval. In other words, the FTR squall line is ingesting more parcels per unit time than the RTF squall line, resulting in a larger number of updraft parcels overall (Fig. 2.13).

These variations in SRI ultimately correspond to differences in precipitation output. As

\(^3\)It is evident in Fig. 2.11 that the maximum pre-line CAPE in the FTR simulation is about 500 J kg\(^{-1}\) lower than in the RTF. This appears to result from a larger latent heat release in the FTR simulation producing a warmer forward anvil region which leads to a smaller area between the environmental and parcel temperature curves near the top of the column. Thus CAPE is reduced in the integrated sense, however inflowing parcels between 1 and 3 km and their properties up through the middle troposphere, are still thermodynamically similar for both simulations.
warm, moist air parcels enter the storm they ascend through the updraft region. In the FTR simulation, more air flowing into the storm results in more parcels being lifted, more condensation taking place per unit time, and greater hydrometeor production, ultimately resulting in an increase in precipitation output (Fig. 2.3g-i). The opposite was observed within the RTF simulation, wherein decreased horizontal mass flux into the storm leads to fewer parcels being ingested into the storm and lifted, decreased hydrometeor production, and ultimately a decrease in precipitation (i.e. Fig. 2.3d-f). This reduction in inflow explains why the RTF simulation contained diminished TUMF compared to the FTR simulation despite having stronger peak updraft speeds (Fig. 2.5a, b).

From Fig. 2.3 it is clear that the addition of the LLJ caused some change in the storm motion\textsuperscript{4}, which can also impact the SRI by changing the system-relative flow. However, the computed storm motions were still quite similar, generally $\pm 1 \text{ m s}^{-1}$ compared to the CTL simulation, which is small in comparison to the $5 \text{ m s}^{-1}$ magnitude of the imposed LLJ. Additionally, the RTF squall line exhibits a faster motion, which would tend to increase SRI (by increasing system-relative flow toward the squall line), contrary to what is observed. Thus the most substantial impacts of SRI upon TUMF are determined by the direction of the jet (toward or away from the system) in the pre-line environment.

To summarize the results of the benchmark simulations, there were two primary impacts from applying a simulated LLJ to our simulations. First, the LLJ alters the low-level wind shear profile, which controls gust front lifting as anticipated by the theory put forth by RKW. The LLJ also changes the storm-relative wind profile, impacting the strength of storm-relative inflow feeding the squall line, and with it the total upward mass flux within the squall line. These results are very much in line with the recent work of James et al. (2005), who found that \textsuperscript{4}These change appear to be due to simple advection by the LLJ. In 2D test simulations wherein the jet was limited to the pre-line region, storm motions were nearly identical for the RTF and FTR configurations.
low-level shear determines the vertical displacement of inflowing parcels, while storm-relative flow governs total upward mass flux. In their study, they used simulations of simple density currents to demonstrate that for constant shear, increasing the speed at which the cold pool moves (and thus the storm-relative flow) produces a concurrent increase in total upward mass flux. This is akin to our FTR simulation, however in our case the enhanced system-relative flow is caused by the LLJ rather than faster storm motion.

4. Sensitivity tests

a. Simulations without low-level cooling

The results of the previous section suggest that the impacts of the LLJ are strongly tied to the transition from surface-based to elevated convection. In order to separate the impact of the LLJ from the impacts of the low-level cooling, we re-ran our series of benchmark simulations without the cooling scheme applied. In essence, these runs simulate the impact of adding an LLJ to a squall line that remains surface-based and cold pool driven throughout its lifetime. We refer to these simulations together as the “NOCOOL” simulations, with CTLNC, RTFNC, and FTRNC corresponding to the benchmark CTL, RTF and FTR runs. Aside from the absence of the low-level cooling, the experimental set-up remains identical to the benchmark simulations.

The NOCOOL simulations are characterized by generally more steady behavior than the benchmark simulations, at least through approximately \( t = 8:00 \) (Fig. 2.14). This is most notable when examining \([\vec{\omega}]_{max}\), as the RTFNC simulation exhibits larger values than the CTLNC run through the majority of the simulation, while the FTRNC run exhibits weaker values through the entirety (Fig. 2.14b). This is because, in the absence of low-level cooling, the layer of relevant vertical wind shear remains between approximately 0-1 km (the depth of
the cold pool, as discussed earlier) throughout the simulation. As a result, the below-jet shear, which is stronger in the RTFNC configuration (Fig. 2.10b) remains important throughout the simulation. As in the benchmark simulations, the FTRNC squall line generates a larger TUMF than the CTLNC squall line throughout the run, despite now having the weaker vertical motion throughout, consistent with the concept that storm-relative inflow governs TUMF.

The primary departure from steady behavior in the NOCOOL simulations occurs within the RTFNC simulation. After about $t = 8:00$ this simulation exhibits a marked increase in TUMF and decline in $[\bar{w}]_{\text{max}}$ (Fig. 2.14). This is concurrent with a rapid increase in system speed (not shown) which is proportional to the cold pool strength, $C$. The increase in $C$ overwhelms the vertical wind shear, leading to the decline in $[\bar{w}]_{\text{max}}$. At the same time, the faster system motion increases the storm-relative flow (and thus SRI, Fig. 2.12b), leading to the uptick in TUMF that is observed. This is very much in line with the findings of James et al. (2005), as discussed earlier. Thus the NOCOOL simulations provide further evidence that 1) while the squall lines are surface-based, their behavior is controlled by the below-jet shear, and 2) TUMF is governed more by system-relative inflow than by updraft intensity.

\textit{b. 500m jet height}

As an additional sensitivity test, we present the results of a set of simulations run wherein the jet was lowered to approximately 500 m (model level $z = 513$ m). This was deemed an appropriate move because 1) in nature\textsuperscript{5} the LLJ tends to be found at approximately 500 m, and

\textsuperscript{5} As discussed in Section 2, our placement of the jet at 1 km AGL for the benchmark simulations was due to this being the top of our nocturnal stable layer, which is where the jet is located in general. In nature this layer is often shallower than 1 km, thus resulting in the LLJ being found to climatologically reside around 500 m AGL. However, in the interest of including a sufficient number of model grid points within our stable layer to facilitate its gradual growth, the 1 km depth was chosen.
by lowering the jet, we were able to assess both how the jet impacts SRI when it is removed from the layer of elevated high $\theta_e$ air, and how the jet-induced shear impacts the squall line when the below-jet region is comparatively shallower, and as a result stabilizes faster. These simulations are referred to as the RTF513 and FTR513 simulations, and will be compared to the same CTL simulation as was discussed previously. Aside from altering the height where the LLJ is applied, these simulations are identical to the benchmark RTF and FTR simulations. Several configurations that placed the jet higher than 1 km were tested as well, however seeing as these configurations removed the jet-related environmental changes from both the level of highest $\theta_e$ and strongest $B'$, the impact on the simulated squall line was minimal and thus will not be discussed in detail.

The 513 m simulations evolve quite similarly to the benchmark simulations in terms of vertical motion through approximately 6 hours. From $t = 3:00-6:00$, the RTF513 squall line produces larger vertical velocities than the FTR513, after which the FTR513 becomes dominant for the remainder of the simulation (Figs. 2.15). This transition occurs approximately 30 minutes sooner in the 513m simulations than in the benchmark simulations, which is to be expected, as the effective inflow layer is cut off at lower levels first and works its way upward with time. It is notable, though, that the squall lines continue to ingest parcels (and thus to flux horizontal vorticity) from below 500 m though this time period, as low-level CAPE decreases markedly and CIN grows (Fig. 2.9). As in the benchmark simulations the respective periods of stronger $\bar{\nu}_{max}$ correspond to periods of stronger $u\eta|_{z}$ in both simulations (Fig. 2.10c).

After $t = 7:30$, the behaviors of the benchmark and 513m simulations diverge as the FTR513 (RTF513) continues to produce stronger (weaker) vertical velocities through the end of the simulation (e.g. Figs. 2.15b, 2.16). Because of the lower jet axis, the FTR513 simulation has favorable shear within the 0.5-1.5 km AGL stable layer, leading to a higher amplitude gravity
wave once the system becomes elevated. For this reason, FTR513 retains the strongest $|\bar{w}|_{max}$ through the end of the 10 hour simulation (unlike in the benchmark case, c.f. Figs. 2.5b, 2.15b)

As in the previous simulations, the FTR-jet configuration generates the strongest TUMF throughout the simulation. Through approximately $t = 7:30$ both the RTF513 and FTR513 simulations have comparable values of TUMF to their benchmark simulation counterparts. However, after approximately $t = 7:30$ the FTR513 exhibits stronger TUMF than was seen in the original FTR simulation, while the RTF513 TUMF values approach those produced by the CTL simulation (Fig. 2.15a). It appears that, when the jet is located at 513 m (within the stable layer) it ceases to have a significant influence on system-relative inflow and TUMF once the system becomes elevated. This is evident in the time series of integrated storm-relative horizontal mass flux, as the FTR513 simulation mimics the CTL during this period (Fig. 2.12c). This suggests that the larger TUMF associated with the FTR513 simulation after $t = 7:30$ is likely due to the substantially larger values of $|\bar{w}|_{max}$ that are also present.

5. Non-periodic 3D simulations

In order to evaluate the effects of an LLJ on a fully three-dimensional system, a set of simulations were run using a non-periodic configuration. Instead of focusing on a segment of an infinitely long squall line, these simulations allowed us to examine the effects of the LLJ on the system as a whole, including the development of 3D asymmetries along the line. The primary changes to the model configuration for these simulations consisted of using an open $y$-lateral boundary condition and a larger domain in the $y$-direction (increased from 60 to 600 km) to allow for a sufficiently long convective line while keeping it away from the lateral boundaries. Additionally, a coarser horizontal grid spacing (dx and dy were increased from 250 to 500 m) was utilized in order to keep run times manageable given the larger domain. The squall line
was initiated using a warm line thermal as before, in this case limited to a $y$-length of 200 km and centered at $y=300$ km. The three benchmark simulations (CTLNP, RTFNP, FTRNP) as well as the line-parallel jet simulation (PARNP) were all re-run using this configuration (Fig. 2.2b).

In general the impact of the RTF and FTR jets in the non-periodic configuration was the same as for the periodic simulations, with the FTR creating more precipitation output and the RTF less precipitation. The impact of adding the line-parallel jet is subtle, with the CTLNP and PARNP simulations maintaining similar characteristics through 8 hours. After this point, the two diverge slightly, with the PARNP simulation featuring an increase (decrease) in precipitation along its southern (northern) flank compared to the CTLNP (Fig. 2.17). This result is in line with those from the periodic simulations discussed earlier. In this case, the southern half of the PARNP squall line is experiencing an effect akin to an FTR jet, with an increase in precipitation due to an increase in horizontal mass flux into the system. At the same time, the northern half of the PARNP squall line is experiencing an effect akin to the RTF jet, with a reduction in storm-relative inflow leading to the observed decrease in precipitation output. Thus the primary impact of the addition of a line-parallel LLJ to a completely 3D simulated system is an alteration of the precipitation distribution around the storm. Overall, though, the impact is subtle, likely owing to the deep-layer westerly shear and homogeneous background environment favoring continued development along the generally north/south oriented bore. The overall system motion continues to be eastward due to the mean westerly winds and system propagation in the downshear direction. In other words, storm motion is primarily governed by the base state wind profile, whereas the effect of the LLJ is to locally modulate the storm-relative inflow.
6. Discussion

This work set out with the goal of examining the impacts of adding a low-level jet to an elevated squall line. Two primary impacts were identified: 1) the LLJ changes the vertical shear, resulting in changes to updraft intensity in the simulated squall lines and 2) the LLJ modulates the intensity of the system-relative inflow within the jet layer, with the orientation of the jet governing whether SRI is increased or decreased. In general, the shear impacts modulated vertical velocities within our simulated squall line, while changes in storm-relative inflow impacted total upward mass flux and precipitation output by providing more mass to be processed by the system. Both of these impacts were affected by the transition from surface-based to elevated convection over the life of the squall line. While the system was surface-based, the shear below the jet appeared to have the largest impact (Fig. 2.18a). As the system became elevated the shear above the jet became increasingly important, as low-level CAPE decreased and the strongest $B'$ shifted upwards due to low-level stabilization (Fig. 2.18b). Once the system became completely elevated, the forcing became that of a bore (as detailed by P08), and the vertical shear ceased to directly impact low-level lifting via density current-shear dynamics. Instead it appeared to control the amplitude of the gravity wave/bore (Fig. 2.18c). Storm-relative inflow impacts also varied with system evolution, with LLJ-induced changes to SRI becoming more pronounced as the system became elevated and the layer of highest- $\theta_v$ air was limited to that of the LLJ.

Fritsch et al. (1994) and Schumacher (2009) have also investigated the impact of an LLJ on the RKW cold pool-shear balance. As in the present work, both studies found that an LLJ oriented toward the system (our FTR configuration) resulted in unfavorable shear (in the RKW-sense) below the jet, inhibiting lifting, and favorable shear above the jet, promoting stronger lifting. In these studies additional low-level lifting, in the form of isentropic ascent associated
with a mesoscale convective vortex (MCV), effectively compensated for the less-favorable RKW condition below the jet. Once parcels were lifted sufficiently through this mechanism, they were accelerated upward by the more favorable RKW conditions in the above-jet layer. Thus in nature additional factors can compensate for unfavorable shear associated with a low-level jet. However, as the present results suggest, when cold-pool lifting is dominant, the shear associated with the layer of high-$\theta_e$ inflow is of primary importance. In light of this, for the case of vertically varying cold pool buoyancy and environmental horizontal vorticity flux, we suggest that the commonly applied “RKW balance” concept may be most applicable to an “effective inflow layer” which we define as follows. The predominant contribution to the integrated horizontal vorticity tendency emerges from a layer defined by the union of a) potentially buoyant inflow parcels (CAPE > 0 J kg$^{-1}$ and small CIN) with b) substantial negative buoyancy (and hence baroclinity) in the cold pool. In the present simulations, even though the total shear over the entire depth of the cold pool (i.e. from the 0-2.5 km AGL) changed very little, the flux of horizontal vorticity within the “effective inflow layer” described the differences among the simulated vertical velocities quite well.

The findings presented herein also suggest that changes to the system-relative inflow, or the rate at which parcels are ingested by a squall line, can have a more significant impact on total upward mass flux and precipitation output than maximum updraft strength (much as shown by James et al. (2005)). Conditions that favored enhanced flow toward the storm universally saw an uptick in total upward mass flux. Perhaps the most dramatic example of this was seen in the RTFNC simulation, where an increase in system speed, and thus system-relative flow, simultaneously resulted in both a precipitous drop in $[\bar{\tau}]_{max}$ and an increase in upward mass flux (Fig. 2.14). Additionally, all of the simulations using the FTR jet configuration demonstrated an increase in TUMF that began just after the LLJ started to develop and lasted for the duration of the simulation. This can make it challenging to quantify squall line intensity,
as two common metrics, updraft velocity and total upward mass flux do not always parallel one another. In effect, an enhanced horizontal mass flux into the system can lead to an increase in TUMF by increasing the mass available to be fluxed upward, even while vertical motion may decline.

One of the motivations for the PARNP simulation was to provide a comparison to the work of Corfidi et al. (1996) and Corfidi (2003), who emphasized the role of the LLJ in governing squall line motion. Our results indicate that enhanced precipitation does appear to occur on the flank of the storm that is interacting with the LLJ (in our case the southern flank, Fig. 2.17). However, in our simulations no significant change in storm motion toward the direction of the LLJ was observed. Rather, the squall line appeared to continue to move in the base-state downshear direction (eastward), apparently due to lifting along the downshear edge of the bore and advection by the mean westerlies in our wind profile. This is in agreement with the findings of Corfidi (2003), that for cold pool-driven (or, in this case, bore-driven) squall lines, the direction of cold pool (bore) motion better represents the propagation component of the squall line’s motion. It is possible, however that in cases of weaker westerly shear, or a stronger jet, the LLJ may have a more significant effect on altering storm motion, as its impacts would be larger in a relative sense. Our simulations only continued for 2 hours following the jet reaching its full intensity. Perhaps a longer period of influence by the jet would have had a more significant impact on storm motion. Additionally, in nature the LLJ is often longitudinally narrow and is the sole region of elevated instability, which would tend to favor propagation into the jet, as that is where the most instability to fuel new cell development would be located.

As a final note, we emphasize that the time and duration of these evolutionary stages (e.g. in Fig. 2.18) may be somewhat different in nature. We allowed artificial low-level cooling to proceed at a constant rate indefinitely. In nature, where the nocturnal temperature decrease varies in rate and total magnitude from day to day, squall lines may reside in any one of these
three stages (e.g. Fig. 2.18) for much longer or shorter periods.

7. Conclusions

A series of numerical simulations have been performed to evaluate the effects of the addition of a low-level jet on a simulated squall line that is subjected to low-level cooling. The jet changes the squall line in two ways: 1) by altering the low-level wind shear and thus modulating low level lifting and 2) by modulating the intensity of the storm-relative inflow. Overall, changes to the vertical shear have the largest impact on vertical velocities within the squall line. To this end, below-jet shear appears to have the most significant impacts while the system is surface-based, with above-jet shear having a larger impact as the system becomes elevated. Once the system becomes entirely bore-driven, vertical shear continues to play a role in modulating vertical motion, however it is by impacting the amplitude of the gravity wave/bore within the low-level stable layer. The exact details of the process remain unclear, although it is possible, as speculated by Schmidt and Cotton (1990), that it could be similar to the vorticity balance discussed by RKW. Future investigation to this end could be enlightening. The changes to system-relative inflow, on the other hand, impacted total upward mass flux and precipitation output, with a jet directed toward the squall line favoring enhancement of these fields, and a jet directed away from the squall line a reduction. The impacts of the system-relative inflow were more dramatic once the stable layer had deepened and the high-$\theta_v$ air was entirely co-located with the LLJ.
FIGURE 2.1: Skew-T ln-p diagram of the base state temperature and dew point (thick black lines) and wind (barbs, half and full barbs = 2.5 and 5 ms$^{-1}$, respectively) profiles for all simulations.
Figure 2.2: (a) Time-series of perturbation u-wind profiles illustrating the development of the simulated LLJ. (b) Schematic illustrating the different LLJ orientations. The wind profiles in (a) are from the rear-to-front (RTF) simulation. The other jet orientations have an identical shape and magnitude, but a different direction.
Figure 2.3: Vertical cross sections of simulated radar reflectivity (dBZ, shaded as shown) for the CTL (a,b,c), RTF (d,e,f), and FTR (g,h,i) simulations at (from left to right) t = 6:00, 8:00, and 10:00.
FIGURE 2.4: Depiction of the CTL simulation over time: (left) values vs time for the reference environmental temperature (Tref) and the minimal surface temperature on the domain (Tmin), using line styles as shown; (center) Hovmöller diagram for 5 km AGL, with along-line maxima in tracer concentration (shaded as shown) and vertical velocity (contoured at 10, 15, 20, and 25 m s$^{-1}$); (right) vertical profile of environmental CIN (shaded as shown) and CAPE (contoured at 500, 1000, and 2000 J kg$^{-1}$, with bold contour at 0) vs time. The ordinate for all three panels is the same (time).
FIGURE 2.5: Time series of (a) total upward mass flux (TUMF, kg), and (b) maximum of along-line averaged vertical velocity ($\bar{w}_{max}$ m/s) from t = 3:00-10:00 for the RTF (gray solid), FTR (dashed) and CTL (black solid) simulations.
FIGURE 2.6: Assorted fields for (a) CTL, (b) RTF and (c) FTR simulations at t = 5:30. Center panel contains vertical cross section of along-line averaged vertical velocity ($\bar{w}$, contoured, dashed contours $> 5$ m s$^{-1}$) and potential temperature perturbation ($\theta'$, alternately shaded every 2 K below -2 K) and wind vectors within the developing stable layer and cold pool (m s$^{-1}$, scale vector at upper right). Right hand panel is the along-line averaged u-wind profile at x = 390 km (well ahead of the squall line). Left hand panel is an along-line averaged buoyancy perturbation (relative to the evolving pre-line environment) 1 km behind the system gust front.
FIGURE 2.7: As in Fig. 2.6, but at t = 7:30.
Figure 2.8: As in Fig. 2.6, but at t = 8:30.
FIGURE 2.9: Time-height plots of along-line averaged (a) CAPE (J kg\(^{-1}\)), (b) CIN, and (c) cold pool buoyancy perturbation (\(B'\) m s\(^{-2}\)) between t = 3:00 and 10:00 for CTL simulation. CAPE and CIN are measured at x = 398 km (well ahead of the squall line), while \(B'\) is calculated within the cold pool (relative to the evolving pre-line environment), 1 km behind the gust front. The gray shaded area represents the effective inflow layer discussed in the text.
FIGURE 2.10: Time series of cold pool strength ($C$, ms$^{-1}$ heavy dashed line, control simulations only) and square root of the vertically integrated horizontal vorticity flux ($u\eta|_z$, thin solid line for RTF jet configurations and thin dashed line for FTR jet configurations) for (a) benchmark, (b) NOCOOL, and (c) 513m simulations. The square root of $u\eta|_z$ is shown to make the units comparable to $C$ and both variables are integrated over the effective inflow layer shaded in Fig. 2.9.
Figure 2.11: Vertical cross section of along-line averaged potential temperature ($\theta$, contours every 2 K starting at 294 K), wind vectors (m s$^{-1}$, scale vector in lower right corner) and CAPE (J kg$^{-1}$ shaded as shown) and vertical profiles of LFC height (m, left-hand panel) for (a) RTF and (b) FTR simulations at $t= 8:00$. Heavy black lines denote the top and bottom of the applied LLJ.
Figure 2.12: Time series of vertically integrated system-relative horizontal mass flux for (a) benchmark, (b) NOCOOL, and (c) 513m simulations. The integration is performed from the bottom of the effective inflow layer shaded in Fig. 2.9 to 3 km AGL in (a) and (c) and from the surface to 3 km AGL in (b). An upper bound of 3 km is chosen as it generally represents the vertical extent of air parcels containing CAPE.
Inflow trajectories for benchmark simulations
\( t = 8:00-9:00 \)

**Figure 2.13:** Vertical cross section of parcels trajectories plotted over 1 hour between \( t = 8:00 \) and 9:00 for the (a) CTL, (b) RTF and (c) FTR simulations. The values reported below the figure label refer to the total number of parcels that passed through the updraft region of the respective squall lines during this time interval. Only a subset of these totals are plotted in the interest of clarity.
FIGURE 2.14: As in Fig. 2.5, but for NOCOOL simulations.
**Figure 2.15:** As in Fig. 2.5, but for FTR513, RTF513 and CTL simulations.
FIGURE 2.16: As in Fig. 2.6, but for CTL, RTF513 and FTR513 simulations at $t = 8:30$. 

53
Figure 2.17: Difference plot of across-line (east/west) averaged accumulated rainfall (cm) comparing PARNP and CTLNP simulations (PARNP-CTLNP). The data are grouped into 30 km wide (north/south) bins that are centered on the points shown along the horizontal axis. The dashed vertical line denotes the center of the squall line at y = 300 km.
Figure 2.18: Schematic diagram illustrating the key layers of vertical wind shear participating in the cold pool/shear balance relative to the height of the maximum LLJ winds (Z_{LLJ} heavy dashed line) and cold pool depth (Z_{cp}) over time. Thin lines are representative isentropes that define the cold pool/bore and the shaded area represents the layer of vertical shear. The evolution is broken into three periods corresponding to different points in the squall line’s evolution.
TABLE 2.1: Vector wind differences between 0 - 2.5 km, 0 - 1 km, 1 - 2.5 km, and 0.5 - 1.5 km AGL for CTL, RTF and FTR simulations at t = 8:00.

<table>
<thead>
<tr>
<th>Layer</th>
<th>CTL</th>
<th>RTF</th>
<th>FTR</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-2.5 km</td>
<td>13.64 ms$^{-1}$</td>
<td>13.99 ms$^{-1}$</td>
<td>13.45 ms$^{-1}$</td>
</tr>
<tr>
<td>0-1 km</td>
<td>5.32 ms$^{-1}$</td>
<td>8.08 ms$^{-1}$</td>
<td>2.64 ms$^{-1}$</td>
</tr>
<tr>
<td>1-2.5 km</td>
<td>8.32 ms$^{-1}$</td>
<td>5.91 ms$^{-1}$</td>
<td>10.85 ms$^{-1}$</td>
</tr>
<tr>
<td>0.5-1.5 km</td>
<td>5.49 ms$^{-1}$</td>
<td>10.90 ms$^{-1}$</td>
<td>0.27 ms$^{-1}$</td>
</tr>
</tbody>
</table>
Chapter 3

Observations of mergers between squall lines and isolated supercell thunderstorms

1. Introduction

It has been observed that different organizational modes of convective storms tend to be associated with different severe weather threats (e.g., Gallus et al. 2008). Generally speaking, significant tornadoes and large hail often occur with supercell thunderstorms (e.g., Doswell and Burgess 1993; Davies-Jones et al. 2001) whereas widespread damaging straight-line winds are more frequently produced by linear modes, particularly bow echoes (e.g., Fujita 1978; Przybylinski 1995). In light of this, severe weather forecasters try to anticipate the predominant mode of organization that storms will take once they form, and how that mode may evolve over time. This can help them anticipate which severe weather hazards may occur, and how those hazards may change with time. This becomes complicated, however, in cases where multiple organizational modes are present within a localized area (e.g., French and Parker 2008), especially when these modes merge into a single system. The present work seeks to improve
our understanding of these situations, by investigating the effect that mergers between isolated supercells and squall lines have on storm organization and severe weather production.

a. Background

The majority of past literature dealing with squall line-supercell mergers has consisted of observation-based analysis of individual cases, many of which produced significant tornadoes. Goodman and Knupp (1993) investigated a case from November 1989 wherein a merger between a squall line and isolated supercell coincided with the development of an F4 tornado that struck Huntsville, AL. Using regional composite radar data, observations from a nearby surface mesonet, and visual observations of the storm, the authors demonstrated that tornadogenesis appeared to coincide with an interaction between the supercell and the gust front associated with the squall line’s cold pool. Furthermore, these observations also showed a “distortion” of the squall line’s gust front, resulting from the merger. As the squall line approached the supercell, forward progress of its gust front slowed in the vicinity of the merger and accelerated south of the merger location, effectively appearing to “wrap around” the supercell’s mesocyclone. This suggests that the supercell effectively altered the structure of the squall line during the merger process. Additional studies also suggest a propensity for the supercell to play a dominant role in the merger process. Wolf (1998) analyzed what he described as the “unexpected evolution” of a merger between a supercell and bow echo that produced a large high-precipitation supercell, and continued to produce tornadoes for over an hour after the merger. Similar results were seen by Sabones et al. (1996) and Wolf et al. (1996), both of which documented interactions between squall lines or bowing line segments and supercells coinciding with tornadogenesis.

Squall line-supercell mergers are not always associated with significant tornadoes, nor do
they always promote sustained supercell structures. Several studies have also documented merger events that lead to the development of bow echoes. Fujita (1978) and Sieveking and Przybylinski (2004) both discussed cases where mergers between supercells and developing bow echoes appear to enhance the bow echo and produced widespread damaging winds. In these cases either weak (Sieveking and Przybylinski 2004) or no (Fujita 1978) tornado damage was reported with the merged system. Additionally, Calianese et al. (2002) discussed a case where a merger between a bow echo and high-precipitation (HP) supercell produced significant flash flooding in the Dallas-Fort Worth, TX metro area, illustrating the variety of hazards posed by these merger events. All three of these cases exemplify events where the post-merger storm evolution closely resembled that commonly observed with bow echoes. It is not surprising that mergers between line segments and supercells lead to the development of bow echoes, as past work has shown that the often-observed evolution of HP supercells to bow echoes (e.g., Moller et al. 1990, 1994) may be related to storm mergers in the more general sense (e.g., Klimowski et al. 2004; Finley et al. 2001). In particular, Finley et al. (2001) found that enhanced precipitation following the merger strengthened the supercell’s cold pool, leading to the development of the bow echo.

Thus, a review of previous works on the subject of squall line-supercell mergers reveals a variety of outcomes that can produce a variety of severe weather threats. A broader examination of a larger number of cases seems warranted to develop an understanding of the common evolutions associated with squall line supercell mergers, and what they mean for severe weather production. It would be useful to also further our understanding of storm evolution in cases where supercells merge with well-organized squall lines that are clearly larger in scale than the supercell. The present study looks to address both of these topics by examining a set of 21 cases wherein at least one supercell merges with a well-organized squall line.

Section 2 details the data and methods used in the study. In Section 3 we introduce two
common background environments associated with our merger cases, and discuss the convective organizations, and severe weather production associated with these environments. Finally in Section 4 we synthesize these results in light of past work and provide some concluding remarks and avenues for future study.

2. Data and Methods

Candidate squall line-supercell merger cases were initially identified using archived regional scale composite radar reflectivity data maintained online by the Microscale and Mesoscale Meteorology (MMM) division at the National Center for Atmospheric Research (NCAR) (http://www.mmm.ucar.edu/imagearchive/). Data were reviewed from 2006-2010, focusing on the months of April, May and June and the central United States (including the states of Texas, Oklahoma, Arkansas, Missouri, Iowa, Kansas, Nebraska, South Dakota, Colorado and Wyoming). To be included, a case needed to contain persistent (present at a quasi-steady intensity for at least an hour prior to the merger) linear (defined as length to width ratio > 5:1) and isolated cellular structures that eventually merged. These cases were then further examined using single-site Weather Surveillance Radar 1988-Doppler (WSR-88D, Crum and Alberty 1993) radar reflectivity and radial velocity data in order to confirm that the individual cells identified were indeed supercells (e.g., contained a mesocyclone and relevant reflectivity structures such as hook echoes and/or weak echo regions), and that the two modes actually merged, which was not always clear given the coarser (approximately 15-30 minute) temporal resolution of the composite data. A merger was defined as a permanent union of the 40 dBZ radar reflectivity contours associated with the squall line and supercell. This resulted in 18 merger cases, which we supplemented with three additional cases from outside of the 2006-2010 date range in order to increase the sample size. This increased the total to 21 cases, and a total of 29 merger events
Our primary dataset for analysis was archived level-II WSR-88D reflectivity and velocity data. For each case, data were obtained from all relevant radar sites to cover the duration of the event. These data were processed using the Warning Decision Support System-Integrated Information (WDSS-II, Lakshmanan et al. 2007) software developed collaboratively by the National Severe Storms Laboratory and University of Oklahoma. The initial processing included quality-controlling radar reflectivity to remove non-meteorological echoes, de-aliasing the radial velocity fields and calculating azimuthal shear and radial divergence from the de-aliased velocity data. WDSS-II uses a two-dimensional, local, linear least squares derivative technique to calculate these last two fields, which tends to be more tolerant of noisy data and is less dependent on a feature’s position relative to the radar than other methods for calculating rotational and divergence signatures (Smith and Elmore 2004). The azimuthal shear data were then subjectively sorted to separate data associated with the pre-merger supercell, the pre-merger squall line, and the final merged system. This allowed us to calculate some statistics and time series data for the individual modes. Finally, to better facilitate comparison between cases, the reflectivity and azimuthal shear data were interpolated to constant latitude-longitude-height grids, using the WDSS-II algorithm described by Lakshmanan et al. (2006). This was done to remove the range dependence of the height of the radar beam that occurs when viewing individual radar tilts, and thus facilitate comparisons between cases where storms and mergers were at varying distances from the radar. The grids had a horizontal dimension of 5° x 5°, with a grid spacing of 0.1° and a vertical dimension of 10 km with a grid spacing of 500 m.

The hourly analysis fields from the Rapid Update Cycle (RUC, Benjamin et al. 2004) forecast model were used to investigate the background environments associated with our merger cases. These data are desirable as they provide a complete three-dimensional picture of the atmosphere on an hourly time scale, and a 20 km horizontal grid-spacing. This provides much
higher spatial and temporal resolution than can be found from observed radiosonde soundings alone. Furthermore, these data have been used extensively in the study of mesoscale phenomena, and found to have small errors when compared with available observations (e.g., Thompson et al. 2003; Benjamin et al. 2004). We used the RUC analyses to categorize each case based on its synoptic environment, as will be discussed in Section 3b, and create mean environments for the two dominant patterns. To create these mean environments, the RUC data for each case were first rotated so that the east/west axis was parallel to the mean squall line motion over the 1-hour time period centered on the merger. We then interpolated the rotated data to a 1400 x 1400 km grid centered in time and space on the merger. The grids associated with the two dominant synoptic patterns were then averaged to create the mean environments. The mean grids were examined for signs of convective “contamination” related to the storms that were present, such as signals of surface outflow or perturbations to the wind field in the vicinity of the observed storms, and we are confident that the mean fields represent the large scale environment, rather than any signals related to the storms. For four cases, the 20 km RUC analyses were unavailable, and in their stead the analysis fields from the 40 km grid-spacing RUC model (2 cases), and 32 km grid-spacing North American Regional Reanalysis (NARR, Mesinger et al. 2006, 2 cases) were used to characterize the background environments. These cases were not included in the mean environments.

Finally, to evaluate the hypothesis that these types of merger events produce an evolving severe weather threat, we examined severe weather reports from the National Climate Data Center Storm Events Database associated with each of our cases. While a number of past studies have discussed the shortcomings and caveats associated with this dataset (e.g., Doswell and Burgess 1988; Weiss et al. 2002; Doswell et al. 2005; Verbout et al. 2006; Trapp et al. 2006), it still remains the most complete source of data of this type, and when used carefully can provide some useful input regarding the actual threat posed by these systems from a severe weather
warning standpoint. We subjectively sorted the reports, using radar data to identify which reports corresponded to the pre-merger supercells, the pre-merger squall line (in the vicinity of the merger), and the post-merger system based on their proximity in time and space. Reports from nearby storms not involved in the merger were excluded. The reports were then binned into 15 minute time periods based on a merger-relative framework (i.e., times before/after the merger) to allow comparisons between cases relative to the time when the merger occurred. This was done for each merger event in each case.

3. Results

a. Overview

One of the primary goals of investigating a number of cases was to try to find commonalities associated with these types of squall line-supercell merger events. In reviewing the single-site radar reflectivity data, we found that in an overwhelming majority of the merger events (22 of the 29 identified) the post-merger evolution was characterized by the development of some type of bow echo (Fujita 1978) structure (the “system-scale bowing”, “hybrid”, and “embedded bowing” rows of Table 3.2). The remaining cases (the “other” row of Table 3.2) showed a variety of evolutions ranging from the supercell being absorbed into the squall line and dissipating, to the merged system evolving into a large supercell. While the evolution toward bow echoes was a common occurrence, we found that rather than falling into clear evolutionary archetypes, these events instead tend to fall within a spectrum of convective evolutions, as will be discussed in Section 3c.

We found it more informative, and deemed it potentially more useful in the forecasting sense, to organize our cases based upon the characteristics of their background environments.
Cases were subjectively sorted into two groups based on the strength of the synoptic-scale forcing, in a similar manner to several previous studies (e.g., Johns and Hirt 1987; Evans and Doswell 2001). The first is a “weak synoptic forcing” environment, characterized by a low-amplitude 500 hPa trough and wind maximum $< 30 \text{ m s}^{-1}$, and a weak or non-existent surface cyclone in the vicinity of the merger. This environment effectively characterized 10 cases (the “weakly forced” column in Table 3.2). The second, is a “strong synoptic forcing,” environment, characterized by a high-amplitude 500 hPa trough, a 500 hPa jet streak $> 30 \text{ m s}^{-1}$, and a mature surface cyclone with a central pressure $< 1000 \text{ hPa}$, in the vicinity of the merger. This environment characterized the remaining 11 of our cases (the “strongly forced” column in Table 3.2). We will refer to these respectively as the “weakly forced” (hereinafter WF) and “strongly forced” (hereinafter SF) environments.

As a measure of how effectively the synoptic-scale forcing separates the present cases, a scatter plot of the maximum sea level pressure gradient (a proxy for the strength of the surface cyclone) versus the maximum 500 hPa height gradient (a proxy for the strength of the 500 hPa trough) is presented in Fig. 3.1. The cases divide into two clusters, indicating that the strength of the synoptic forcing effectively separates the merger cases. In the following section, mean features representing these two environments will be presented. In order to limit the impacts of any outliers, these means were computed using a subset of cases that consisted of the most closely clustered distribution, as illustrated for the fields shown in Fig. 3.2. This was necessary step prior to calculating the means as the full distributions for most of the variables examined were not Gaussian. Individual subsets were used to compute each mean field (e.g., different cases were used to compute the mean sea level pressure and mean shear, etc) in order to produce the most representative values. Generally speaking, the basic patterns of the means computed using the subset values were quite similar to the mean patterns computed using all of the cases, however the values differed slightly.
b. **Background environment**

The mean WF environment at the time of merger is characterized by a low-amplitude 500 hPa short-wave trough to the west of the merger location, with a 20-25 m s$^{-1}$ wind maximum over the merger location (Fig. 3.3a). The primary surface feature is a weak warm front oriented nearly parallel to the squall line motion extending through the merger location (Fig. 3.4a). Also notable is the absence of a deep surface cyclone and attendant cold front. This is consistent with surface analyses in several of these cases suggesting that squall lines often formed in weak convergence zones along dry lines or troughs rather than along a strong cold front (not shown). The thermodynamic environment was fairly homogeneous, with a large region of 500-1500 J kg$^{-1}$ of convective available potential energy (CAPE) and $>-100$ J kg$^{-1}$ of convective inhibition (CIN) in the vicinity of the merger location (Fig. 3.5a). The magnitude of the deep layer (0-6 km AGL) bulk shear is approximately 20 m s$^{-1}$ (Fig. 3.6a), which is generally considered to be toward the low-end of values expected for supercells (e.g., Weisman and Klemp 1982; Weisman and Klemp 1984; Thompson et al. 2003). These values are, however, favorable for bow-echoes (e.g., Weisman 1993; Doswell and Evans 2003), especially given the large component parallel to the squall line motion vector. Outside of a localized region of 0-1 km storm-relative helicity (SRH) $>200$ m$^2$s$^{-2}$ in the immediate vicinity of the merger (Fig. 3.6a), SRH values were generally toward the lower end of what is expected for tornadoes and rotating storms (e.g., Thompson et al. 2003) with values around 100 m$^2$s$^{-2}$ over most of the region (Fig. 3.6a). The synoptic scale features of this environment bear a strong resemblance to the “progressive derecho” environment identified by Johns and Hirt (1987), and the “warm season” bow echo environment discussed by Johns (1993), that tend to favor isolated bow echoes rather than long squall lines with embedded bowing structures.

In contrast to the WF environment, the mean SF environment is characterized by a high-
amplitude trough and $>30$ m s$^{-1}$ jet at 500 hPa (Fig. 3.3b) and strong surface cyclone (Fig. 3.4b). The strong surface cold front and dry line (Fig. 3.4b) appear to be primary initiating mechanisms for the squall line in these cases, with analysis of radar fine-lines in several cases (not shown) suggesting that the squall lines were initiated by enhanced forcing when the cold front overtook the dry line as discussed by Dial et al. (2010). The thermodynamic environment for these cases was much more variable than in the WF evolution, with a narrow region of high CAPE ($>1500$ J kg$^{-1}$) and low CIN ($>-50$ J kg$^{-1}$) extending southward from the merger location along the dry line/cold front, while locations further east into the warm sector were characterized by lower CAPE and larger CIN (Fig. 3.5b). The region of low CIN along the linear forcing likely favored squall line formation, by promoting widespread convective storm development. In another significant departure from the WF environment, the SF environment featured a 0-6 km bulk shear $>25$ m s$^{-1}$, favoring supercell storms, and 0-1 km SRH values $>300$ m$^2$s$^{-2}$ increasing the likelihood of tornadic supercells (Fig. 3.6b). In light of this, the strong linear forcing present due to the cold front/dry line interactions likely played a key role in the development of the squall line in what could otherwise be described as a supercell environment (Dial et al. 2010). This synoptic environment is quite similar to the “serial derecho” pattern of Johns and Hirt (1987) and the “dynamic” bow echo environment of Johns (1993). Particularly, the strong linear surface forcing along the cold front suggests an environment more favorable for comparatively longer squall lines with embedded bowing segments, rather than a single, large bow echo. In addition, as noted by Johns (1993) this environment shares a number of similarities with the “classic” Great Plains tornado outbreak pattern, which would suggest that sustained supercell structures may also be favored.

In short, the mean background environments associated with our merger cases are very similar to those commonly associated with different organizations of bow echoes (Johns and Hirt 1987; Johns 1993). In particular, the background synoptic pattern and vertical wind shear
in the WF environment would traditionally favor large, isolated bow echoes, while the synoptic pattern vertical wind shear in the SF environment is more favorable for large squall lines with embedded bowing segments. Both environments feature vertical wind shear profiles that can support isolated supercell storms, however the SF environment would be more favorable for these storms, and likely more supportive of tornadic supercells given the larger 0-1 km SRH values.

It should be noted that this analysis of the background environment focuses on the upper-end mesoscale to synoptic scale features that were present in the cases examined. This neglects any influences that the squall line and/or supercell may be having on the environment which may also be important. Past studies have shown that squall lines can significantly perturb their local thermodynamic and kinematic environment in a variety of ways (e.g. Lafore and Moncrieff 1989; Nicholls et al. 1991; Weisman and Davis 1998; Fovell 2002; Trier and Sharman 2009; Bryan and Parker 2010). Supercells have also been shown to alter the near-storm environment (e.g. Brooks et al. 1994), however these effects tend to be localized near the supercell, and given their relative sizes, it is likely that the squall line’s impact on the supercell’s environment may be more significant. Indeed, a recent study by LaPenta et al. (2005) speculated that the presence of a nearby squall line may have played a role in the development of a tornado with an isolated supercell that later merged with the line. They hypothesized that the presence of a strong line-end vortex, and associated low pressure center, may have caused a local backing of the low-level winds, creating an environment rich in storm-relative helicity, and thus more favorably for tornado development. While the present observations do not permit identifying such processes in the present cases, they may nonetheless play an important role in these types of cases.
c. Reflectivity Analysis

As mentioned earlier, the reflectivity structures associated with these merger events can best be described as covering a spectrum of convective evolutions that frequently produce bow echo structures (Table 3.2). At one extreme, after the merger the entire squall line evolved into a large bow echo, as illustrated in Fig. 3.7a and the examples in Fig. 3.8a-e. We have termed this evolution system-scale bowing (hereinafter SSB). It was observed exclusively in the WF environment, and was the most common evolution in that environment (Table 3.2). In these cases, the squall line and supercell tend to have similar directions of motion (Fig. 3.7a, t=1) and the merger typically results from the squall line overtaking the supercell. As it approaches the supercell, the squall line tends to slow its eastward progress and weaken north of the eventual merger location\(^1\) (Fig. 3.7a, t=2). The squall line typically merges with the rear flank of the supercell, leading first to a “Y-shaped” echo (as the forward flank precipitation associated with the supercell continues to extend eastward from the squall line; Fig. 3.7a, t=3, Fig. 3.8c). As the merger progresses, it is associated with an increase in radar reflectivity values near and south of the merger location, and the squall line begins to take on a “S-shape” (Fig. 3.7a, t=4, Fig. 3.8d). Eventually a “swirl” pattern becomes evident near the north end of the squall line and the bowing becomes more pronounced (Fig. 3.7a, t=5). By this point the merger location/remnant supercell now represents the north end of the squall line, and any remaining radar echoes north of this point have weakened considerably. Typically, a large “comma-echo” (Fujita 1978) emerges as the bowing structure becomes most evident within 1-2 hours following the merger (Fig. 3.7a, t=6, Fig. 3.8e).

This evolution typically occurs in cases where a single supercell was present and merged with the squall line, and the post-merger evolution appears similar to that detailed in several

\(^1\)For the sake of simplicity, we will assume an eastward moving squall line oriented north/south, as shown in the schematic in Fig. 3.7.
past studies (e.g., Fujita 1978; Sieveking and Przybylinski 2004). Additionally, there are a number of qualitative similarities between the reflectivity structures in the SSB evolution, and those associated with the often observed high-precipitation supercell-to-bow echo transition (e.g. Moller et al. 1990; Moller et al. 1994). This includes the development of strong bowing south of the remnant supercell circulation and the presence of “swirl” patterns in the reflectivity field that appears to be associated with this circulation. That the SSB evolution was the preferred outcome in the WF environment is not surprising, as this environment strongly resembles one associated with large bow echoes (e.g., the “progressive derecho” of Johns and Hirt 1987; Johns 1993). In fact, a number of SSB cases exhibited varying degrees of bowing prior to the merger (e.g., Fig. 3.8b), suggesting that the environment may play a strong role in governing the outcome of the merger.

At the other end of the evolutionary spectrum, we observed what we have termed embedded bowing (hereinafter EMB). In these cases, following the merger, a small scale bowing segment develops along the squall line but the entire line does not evolve into a bow echo (Fig. 3.7b, Fig. 3.8f-j). This evolution was only observed in the SF environment, in 4 cases total (Table 3.2). In these situations the supercell typically has a direction of motion that is largely parallel to the squall line’s major axis (Fig. 3.7b, t=1). As the squall line approaches it typically weakens or “breaks” in the vicinity of the supercell (Fig. 3.7b, t=1-2, Fig. 3.8g). The forward flank of the supercell then will merge first at the northern end of this break, followed by the rear flank of the supercell merging with the line south of the break (Fig. 3.7b, t=3, Fig. 3.8h). Following the merger the supercell remains evident as an embedded structure within the squall line, typically characterized by a notch-like feature within the line (Fig. 3.7b, t=4, Fig. 3.8h-i). Eventually this feature evolves into a small scale bow echo embedded within the larger line, sometimes also exhibiting reflectivity swirl features as discussed for the SSB evolution above, albeit on a smaller scale (Fig. 3.7b, t=5-6, Fig. 3.8j). The whole process from initial merger to embedded
bowing structure can take upwards of 1 - 2 hours and thus influence the local organization of the squall line well after the merger occurs.

The EMB cases often occur in situations where multiple supercells are present ahead of a comparatively long squall line, and thus multiple mergers can occur within a single case leading to a line-echo wave pattern (LEWP, Nolen 1959) organization to the squall line. It should also be noted that in these cases there were often additional bowing structures present away from the merger (i.e., the gray arrow in Fig. 3.8j). Thus it would appear that in the EMB cases the merger was not a necessary condition to get a bowing segment, but it still may have served as a trigger to promote bowing at a particular location along the squall line. Additionally, it makes sense that the EMB evolution was observed exclusively in the SF environment as this environment is very similar that commonly associated with LEWP-type squall lines (e.g., the “serial derecho” of Johns and Hirt 1987; Johns 1993). Thus, as in the SSB evolution, the post-merger in the EMB cases appears to be strongly governed by the background environment as well.

While the SSB and EMB evolutions represent the extremes on the spectrum of post-merger organizations, many cases contained features common to both of these evolutions, and are thus best described as a “hybrid” of the two (Fig. 3.7c). We found examples of this evolution in both the WF and SF environments (e.g Fig. 3.9a-e, and f-j), although it was more common in the SF environment, representing the most frequently observed evolution in that environment (Table 3.2). These cases often begin with a large squall line, as in the EMB evolution (Fig. 3.7c, t=1), however as the supercell approaches the line weakens and ultimately dissipates to the north of the merger point (Fig. 3.7c, t=2 Fig. 3.9a-c, and g-h). As the supercell merges, it becomes the north end of the squall line, proceeding through the Y- and S-shaped echo evolutions common to the SSB evolution (Fig. 3.7c, t=3-4, Fig. 3.9c-d, and h-i) and eventually developing a small-scale bow and comma echo (Fig. 3.7c, t=5-6, Fig. 3.9e, and j).
difference, however, is that the resultant bow remains similar in scale to the merged supercell, and while the line may reorient south of the merger, it does not evolve into a large bow echo, as seen in the SSB cases (c.f., Figs. 3.7a, c, t=3-6). In fact, in some cases additional embedded bowing segments are observed away from the merger location, not unlike what is observed in the EMB cases, and in several cases multiple mergers occurred and followed the hybrid evolution. While the observation of this evolution in both the WF and SF environments implies that the background environment is not the primary control, it still appears to play an important role. The hybrid cases observed in the WF environment appeared closer to the SSB end of the spectrum (c.f. Figs 3.8a-e and 3.9a-e), while those observed in the SF environment had more similarities with the EMB evolution (c.f. Figs 3.8f-j and 3.9f-j). Thus for a given environment the delineation between the SSB or EMB and hybrid evolutions may ultimately come down to storm-scale details such as the relative size or maturity of the squall line/supercell to the relative location of the merger (i.e., near the north end vs. near the center of the line).

To summarize, the overwhelming majority of the merger cases that we examined produced an ultimate storm organization that resembled a bow echo. The behavior ranged from the development of a large bow echo (the SSB evolution), seen in most of the WF cases, to a small-scale bowing segment embedded within a larger line (the EMB evolution) seen in several SF cases. The remaining cases, primarily in the SF environment, evolved as a hybrid of these extremes. These evolutions are consistent with non-merger bow echo evolutions observed for similar weak- and strongly-forced environments by Johns and Hirt (1987) and Johns (1993). In these studies, large-scale bow echoes (i.e., the “progressive” derecho of Johns and Hirt 1987, their Fig. 3) tended to be associated with more weakly forced events, while lines with smaller-scale embedded bow echoes (i.e., the “serial” derecho of Johns and Hirt 1987, their Fig. 6) were associated with strongly forced events.
d. Velocity analysis

In addition to examining the reflectivity features associated with squall line-supercell merger events, we were also interested in examining what happens to the velocity signatures associated with these two modes when the merger occurs. Specifically, how does the existing mesocyclone associated with the supercell evolve as the merger takes place, and what influence might this have on subsequent storm organization? In order to facilitate comparison among multiple cases, we focused our analysis on azimuthal shear calculated from the de-aliased radial velocity data as a means of identifying and tracking rotational features. As discussed in Section 2, these data were interpolated to a three-dimensional grid and subjectively sorted, so that only those data associated with the pre-merger supercell and merged system were evaluated.

One of the most basic questions pertaining to the evolution of the supercell’s mesocyclone in these cases is whether or not it remains evident following the merger (i.e., does the mesocyclone persist within the merged system?). To address this question, we tracked azimuthal shear values over time associated with each supercell and its subsequent merged system. This was done both by looping images, and looking at plan view plots of the accumulated azimuthal shear over time to produce “rotation tracks” associated with these features (e.g., Fig. 3.10). In most cases, the rotational signature initially associated with the supercell could be tracked following the merger. The exact evolution of the rotational features post-merger varied considerably, with azimuthal shear weakening (e.g., Fig. 3.10b, d), remaining constant (Fig. 3.10c), or intensifying (Fig. 3.10a) after the merger depending on the case. Additionally the direction of the rotation tracks varied after the merger as well, although most cases either saw little change in the path of rotation (approximately 50% of cases, e.g., Fig. 3.10b, d), or a turn to the right relative to the initial supercell motion (approximately 36% of cases, e.g., Fig. 3.10a, c). Most of the cases exhibiting no change in direction occurred in the SF environment, while examples
in both environments were found that turned to the right. We interpret this as representing two different avenues for supercell behavior post-merger. The cases where the rotation track is largely unchanged indicates a sustenance of supercell features post-merger, as the storm does not appear to be disrupted or altered by the squall line. This is consistent with the “embedded supercell” period of the evolution seen in some of the SF cases (e.g., the “EMB” evolution, Fig. 3.7b, T=4). The cases where there is a pronounced turn to the right likely indicate that the supercell has acquired a motion vector similar to the squall line’s, suggesting that the squall line is playing a dominant role in the merger in these cases. This would indicate that the supercell has been disrupted in some way by the squall line, and its remnant supercell circulation becomes part of the squall line, moving with the squall line’s gust front. Which supercell pathway occurs may depend on the storm scale details of a given event, including the strength of the squall line’s cold pool.

To better understand the details of the evolution of rotation features associated with the merger case, the maximum azimuthal shear associated with the supercell, and eventual merged system was examined over time and height for each merger case. From the example WF cases shown in Fig. 3.11 it is clear that the details of this evolution vary case-to-case, however there are some common features that stand out. In most of the WF cases, azimuthal shear is observed to weaken around the time of merger (t=0 in Fig. 3.11a-f). For some cases this occurs just after the merger (e.g. Fig. 3.11a, b, d) while in other cases it appears to precede, or occur coincident with the merger (e.g. Fig. 3.11c, e, f). Following this initial decline in rotation around the merger time, a subsequent re-intensification of rotation was often observed (e.g. Fig. 3.11a, c, d, e, f). Generally this re-intensification was concentrated in lower levels than the pre-merger supercell rotation (e.g. generally below 3 km AGL, Fig. 3.11a, c, d, f), although in some cases a strong, deep, rotational feature developed (Fig. 3.11e). We interpret the initial decline in azimuthal shear as resulting from a broadening of the circulation that appears to closely follow...
the merger in a number of the WF cases (e.g., Fig. 3.12a-d, and e-h). As the diameter between
the maximum inbound and outbound winds increases, azimuthal shear (e.g., vertical vorticity)
decreases. The comparatively broad post-merger circulations were generally observed north of
the bow, collocated initially with the “S”-shaped echo (Fig. 3.12c, g) and later the reflectivity
swirl/comma echo structures (Fig. 3.12d, h) common to the observed reflectivity evolutions.
In most cases this circulation was strongest in the low-mid levels (e.g., at or below 3 km AGL,
Fig. 3.12), consistent with shift in maximum azimuthal shear to lower levels post-merger (Fig.
3.11). Qualitatively, these post-merger circulations appear very similar to the line-end vortices
often observed with bow echoes (e.g., Weisman and Davis 1998; Atkins et al. 2004). From
the observations it is unclear whether these features facilitate the developing bow echo, or are
instead a result of it.

The mechanism responsible for re-intensification of low-level rotation later on (e.g., after
t=+30 min. in Fig. 3.11a, t=+15 min. in Fig. 3.11d, e) is less clear. As discussed above, the
post-merger circulations in many of the WF cases appeared qualitatively similar to line-end
vortex structures. However, line-end vortices tend to reside in the mid levels (e.g., 3-6 km
AGL, Weisman and Davis 1998; Atkins et al. 2004), whereas the features in Fig. 3.11a, c-f
become maximized in low-levels over time (e.g., below 3 km AGL by t = 40 min.). This is
more often observed with squall line mesovortices (e.g., Funk et al. 1999; Weisman and Trapp
2003; Trapp and Weisman 2003; Atkins et al. 2004; Atkins and St. Laurent 2009a; Atkins
and St. Laurent 2009b), which tend to be smaller in scale and focused in lower levels. Figure
3.13 provides an example of mesovortices associated with one of our cases. Following the
merger, the original supercell circulation (labeled SC in Fig. 3.13a-c) moves rearward relative
to the developing bow echo, while multiple mesovortices develop just north of the apex of the
bow echo (labeled MV1-MV3 in Fig. 3.13b-d) and move rearward along a similar path as
the remnant supercell circulation. Similar to the initial supercell circulation, several of these
mesovortices widen over time, appearing to evolve toward line-end vortices (e.g. MV1 Fig. 3.13c-d). Thus the post-merger low-level maximum in rotation may result from the presence of line-end vortex and mesovortex features, both of which were present in our cases, and both of which are often observed with bow echoes.

Several common rotational features were also observed in the SF cases, however these represent a slightly different evolution than that seen in the WF cases. First, in a number of SF events, azimuthal shear is observed to increase prior to the merger at varying depths throughout the troposphere (e.g. Fig. 3.14b, d, e, f). While this may simply be capturing fluctuations in intensity common to the life-cycle of supercell thunderstorms, it is also possible that the squall lines in these cases are altering the local environment in a way that favors storm rotation, as hypothesized by LaPenta et al. (2005). While the present observations are insufficient to ascertain to what extent this may be occurring, it has been well documented that squall lines can perturb the nearby wind and thermodynamic fields (e.g. Lafore and Moncrieff 1989; Nicholls et al. 1991; Weisman and Davis 1998; Fovell 2002; Trier and Sharman 2009; Bryan and Parker 2010), and the impact that such changes may have on nearby storms deserves further consideration in a future study.

A second common feature to these cases, which shares a similarity with the WF cases, is that the strongest rotation generally becomes confined to lower levels (e.g. below 3 km AGL) following the merger (Fig. 3.14a, b, d, e, f). However, in contrast to the WF cases, there is no significant weakening of the initial rotation prior to the development of this low-level feature. Rather, it appears that the low-level rotation gradually becomes dominant as the mid- and upper level rotation weakens. As with the WF cases, an analysis of the actual radial velocity data sheds some light on how to interpret this behavior. In many of the SF cases the post-merger circulation does not appear to broaden as much as those in the WF cases (e.g., Fig. 3.12i-l, and m-p), which may account for the maintenance of strong rotation following the merger. This is
more consistent with the maintenance of an embedded supercell type feature, as suggested in the EMB reflectivity evolutions.

e. Storm Reports

As a means of quantifying the impact of squall line-supercell mergers in terms of severe weather production, we examined storm reports associated with the each of our cases. This includes reports of tornadoes, wind $> 25 \text{ m s}^{-1}$ (50 kt) and hail $> 2.0\text{cm} (0.75")$ in diameter associated with the isolated supercell(s) and the portion(s) of the squall line involved in the merger(s)$^2$, and the subsequent merged system(s) in each case. Fig. 3.15 shows these reports plotted over a merger relative time frame for the WF (Fig. 3.15a) and SF (Fig. 3.15b) environments. In both cases the peaks in tornado reports occur at or just before merger time, the peak in hail reports approximately 30-60 minutes prior to the merger, and the peak in severe wind reports approximately 60-90 minutes after the merger begins. The peak in hail reports pre-merger suggests that these are likely associated with the isolated supercells, while the peak in severe wind gusts post-merger implies that they were generated by the merged system, organized as a bow-echo. These results are very much in line with past research connecting severe weather type to convective mode (e.g., Gallus et al. 2008). The overall distribution of severe reports is approximately centered on the merger in the WF environment (black curve, Fig 3.15a), while it is shifted approximately 15 minutes pre-merger in the SF environment (black curve, Fig 3.15b), owing to a larger number of pre-merger tornadoes with the SF cases. Interestingly, the WF cases produced a great deal more severe weather reports (335) than the SF cases (207), owing primarily to a larger number of damaging wind reports (203 vs. 72) in the WF cases.

$^2$Only reports that occurred with the portion of the squall line eventually involved in the merger were counted, as in some cases the squall line extended for 10s or 100s of km away from the merger location. This was done subjectively using radar animations to track the section of the squall line eventually involved in the merger backwards in time, and generally was restricted to the time period where the merging supercell was also present.
Thus, for the cases investigated, the SF cases tended to produce more tornadoes, while the WF events produced a more widespread severe weather threat overall.

Because several past studies (e.g., Goodman and Knupp 1993; Sabones et al. 1996; Wolf et al. 1996; Wolf 1998) have investigated the role of squall line-supercell mergers as a trigger for tornadogenesis, it is of interest to look in detail at the tornado reports associated with these cases. In the WF cases, the general trend is one of increasing tornado reports leading up to the merger, with the peak occurring just after the merger followed by a gradual decline (Fig. 3.15a). The majority of these reports occur over a window from approximately 30 minutes pre-merger through approximately 60 minutes post-merger. This appears to indicate that the merger plays a key role in producing the tornadoes that do occur in this environment, which would be in line with the past research cited above. In the SF cases, tornado reports increase rather rapidly beginning approximately two hours prior to the merger, with the overall trend (e.g. the red line in Fig. 3.15b) suggesting that reports peak approximately 45 minutes prior to the merger, followed by a fairly quick decline following the merger. In these cases the majority of reports occur over the 90 minutes prior to the merger, with very few tornado reports post-merger. During this 90 minute window there are two peaks in tornado activity, one approximately 60 minutes pre-merger, and a second just prior to the merger. We attribute the first peak (around t = -60 in Fig. 3.15b) to tornadoes associated with the isolated supercells, which is to be expected in the SF environment as the large 0-1 km SRH values and strong vertical shear in these cases (e.g., Fig. 3.6) tend to favor tornadoes (e.g., Rasmussen and Blanchard 1998; Thompson et al. 2003). The secondary peak (around t=-15 in Fig. 3.15b) would appear to represent an increase in tornado activity associated with the merger. Thus for both environments, there appears to be a signal indicating an enhanced tornado threat corresponding with the squall line-supercell merger.

Tornadoes continue to occur after the merger takes place for both environments, particularly
within the first 60 minutes following the merger (Fig. 3.15), however as illustrated by Fig. 3.16, these are generally weaker (< EF2) and shorter track (path lengths generally less than 10 km). This indicates that the post-merger tornadoes tend to behave similarly to those often observed with squall lines and bow echoes, generally being weaker and shorter lived and occurring early in the bow-echo’s lifetime (e.g., Fujita 1978; Forbes and Wakimoto 1983; Wakimoto 1983; Atkins et al. 2004; Trapp et al. 2005). In summary, the post-merger severe weather threat appears to fit very well with that often observed with bow echoes, namely a large threat of widespread damaging winds along with weaker, short-lived tornadoes.

Finally, the increase in tornado activity leading up to merger time in the SF cases (Fig. 3.15b) is quite striking. While some of this signal is no doubt due to the fact that some of the supercells did not mature until this time frame (i.e., the supercell developed not long before the merger), it also corresponds to the increase in low-level rotation seen the SF cases (Fig. 3.14). Furthermore, in several of the cases the longest-track (although not necessarily strongest) tornado of the event occurred with a supercell that later merged with the squall line. The average path length for tornadoes associated with the supercells that merged was over twice as long (11 km) as that associated with supercells that were present in the region, but did not merge with the squall line (4.9 km). This may indicate that some type of pre-merger interaction between the two storms favors tornado genesis and sustenance in the isolated supercell. Past studies have noted apparent increases in storm intensity (Przybylinski 1995) and development of tornadoes as squall lines and supercells converge (e.g., MacGorman and Nielsen 1991; Carey et al. 2003; LaPenta et al. 2005), which would be in line with the idea that the presence of the squall line is in some way impacting the evolution of the supercell. While elucidating the details of these impacts is beyond the scope of the present work, it clearly deserves additional attention in a future study.
4. Conclusions and future work

A radar and RUC analysis-based study has revealed two basic environments in which squall line-supercell mergers occur, one characterized by strong synoptic forcing and strong shear (the SF environment), and the other by weak synoptic forcing and weak-moderate shear (the WF environment). Across these two environments, a spectrum of convective evolutions were observed, generally leading to the development of bow echo structures following the merger. For cases in the WF environment this was most often characterized by the entire squall line evolving into a large bow echo following the merger (the SSB evolution). In the SF environment, a handful of cases followed an evolution that produced small-scale bowing segments embedded within a larger squall line after the merger (the EMB evolution), while the majority exhibited an evolution best described as a hybrid of the SSB and EMB (the “hybrid” evolution). Analysis of radial velocity data revealed that the evolution of rotational features during the merger varied between the WF and SF environments as well. In general, an initial weakening of the supercell’s mesocyclone was observed in the WF cases, associated with a broadening of the circulation as the merger occurs. This appeared qualitatively similar to the development of a line-end vortex associated with the developing bow echo. In the SF cases, strong low-level rotation was maintained and the initial supercell circulation did not broaden as dramatically post-merger, suggesting more of an embedded supercell structure. In both environments, the strongest rotation became concentrated in lower levels after the merger occurred.

A number of past studies have examined cases where the types of mergers discussed in this paper appear to lead to tornadogenesis. Indeed, both environments produced a peak in tornado activity around merger time, with the peak in tornado reports in the WF cases being nearly centered on the merger, suggesting that the merger may play an important role in tornado development in this environment. Also, there appeared to be a pronounced increase in
tornado production in the SF cases over the 60-90 minutes prior to merger, corresponding to an increase in low-level rotation seen in these cases. The possibility that the proximity of the approaching squall line may be related to this signal (e.g. by altering the low-level shear profile) is something that needs to be addressed in a future study. It should be noted, though, that for both environments, the majority of the strongest and/or longest-lived tornadoes occurred with the isolated supercells prior to the merger, and that the trend in both tornado intensity and longevity declined quite rapidly following the merger as bow echo structures developed. This is very similar to what is commonly observed with tornadoes associated with squall lines and bow echoes, and leads us to argue that the tornado threat associated with the merged systems is not unlike the tornado threat commonly associated with bow echoes that occur absent a merger with a supercell.

Additionally, it is important to point out that most of these cases also included severe weather reports away from the storms of interest associated with the merger, either from additional supercells or isolated storms that did not merge, or from other portions of the squall line away from the merger. Seeing as the mean environments identified with these cases bear a strong resemblance to common severe weather patterns identified in the literature, this is not all that surprising. The key point is not so much that the merger is the source for severe weather production in these cases, but rather a source for severe weather. In particular, the merger appears to represent a location favorable for tornadoes at or just before merger time, followed by damaging winds within the 30-60 minutes post-merger. While it would be tempting to claim that the merger location provides a more focused severe weather threat relative to the rest of the squall line, akin to recent research relating squall line mesovortices to swaths of wind damage (e.g., Wakimoto et al. 2006; Wheatley et al. 2006), the present data are insufficient to justify such a claim. More detailed damage survey data would be necessary to effectively determine if the merger location poses an increased threat for severe weather compared to other locations.

80
along the squall line.

Finally, the present results have provided some insights as to some of the common characteristics of squall line-supercell mergers, however the limitations of the observed data used preclude a great deal of in-depth analysis regarding the storm-scale processes at work. In particular, we are interested in more fully understanding why low-level rotation increases during the merger. Based on the present results, and past studies of meso-vortex and low-level mesocyclone development, we hypothesize that the horizontal vorticity present in the squall line’s cold pool may be an important vorticity source for this enhanced low-level rotation. Additionally, given that the supercell structure appeared to persist to some degree in many of the present cases it is also of interest to determine just how the supercell is sustained in these cases rather than being overwhelmed by the squall line’s cold pool. Goodman and Knupp (1993) speculated that outflow from the supercell effectively “blocked” the squall line’s gust front in their analysis of a merger cases, a finding that appears consistent with the present observations of squall line weakening in the vicinity of the merger. To address these hypotheses, and provide a more complete understanding of squall line-supercell mergers, we will now delve into some of the storm-scale processes at work in these types of merger cases using idealized numerical simulations.
FIGURE 3.1: Scatter plot of RUC analysis maximum sea-level pressure gradient (Pa/km) compared to maximum 500 mb height gradient (m/km) for (blue dots) weakly forced cases and (red triangles) strongly forced cases. Maximum values were computed over a 1400 x 1400 km domain centered on the merger.
Figure 3.2: Distributions of mean values of RUC analysis (a) sea-level pressure gradient (Pa km$^{-1}$), (b) mean 500 hPa height gradient (m km$^{-1}$), (c) mean 0-6 km shear (m s$^{-1}$), and (d) mean CAPE (j kg$^{-1}$). Means were computed over a 1400 x 1400 km area centered on the merger. Dashed red and blue boxes denote the portions of the distributions used to compute the means in Figs. 3.3-3.6.
**Figure 3.3:** Mean RUC analysis 500 hPa height (contours, m), wind barbs (flag = 25 m s\(^{-1}\), full barb=5 m s\(^{-1}\), half barb= 2.5 m s\(^{-1}\)), and wind speed (shaded, m s\(^{-1}\)) for (a) WF and (b) SF environments. The black “X” marks the mean merger location.
Figure 3.4: Mean RUC analysis sea level pressure (hPa, solid contours), surface temperature (C, shaded as shown), dew point temperature (C, dashed contours) and wind barbs (flag = 25 m s\(^{-1}\), full barb=5 m s\(^{-1}\), half barb= 2.5 m s\(^{-1}\)), for (a) WF and (b) SF environments at the time of merger. Approximate locations of the surface warm front, cold front, and dry line are annotated with traditional symbols. The black “X” marks the mean merger location.
FIGURE 3.5: Mean RUC analysis surface-based CAPE (shaded as shown, J kg$^{-1}$) and CIN (contours, J kg$^{-1}$), for (a) WF and (b) SF evolutions at the time of merger. The black “X” marks the mean merger location.
Figure 3.6: Mean RUC analysis 0-1 km AGL storm-relative helicity (shaded as shown, m²s⁻²), 0-6 km AGL bulk shear magnitude (contours, m s⁻¹) and bulk shear (wind barbs, flag = 50 m s⁻¹, full barb = 10 m s⁻¹, half barb = 5 m s⁻¹), for (a) WF and (b) SF evolutions at the time of merger. The black “X” marks the mean merger location.
Figure 3.7: Schematic diagrams illustrating the (a) SSB evolution (b) EMB and (c) hybrid evolutions as they would appear on radar (gray shading denotes higher radar reflectivity values). The dashed arrows at T=1 represent initial supercell motion vectors.
Figure 3.8: Examples of (a-e) the system-scale bowing evolution in a WF environment and (f-j) the embedded bowing evolution in a SF environment. Data are WSR-88D 0.5° tilt radar reflectivity from (a)-(e) Fort Worth, Texas (KFWS) between 2333 UTC 5 May 1995 and 0233 UTC 6 May 1995 and (f)-(j) Little Rock, Arkansas (KLZK) between 0205 and 0339 UTC 5 February 2008.
FIGURE 3.9: Examples of the hybrid evolution in (a-e) a WF environment and (f-j) a SF environment. Data are WSR-88D 0.5° tilt radar reflectivity from (a) Amarillo, Texas (KAMA) at 0302 UTC 16 May 2003, (b-e) Frederick, Oklahoma (KFDR) from 0335-0533 UTC 16 May 2003, and (f-j) Topeka, Kansas (KTXW) between 2203 UTC 23 March 2009 and 0008 UTC 24 March 2009.
FIGURE 3.10: Maximum azimuthal shear (s$^{-1}$ shaded as shown) accumulated over time to produce rotation tracks associated with the supercell and merged system from (a) the 18 April 2009 WF case, (b) supercell 1 in the 10 November 2002 SF case, (c) the 30 May 2008 WF case, (d) supercell 4 in the 6 February 2008 SF case. The vertical dashed black lines indicate the longitude of the merger in each case.
FIGURE 3.11: Time versus height plots of maximum azimuthal shear (contoured, (s\(^{-1}\)), color scheme on right side of figure) associated with the isolated supercell (pre-merger) and merged system (post-merger) for (a) 6 May 1995, (b) 21 April 2007, (c) 9 May 2008, (d) 29 May 2008 (merger 1), (e) 18 April 2009, and (f) 6 May 2009 WF cases. Time is in a merger-relative framework, with t=0 corresponding to the merger time, which is annotated with a vertical black line.
Figure 3.12: Constant-height, ground-relative WSR-88D velocity data (shaded as shown m s$^{-1}$) and 45 dBZ radar reflectivity contour from: (a-d) Frederick, Oklahoma 15 May 2003 at 3 km AGL, (e-h) Vance Air Force Base, Oklahoma, 18 April 2009 at 2 km AGL, (i-l) Wichita, Kansas, 25 May 2008 at 3 km AGL and (m-p) Topeka, Kansas 23 March 2009 at 2 km AGL. In all panels dashed circles represent approximate diameters of the circulation features initially associated with the pre-merger supercell. Black arrows point toward the radar location in panels where it is outside the plotting area, otherwise the radar location is denoted by a black “X”. Radar reflectivity from the cases in (a-d) and (m-p) is also presented in Figs. 3.9a-e and 3.9f-j respectively.
Figure 3.13: (a-d) Constant-height, ground-relative WSR-88D velocity data (m s$^{-1}$, positive (outbound) values shaded as shown, negative (inbound) values contoured using the same color scheme) and (e-f) radar reflectivity from Fort Worth, Texas (KFWS) at 3 km AGL on from 0054-0227 UTC 6 May 1995. In all panels dashed circles represent the approximate diameters of the circulation initially associated with the pre-merger supercell (labeled SC) and subsequent mesovortices (labeled MV). White (black) arrows point toward the radar location in panels b-d (f-h) where it is outside the plotting area, otherwise the radar location is denoted by a black “X”.
As in Fig. 3.11, but for the (a) 10 November 2002 (merger 3), (b) 5 May 2007, (c) 5 February 2008 (merger 3), (d) 10 February 2009, (e) 23 March 2009, and (f) 26 April 2009 SF cases.
Figure 3.15: Storm reports vs time for (a) WF and (b) SF environments. Black dots represent all reports, red diamonds tornado reports, green triangles hail reports and blue asterisks wind reports. Colored lines correspond to a 6th-order polynomial trend line fit to the respective data. The vertical gray line denotes the merger time.
**Figure 3.16:** Enhanced Fujita (EF) scale rating (triangles) and path length (squares) vs time for tornadoes in all merger cases (WF and SF). The black and gray lines represent a 4th-order polynomial trend line fit to the EF scale and path length reports, respectively. The vertical gray line denotes the merger time.
TABLE 3.1: Details for each merger event, letters following the date denote cases with multiple mergers. Time, lat. and lon. refer to the approximate time and latitude/longitude that the merger occurred, environ. refers to the two synoptic environments discussed in 3b, and type refers to the observed post-merger evolutions discussed in 3c with SSB, EMB, HYB, and OTR corresponding to the "system-scale bowing", "embedded-bowing", "hybrid" and "other" evolutions, respectively.

<table>
<thead>
<tr>
<th>date</th>
<th>location</th>
<th>time</th>
<th>lat.</th>
<th>long.</th>
<th>environ.</th>
<th>type</th>
</tr>
</thead>
<tbody>
<tr>
<td>06 May 1995</td>
<td>Texas</td>
<td>01:00</td>
<td>32.79</td>
<td>-97.07</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>10 November 2002a</td>
<td>Indiana/Ohio</td>
<td>20:54</td>
<td>41.11</td>
<td>-84.35</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>10 November 2002b</td>
<td>Indiana/Ohio</td>
<td>22:12</td>
<td>41.1</td>
<td>-83.23</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>10 November 2002c</td>
<td>Indiana/Ohio</td>
<td>23:56</td>
<td>41.11</td>
<td>-81.7</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>16 May 2003</td>
<td>Texas/Oklahoma</td>
<td>04:02</td>
<td>35.44</td>
<td>-100.17</td>
<td>WF</td>
<td>HYB</td>
</tr>
<tr>
<td>21 April 2007</td>
<td>South Dakota</td>
<td>02:25</td>
<td>45.34</td>
<td>-98.27</td>
<td>WF</td>
<td>OTR</td>
</tr>
<tr>
<td>22 April 2007</td>
<td>Texas</td>
<td>01:35</td>
<td>34.67</td>
<td>-101.99</td>
<td>SF</td>
<td>OTR</td>
</tr>
<tr>
<td>06 May 2007a</td>
<td>Kansas</td>
<td>01:49</td>
<td>38.17</td>
<td>-98.91</td>
<td>SF</td>
<td>OTR</td>
</tr>
<tr>
<td>06 May 2007b</td>
<td>Kansas</td>
<td>02:10</td>
<td>37.94</td>
<td>-99.12</td>
<td>SF</td>
<td>OTR</td>
</tr>
<tr>
<td>09 May 2007</td>
<td>Texas</td>
<td>23:38</td>
<td>30.97</td>
<td>-99.37</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>06 February 2008a</td>
<td>Arkansas</td>
<td>00:31</td>
<td>36.06</td>
<td>-91.93</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>06 February 2008b</td>
<td>Arkansas</td>
<td>00:06</td>
<td>35.35</td>
<td>-92.73</td>
<td>SF</td>
<td>EMB</td>
</tr>
<tr>
<td>06 February 2008c</td>
<td>Arkansas</td>
<td>01:44</td>
<td>34.96</td>
<td>-91.67</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>06 February 2008d</td>
<td>Arkansas</td>
<td>02:39</td>
<td>34.25</td>
<td>-91.37</td>
<td>SF</td>
<td>EMB</td>
</tr>
<tr>
<td>09 May 2008</td>
<td>Kansas</td>
<td>02:52</td>
<td>37.87</td>
<td>-98.29</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>24 May 2008a</td>
<td>Kansas</td>
<td>03:35</td>
<td>37.58</td>
<td>-99.2</td>
<td>SF</td>
<td>EMB</td>
</tr>
<tr>
<td>24 May 2008b</td>
<td>Kansas</td>
<td>04:49</td>
<td>37.97</td>
<td>-98.36</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>30 May 2008</td>
<td>Nebraska</td>
<td>03:45</td>
<td>40.25</td>
<td>-97.33</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>11 February 2009</td>
<td>Oklahoma</td>
<td>2:10</td>
<td>34.5</td>
<td>-96.94</td>
<td>SF</td>
<td>EMB</td>
</tr>
<tr>
<td>23 March 2009</td>
<td>Kansas</td>
<td>23:07</td>
<td>39.71</td>
<td>-96.33</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>11 April 2009</td>
<td>Texas</td>
<td>22:43</td>
<td>33.12</td>
<td>-102.82</td>
<td>SF</td>
<td>OTR</td>
</tr>
<tr>
<td>18 April 2009</td>
<td>Oklahoma</td>
<td>21:35</td>
<td>36.83</td>
<td>-98.82</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>26 April 2009</td>
<td>Kansas</td>
<td>23:36</td>
<td>38.13</td>
<td>-97.39</td>
<td>SF</td>
<td>HYB</td>
</tr>
<tr>
<td>03 May 2009</td>
<td>Louisiana/Mississippi</td>
<td>14:36</td>
<td>31.93</td>
<td>-91.38</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>06 May 2009</td>
<td>Arkansas</td>
<td>06:45</td>
<td>34.62</td>
<td>-91.09</td>
<td>WF</td>
<td>HYB</td>
</tr>
<tr>
<td>10 June 2009</td>
<td>Texas</td>
<td>23:40</td>
<td>33.38</td>
<td>-97.37</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>12 June 2009a</td>
<td>Oklahoma</td>
<td>16:34</td>
<td>34.02</td>
<td>-94.66</td>
<td>WF</td>
<td>OTR</td>
</tr>
<tr>
<td>12 June 2009b</td>
<td>Oklahoma</td>
<td>17:50</td>
<td>35.21</td>
<td>-94.16</td>
<td>WF</td>
<td>SSB</td>
</tr>
<tr>
<td>07 May 2010</td>
<td>Kansas</td>
<td>07:43</td>
<td>39.69</td>
<td>-94.75</td>
<td>SF</td>
<td>OTR</td>
</tr>
</tbody>
</table>
Table 3.2: Number of merger events categorized by convective organization (SSB, hybrid, EMB and other) and background environment. The total number of merger events is larger than the total number of cases owing to some cases producing multiple merger events.

<table>
<thead>
<tr>
<th>observed evolution</th>
<th>weakly forced</th>
<th>strongly forced</th>
</tr>
</thead>
<tbody>
<tr>
<td>system-scale bowing</td>
<td>7</td>
<td>0</td>
</tr>
<tr>
<td>hybrid</td>
<td>2</td>
<td>9</td>
</tr>
<tr>
<td>embedded bowing</td>
<td>0</td>
<td>4</td>
</tr>
<tr>
<td>other</td>
<td>2</td>
<td>5</td>
</tr>
<tr>
<td>total merger events</td>
<td>11</td>
<td>18</td>
</tr>
<tr>
<td>total cases</td>
<td>10</td>
<td>11</td>
</tr>
</tbody>
</table>
Chapter 4

Idealized numerical simulations of a squall line-supercell merger

1. Introduction

As outlined in the previous chapter, past research on squall line-supercell mergers has focused primarily on documenting observed behaviors, particularly on the apparent relationship between the merger and tornadogenesis. Little is known, however, about the storm-scale dynamics at work in these types of events. Perhaps the most detailed storm-scale analysis comes from the observational study of Goodman and Knupp (1993), who utilized mesonet data and visual observations to investigate the evolution of surface features associated with a squall line-supercell merger. They observed that the squall line’s gust front was “distorted” in the vicinity of the merger with the supercell. They hypothesized that the distortion of the gust front may have been the result of its eastward advance being locally “blocked” by the surface meso-high associated with the supercell’s outflow. The possibility of the supercell in some way distorting the squall line’s gust front is of interest as a slowing or weakening of the squall line in
the vicinity of the merger was a common feature in the observations presented in Chapter 3. Given the importance of lifting along the gust front to squall line maintenance and organization (e.g. Rotunno et al. 1988), determining how it evolves during the course of the merger is an important element in understanding these events.

Another area of interest is determining the details of the evolution of the supercell’s mesocyclone during the course of the merger. While our observations (Chapter 3) suggest that the details of this evolution vary with merger environment, we did observe an overall trend of the strongest rotation descending with time following the merger. This appears to coincide with a maximum in tornado reports near the merger time, particularly in the “weakly forced” cases. These observations are in line with a number of past studies that observed strong low-level rotation and/or tornado production coincident with squall line-supercell mergers (e.g. Sabones et al. 1996; Wolf et al. 1996; Wolf 1998; Sieveking and Przybylinski 2004). Determining the process that facilitates this transition is important, and may shed some light on the apparent connection between mergers and tornadogenesis. One possible hypothesis is that, as the merger occurs, low-level rotation is enhanced through the tilting of horizontal vorticity present in the squall line’s cold pool.

In the most general sense we are interested in isolating which characteristics of the post-merger evolution are directly attributable to the merger, and which are just part of the squall line’s maturation. In other words, which of the observed behaviors would have occurred absent a merger? While we can obviously speculate on the role of the merger in our observed cases, we cannot say for certain that a given feature was a result of the merger, merely that it occurred following the merger. Of particular interest to this end is determining the relative importance of the merger in subsequent bow echo development, seeing as the background environments in most of the observed cases were favorable for bow echoes. In addition, we also seek to understand whether the merger location has a higher likelihood for severe weather production.
than other parts of the squall line. Our observations showed that the location of the merger frequently experienced severe weather; however, it was unclear if it was a particularly favored location relative to other portions of the squall line.

To address these issues we have run a series of idealized model simulations that capture a realistic squall line-supercell merger. To our knowledge, these simulations represent the first time that this behavior has been investigated using a numerical model, and thus may provide significant insight into some of the storm-scale features associated with these events. Section 2 details the experimental set-up for our simulations, including a novel means of introducing two distinct modes of convection into an idealized cloud model simulation with a homogeneous base-state. This is followed in Section 3 by an overview of the basic simulations, and a comparison between simulations with and without a merger. In Section 4 we provide a detailed analysis of the cold pool and low-level vertical vorticity evolutions associated with our simulated squall line-supercell merger. Finally, in Section 6, we conclude by summarizing our results in relation to the observations presented in Chapter 3 as well as past works, and provide some avenues for future work.

2. Idealized simulation set-up

This work utilized 3D idealized numerical model simulations using version 1.15 of the Bryan cloud model (CM1) described by Bryan and Fritsch (2002). This is a newer version of the same model used in the simulations discussed in Chapter 2. We used a horizontal grid spacing of 500 m as a compromise in order to sufficiently resolve convective-scale processes while also keeping computing costs manageable given the 300 x 400 x 20 km grid necessary to simulate a squall line, supercell and merged system. The vertical grid spacing was stretched from 100 m at the surface to 250 m above z=2500 m. We employed open x- and y- lateral
boundary conditions, free-slip upper and lower boundary conditions, and a Rayleigh damping layer above 14 km. In the interest of keeping these simulations as simple as possible radiative effects, surface friction and surface fluxes were all neglected. The simulations do include Coriolis forcing, applied to perturbation values only at a constant value of $f = 1 \times 10^{-4} \text{s}^{-1}$ across the entire domain (i.e. an $f$-plane). This was included because initial tests revealed that it was necessary in order to produce the asymmetric structures (i.e. a dominant cyclonic line-end vortex at the north end of the squall line) similar to what was observed for real-world mergers. This is not surprising as the convergence of planetary vorticity has been shown by a number of studies to be important to the development of cyclonic mesoscale vortices over a wide range of scales (i.e. Skamarock et al. 1994; Weisman 1993; Atkins and St. Laurent 2009b). The present simulations used a horizontally homogeneous background environment (Fig. 4.1a), based on the idealized environment of Weisman and Klemp (1982), which has been widely used in the simulation of convective storms. The squall line was triggered using a 200 km long ($y$-dimension), 10 km wide, 3 km deep line thermal with a potential temperature perturbation of +2 K. Random noise of $+/-0.1 \text{K}$ was added to the thermal to help develop 3 dimensional structures along the line. A supercell was triggered 3 hours into the simulation using a single warm bubble positioned approximately 60 km ahead of the developing squall line.

Experience revealed that the main challenge in simulating a squall line-supercell merger lies with producing both convective modes simultaneously within a single simulation: in many cases a simulation that produced a reasonable supercell storm would not produce an effective squall line, and vice-versa. This stems from the long-understood concept that convective organization is strongly tied to the background environment, particularly the wind profile. Indeed, past idealized modeling studies have demonstrated that squall lines are favored in environments with strong unidirectional wind shear, generally isolated within the lowest 2-3 km AGL (e.g.
Rotunno et al. 1988; Weisman et al. 1988; Weisman 1993; Weisman and Rotunno 2004; Bryan et al. 2006) while supercells are favored with strong deep-layer shear, particularly with a curved hodograph (e.g. Rotunno and Klemp 1982; Weisman and Klemp 1982; Weisman and Klemp 1984; Rotunno and Klemp 1985). In nature, the presence of environmental heterogeneity and strong linear forcing are often important to producing multiple modes in a localized region (e.g. French and Parker 2008). However, trying to include such heterogeneity in our idealized model would limit our ability to run controlled tests focused on the role that the storm merger is playing in convective evolution.

To address this issue, we initiated a squall line in an environment characterized by a favorable, unidirectional wind profile (Fig. 4.1a) and let it mature for three hours, essentially the time it took for the line to become steady. At this point, we wrote a restart file containing all of the run-time model fields, and modified the base-state wind profile within the restart file to resemble one more characteristic of a supercell environment, namely moderate (25 m s$^{-1}$) deep-layer shear and a low-level shear vector that veers with height (Fig. 4.1b). This was done by separating the original base-state wind profile from the perturbations that had developed in the course of running the 3-hour squall line simulation, introducing the new base-state wind profile, and then adding the original storm-induced perturbations back on to the new wind profile. In doing this we are able create a more favorable environment for supercells, while still maintaining the physical perturbations to the wind and thermodynamic fields produced by the squall line. The base state thermodynamic profile is left untouched; the small changes in values of CAPE and CIN between Figs. 4.1a and b are perturbations solely due to the presence of the squall line. Once the modifications are complete, we restart the simulation using the modified restart file, and trigger the supercell ahead of the squall line with a warm bubble.

We ran four primary simulations using this method, which we have termed ”base-state substitution”. The first simulation consists of a squall line initiated by the line thermal and
permitted to evolve without the addition of the modified environment. This is intended as a baseline to help evaluate the impacts of adding the stronger wind shear profile on the squall line, and will be referred to hereafter as the ”BASE” simulation. The second simulation uses the same set-up as BASE, but adds the wind profile modification after 3 hours. The supercell is not triggered in this environment either, as it is intended to isolate the effects of the stronger shear environment on the squall line. It will be referred to as the ”NOMERGER” simulation. In the third simulation, the isolated supercell is initiated in the higher shear environment and allowed to mature for three hours to provide an indication of how the supercell evolves in the absence of the merger with the squall line. This simulation will be referred to as the ”SUPE” simulation. Finally, in the fourth simulation, the squall line is initiated as in BASE, and the modified wind profile and supercell are added 3 hours into the simulation in order to simulate the squall line-supercell merger. This will be referred to as the “MERGER” simulation.

The effects of changing the base state wind profile are summarized in Fig. 4.2. The top panels (a-e) illustrate the evolution of a squall line in the unchanged environment of the BASE simulation, and the bottom panels (f-j) illustrate a squall line in the NOMERGER simulation, where the modified wind profile is added 3 hours (180 min.) into the simulation (e.g. between panels g and h). The simulations are identical for the first three hours (Fig. 4.2a-b, f-g) and remain quite similar for the first 10-15 minutes following the restart time (Fig. 4.2c, h). However, by approximately 20 minutes after the new environment is introduced the squall line in the modified environment (Fig. 4.2i) starts to intensify compared to the one in the base state simulation (Fig. 4.2d). After an hour, the modified squall line (Fig. 4.2j) has begun to develop bow echo characteristics, and has accelerated eastward compared to the base state squall line (Fig. 4.2e). Thus the primary impact of adding the stronger wind profile is that the simulation produces a stronger squall line, as would be expected in a stronger-shear environment (e.g. Weisman et al. 1988; Weisman 1993). This test reassures us that the method provides
an effective transition from a squall line to supercell wind profile, with no unphysical noise or detrimental effects to the simulated storms. Furthermore, by altering the environment instantaneously we are able to do so in a controlled manner. This allows us to study the impact of the storm merger in isolation, with the confidence that any changes in storm behavior are a direct result of the merger. If we had used a more gradual means of changing the wind profile, such as some type of forcing over time, or a horizontally varying wind, it would be challenging to isolate the changes in storm evolution owing to the merger from those owing to the changing wind profile.

3. **Overview and comparison of the basic simulations**

We begin our discussion by detailing the basic storm structures simulated in the NOMERGER and MERGER simulations. This will serve as a baseline to facilitate subsequent comparisons between these runs while also demonstrating that the MERGER simulation captures the salient features of observed mergers (Chapter 3).

*a. Overview of the NOMERGER, SUPE, and MERGER simulations*

The early evolution of the NOMERGER simulation is characterized by a quasi-two-dimensional squall line as shown in Fig. 4.2f-g. Once the stronger wind profile is introduced after 3 hours of simulation time, the squall line rapidly intensifies with an unbroken region of 40+ dBZ simulated radar reflectivity extending from y = 90 km to y = 215 km (Fig. 4.3a). The line continues to intensify as the simulation progresses, beginning to exhibit classic bow echo structure in the simulated reflectivity field by t = 245 min. (Fig. 4.3b). This bowing increases with time, as other hallmarks of bow echo structure (e.g. Weisman 1993) begin to materialize including
bookend vortices and a strong elevated rear-inflow jet (Fig. 4.3d-f). Eventually the squall line develops an asymmetric structure with a dominant cyclonic vortex at its north end (Fig. 4.3f) owing to the presence of Coriolis forcing in the simulation (e.g. Davis and Weisman 1994; Skamarock et al. 1994). The development of a large bow echo after the vertical wind shear increase is consistent with past studies on the environments of bow echoes (e.g. Weisman et al. 1988; Weisman 1993), and indicates that the merger is not a necessary condition to develop a bow echo in this environment.

In the SUPE simulation, the warm bubble triggers an initial isolated cell that begins to split by \( t = 230 \text{ minutes} \) (Fig. 4.4a-b). By 250 minutes, a dominant right-moving storm emerges with a hook echo structure visible in the simulated radar reflectivity (Fig. 4.4c). Through \( t = 280 \text{ min} \) (Fig. 4.4d-f) the right-moving split continues to grow in size and begins to turn toward the right, as would be expected with a supercell thunderstorm. Furthermore, throughout the time window illustrated in Fig. 4.4 the right-moving storm is characterized by large values of cyclonic vertical vorticity (not shown) co-located with the mid-level updraft, a defining feature of a supercell (e.g. Weisman and Klemp 1982; Weisman and Klemp 1984; Moller et al. 1994). Together, the NOMERGER and SUPE simulations represent the behavior of the squall line and supercell modes when they occur in isolation in the high shear environment.

The squall line in the MERGER simulation begins in the same manner as that in the NOMERGER simulation, and the two squall lines are largely identical through approximately the first 4 hours of the simulation (c.f. Figs 4.3a-b and 4.5a-b). In a similar vein, the supercell initiated by the warm bubble at \( t = 180 \text{ minutes} \) evolves quite similarly to that in the SUPE simulation through the onset of the merger. The merger begins at approximately \( t = 265 \text{ minutes} \), as the squall line overtakes the supercell along its rear flank (Fig. 4.6c). At this point, radar

---

\( ^1 \)In order to compare the SUPE simulation to the MERGER simulation, we treat SUPE as though the simulation was started at \( t = 180 \text{ minutes} \). Thus 230 minutes is approximately 50 minutes after the supercell is triggered.
reflectivity values begin to decline to the north of the merger location (Fig. 4.6b-d), consistent with observations of squall lines weakening north of the merger location as discussed in Chapter 3. As the merger progresses, several reflectivity structures common to the observed cases begin to emerge. These include the “Y”-shaped, and “S”-shaped reflectivity patterns early in the merger (Fig. 4.6b-c and d, respectively), followed by the “reflectivity swirl” and “comma echo” configurations as the bow echo becomes predominant (Fig. 4.6e-f). As in a number of observed cases, as the merger concludes, the remnant supercell becomes the northern end of the squall line as it subsequently evolves into a bow echo (Fig. 4.6e-f). This bow echo persists through the remainder of the simulation.

Given the many qualitative similarities between the structures in the MERGER simulation, and those found in the observed cases discussed in Chapter 3, we are confident that this simulation is effectively capturing the impacts of a squall line-supercell merger. In particular, this simulation appears to be a fairly accurate representation of what occurs in cases of the Weakly Forced/System-Scale Bowing archetype. With this in mind, we will now move forward with a more in depth discussion of the processes at work as the simulated merger takes place. We will first isolate the key components of system evolution that are a direct result of the merger, and then proceed to a detailed investigation of the evolutions of the squall line cold pool and low-level vorticity field.

b. Impacts of merger on storm-scale structure

We begin our analysis with a comparison of the simulated radar reflectivity fields from the MERGER and NOMERGER simulations. One of the common features observed in many merger cases (Chapter 3) was an apparent weakening of the squall line in the vicinity of the supercell just prior to the merger. As noted above, this feature is captured by the MERGER
simulation. If we compare the two simulations during the time in which this weakening is present in the MERGER simulation, we see that no such weakening occurs in the NOMERGER case (c.f. Figs 4.3c-d and 4.5c-d). In particular, the unbroken region of strong reflectivity $> 45$ dBZ persists north of $y = 180$ km throughout the NOMERGER simulation, which is the eventual position of the northern end of the line in the MERGER simulation. This suggests, as expected, that the decline of the squall line north of the merger point happens as a direct result of the merger. As will be shown shortly, this weakening appears to be tied to the evolution of the squall line’s cold pool, and represents a key part of the merger evolution.

The frequently observed post-merger evolution into a bow echo is also captured by the MERGER simulation, as is evident in the reflectivity field in Fig. 4.5. However, this does not appear to be completely due to the merger. As shown in Fig. 4.3b-f, the NOMERGER simulation also evolves into a bow echo. This suggests that for the high shear restart environment, bow echo structures will eventually evolve from a squall line, much as was hypothesized from the environments in the observed cases. However, the merger does appear to influence the location of the strongest bowing. Specifically, the apex of the bow echo in the MERGER simulation is further south than that in the NOMERGER simulation, and also appears to have a more compact region of strong bowing (c.f. Figs. 4.5e-f and 4.3e-f).

An additional feature that was observed in many actual merger cases, particularly in the weakly forced environment, was the development of a large, line-end vortex like feature associated with the merged supercell. While in some cases this appeared to evolve from the supercell, it was not always clear if this feature was a direct result of the merger, or, as with other bow echo characteristics, a product of the basic convective evolution in the observed background environment. To evaluate the production of such structures in both of our simulations, we examined “swaths” of vertical vorticity accumulated over time at 1 km, 3 km and 6 km AGL for the two simulations. Both simulations produce enhanced regions of vertical vor-
ticity near the north end of the squall line at 3 km AGL (dashed ovals, Fig. 4.7c-d), consistent with the line-end vortices present in each simulation (e.g. wind vectors in Figs. 4.3d-f and 4.5d-f). However, the region in the MERGER simulation appears to be more extensive and contain larger values. More significantly, however is the swath of enhanced vertical vorticity at 1 km in the MERGER simulation (dashed oval Fig. 4.7a). This region corresponds to the path taken by the merged supercell\textsuperscript{2}, and no such region exists in the NOMERGER simulation (Fig. 4.7b). The enhanced vertical vorticity in the MERGER simulation is also evident more at 6 km AGL (dashed oval, Fig. 4.7e-f), suggesting stronger rotation throughout the depth of the merged supercell compared to the NOMERGER squall line. However, the signal is most pronounced at lower levels. These results suggest that the merger is indeed responsible for generating a stronger line-end circulation in the resultant squall line, especially at low altitudes. This lends credence to the hypothesis that the merger is a key part of generating the enhanced circulations observed in the actual merger cases.

The merger also had a dramatic impact on the squall line’s cold pool. As seen in a swath of the minimum surface potential temperature perturbation accumulated over the lifetime of the squall line, north of the merger location a significant weakening of the MERGER system’s cold pool occurs that is not evident in the NOMERGER simulation. (dashed ovals, Fig. 4.8a-b). This is further illustrated in a difference plot comparing the MERGER and NOMERGER simulations 30 minutes after the merger begins (Fig. 4.9). In the region where the merger occurred (between y = 150 and 190 km, Fig. 4.9) the MERGER simulation cold pool is generally 3-5 K warmer than the NOMERGER cold pool. This suggests a significant weakening of the cold pool associated with the merger, and corresponds to lower simulated reflectivity values seen north of the merger in Fig. 4.5c-d. In contrast, south of the merger point the gust

\textsuperscript{2}It is clear that following the merger, the supercell and squall line have become one system, however we will refer to the northern end of the post-merger system as the “merged supercell” as a means of delineating the area impacted by the merger from the rest of the squall line.
front has surged farther east compared to that in the NOMERGER simulation (solid black vs dashed black contours, Fig. 4.9), consistent with the more pronounced bowing produced in the MERGER simulation on (Fig. 4.5e-f). The mechanisms responsible for both the weakening and subsequent surge in the squall line’s cold pool will be discussed in detail in the next section.

The cold pool evolution also reflects a significant change in gust front lifting between the two simulations. In both runs, the early evolution of the squall line is characterized by a region of unbroken, quasi 2-D slabular (James et al. 2005) lifting extending along the length of the cold pool (e.g. x = 120-180 km, Fig. 4.8c-d), which remains the case throughout the remainder of the NOMERGER simulation (Fig. 4.8d). In the MERGER simulation, the onset of the merger brings a rapid decline in low-level vertical velocities at the north end of the squall line (e.g. north of y= 175, Fig. 4.8c). South of this area, following the merger an area of strong vertical velocities persists associated with the merged supercell at the new north end of the squall line (dashed oval, Fig. 4.8c). These stronger vertical velocities are evident throughout the depth of the storm, with a region of comparatively stronger vertical motion persisting in the MERGER simulation at 5 km AGL as well (dashed oval, Fig. 4.8e). This suggests that the merger produced a lasting enhancement to the squall line with stronger vertical motions throughout the depth of the merged supercell. This is consistent with the results of several past studies that have investigated storm mergers in the general sense and found that the merged systems generally produce stronger updrafts (e.g. Tao and Simpson 1989; Kogan and Shapiro 1996; Finley et al. 2001). In the present simulations these appear to result from a super-positioning of the lifting along the squall line’s gust front and the dynamic lifting associated with the supercell (not shown).

As a final point of comparison, in Chapter 3 we speculated that the merger may play a role in producing an enhanced severe weather threat along the merged portion of the squall line.
In particular, tornado and severe wind reports increased during, and following the merger in most of the cases identified. To evaluate the role of the merger in altering the sensible weather along the squall line, we examined swaths of several fields calculated at the lowest model level (Fig. 4.10). Consistent with the observations presented in Chapter 3, we found that the merged supercell appeared to be a favored location for enhanced straight-line winds (dashed oval, Fig. 4.10a) and near-surface rotation (dashed oval, Fig. 4.10c). The maxima seen in both these fields in the MERGER simulation were absent in the NOMERGER simulation (e.g Fig. 4.10b,d) suggesting that the merger does indeed increase these indicators of potentially severe weather. Additionally, a strong signal of enhanced rain-rate was also evident along the path of the merged supercell (dashed oval, Fig. 4.10e), which was absent in the NOMERGER simulation (Fig. 4.10f). While this is not necessarily surprising given the well known propensity for storm mergers to produce enhanced rainfall (e.g. Simpson and Woodley 1971; Tao and Simpson 1984; Westcott 1984; Finley et al. 2001), it is notable that the enhanced rain rate persists along the path of the merged supercell through the end of the simulation. This indicates that rather than inducing a short-lived burst of intense rainfall, the merger appears to result in a persistently stronger northern end of the line. This is in line with the enhanced $\psi$ in this region follow the merger, as stronger updrafts will ultimately lead to greater precipitation production. The implications of this enhanced rainfall to the evolution of the system’s cold pool will be discussed further in the next section.

To summarize, as hypothesized from the observations presented in Chapter 3, the merger appears to play a significant role in altering the squall line. While bow echo structures appear to be favored in the background environment, even in the absence of a merger, the merged supercell still appears to promote stronger, more compact bowing. Furthermore, the merger to alters both the cold pool and low-level vorticity field, while also impacting the sensible weather being produced by the squall line. We will now examine in detail these merger-induced changes
4. Cold pool and low-level vorticity evolution

a. Cold pool evolution

PRE-MERGER COLD POOL WEAKENING

Given that the cold pool is central to squall line dynamics, understanding how it evolves during the course of a merger event is key to determining why the squall line behaves as it does. As mentioned previously in Chapter 3, one might intuitively think that the large cold pool associated with the squall line would effectively overwhelm the supercell, cutting off its inflow and causing it to weaken following the merger. However, based on the observations of a squall line’s gust front being “distorted” during a merger by Goodman and Knupp (1993), not to mention the repeated observations of apparent sustained supercell structures post-merger detailed in Chapter 3, this does not appear to be the case. To understand why, we will now look in detail at the evolution of the squall line’s cold pool in our MERGER simulation.

As noted in the previous section, the cold pool in the MERGER simulation appears to weaken considerably following the merger compared to the NOMERGER run. This weakening appears to be the direct result of an interaction between the system’s cold pool and the pre-line supercell outflow\(^3\). As the supercell develops ahead of the squall line, it begins to produce cold outflow along its rear flank (Fig. 4.11a). The squall line’s cold pool eventually encounters this spreading supercell outflow (black arrow, Fig. 4.11b), which initiates a weakening trend in the cold pool.

\(^3\)The outflow from the supercell obviously constitutes a cold pool of its own. However, for the purposes of this discussion and the sake of clarity, “cold pool” will refer to the squall line’s cold pool, and later that of the merged system, while “supercell outflow” will refer to the cold air associated with the supercell while ahead of the squall line.
squall line’s gust front. As the comparatively cooler air associated with the supercell’s outflow encounters the squall line’s gust front, the cross-gust front potential temperature gradient begins to weaken (e.g. black arrows, Fig. 4.11a-c). Over time, this gradient weakens further, over a larger region (Fig. 4.11d), eventually becoming non-existent by t = 315 min. (dashed oval in Fig. 4.11e). At this point, the gust front associated with the supercell outflow effectively becomes the new leading edge of the system’s cold pool in the vicinity of the merger.

The weakening of the gust front temperature gradient impacts the squall line in two ways. First, a weaker temperature difference across the gust front means that the density difference driving gust front motion is weaker, resulting in a slower forward motion of the gust front. This is evident in Fig. 4.11b-d, as the portion of the squall line’s gust front interacting with the supercell outflow begins to lag behind the segment that is farther south. This diminished temperature gradient is also detrimental to low-level lifting associated with the gust front. The cross-gust front temperature gradient is responsible for creating a pressure perturbation that drives upward motion at the leading edge of the gust front, forcing inflowing air to ascend as it reaches the gust front interface. As this temperature gradient weakens, so too does the cold pool’s ability to force vertical motion at its leading edge (e.g. Fig. 4.11f-j). This is evident as a decline (black arrows, Fig. 4.11f-i) and eventual dissipation (dashed oval, Fig. 4.11j) of the otherwise continuous slabular ascent initially associated with the squall line north of the merger.

Along with the weakening gust front lifting, the supercell’s outflow also weakens the squall line by locally cutting off its supply of high-$\theta_e$ inflow in the vicinity of the merger. Early in the simulation the squall line is being sustained by 2 km deep layer of high-$\theta_e$ inflow along the length of its gust front (Fig. 4.12a, d). As the spreading outflow associated with the supercell encounters the squall line, it begins to limit the amount of this inflow that reaches the squall line (Fig. 4.12b). Initially, the supercell’s outflow is comparatively shallow, and an elevated region
of high-$\theta_e$ parcels continues to fuel the squall line (Fig. 4.12e). However, as the supercell’s outflow deepens over time, this elevated layer is eroded, and the parcels being lifted by the squall line have comparatively lower values of $\theta_e$ (Fig. 4.12f). These lower-$\theta_e$ parcels have less CAPE, and contribute to the overall decline in convection along the squall line in the vicinity of the merger.

The combined effects of a cessation of high-$\theta_e$ inflow and diminished low-level lifting result in the squall line weakening north of the merger, and the merged supercell becoming the new northern end of the squall line. At the same time, the presence of the supercell’s outflow appears to be the key to the supercell being sustained through the merger (instead of being overwhelmed by the squall line’s cold pool). As this outflow from the supercell interacts with the squall line and weakens the cross-gust front temperature gradient, the squall line’s gust front effectively stalls, and the gust front associated with the supercell becomes the new leading edge of the merged system. As a result, the supercell’s updraft and mesocyclone continue to ingest environmental air, sustaining these structures during the early stages of the merger. These behaviors were frequently observed in the cases discussed in Chapter 3. In the system-scale bowing and hybrid evolutions, the eventual weakening of the northern end of the squall line was a hallmark of the evolution, while temporary “breaks” in the squall line were observed with the embedded bowing evolutions. In most of these cases, structures associated with the supercell could be tracked throughout the merger, indicating that it was being sustained. Based on the results presented in this section, it would appear that the weakening of the squall line’s cold pool by the supercell’s outflow is directly responsible for both the weakening of the squall line, and the sustenance of the supercell.
POST-MERGER COLD POOL SURGE

As illustrated in Fig. 4.9, after the merger the gust front in the MERGER simulation appears to accelerate eastward compared to that in the NOMERGER simulation. This behavior is reminiscent of the evolution seen in many of our observed cases wherein the squall line would appear to surge towards the east following the merger as the bow echo developed. Seeing as this process may be an important part of the post-merger bow echo evolution we will now examine this post-merger surge in more detail.

One hypothesis as to the driving mechanism behind this surge is that the system’s cold pool strengthens following the merger. A stronger cold pool would produce a stronger pressure gradient across the squall line’s gust front, facilitating a faster motion, and thus producing the surge. A qualitative comparison between the cold pools in the MERGER and NOMERGER simulations appears to reveal a stronger cold pool in terms of a larger, more continuous area of colder air (e.g. $\theta' < -10$ K with lower minimum values (i.e. areas of $\theta' < -12$ K) during the 30 minutes following the merger (c.f. Fig. 4.13a-c and d-f). The difference in cold pool intensity is more evident in a time series of the maximum cold pool strength, $C$ (Fig. 4.14), calculated as in Chapter 2 for the segment of the squall line that evolves into a bow echo post-merger (between $y = 100$ and 170 km). It is clear that the MERGER simulation produces a stronger cold pool for approximately an hour following the merger, which likely accounts for at least part of the observed gust front acceleration. A likely source for this enhanced cold pool is the increased rainfall that occurs following the merger, shown in Fig. 4.10. As rain falls into the sub-saturated air below the cloud base, some of the rain evaporates, causing cooling. This evaporatively chilled air is a key contributor to the production of surface outflow, thus heavier rainfall will generate stronger outflow through enhanced evaporative cooling.

As shown in Fig. 4.13b-c, the surging gust front in the MERGER simulation (e.g. north of
y = 150 km in Fig. 4.13) is co-located with a region of strong (> 35 m s\(^{-1}\)) surface winds that are not present in the NOMERGER simulation (Fig. 4.13e-f). These strong winds may also contribute to the gust front surge as the gust front is locally accelerated eastward by these strong surface winds. Recent work by Mahoney et al. (2009) has demonstrated that such a scenario can result when strong winds from aloft are transported to the surface in convective downdrafts. Ostensibly the strong winds accelerate the gust front motion by adding an advective component to the gust front propagation governed by density current dynamics. Furthermore, Mahoney and Lackmann (2011) also identified this mechanism as an avenue for the development of severe surface winds, similar to what was seen in the present simulations (e.g. Fig. 4.10a).

There is evidence that the strong surface winds in the present simulations are the result of downward momentum transport as well, as they emanate from regions of very strong downward momentum flux in the vicinity of the merged supercell (Fig. 4.15a-c). Vertical cross sections through the region of strong winds indicate that they descend from a belt of strong flow centered approximately 2 km AGL (Fig. 4.16a-c). This is consistent with strong winds associated with the rear-inflow jet aloft (e.g. Fig. 4.5c-f) descending to the surface. The apparent driving force for this descent is strong downward motion associated with several regions of enhanced precipitation that develop following the merger (Fig. 4.17a-c). This enhanced precipitation is roughly co-located with the merged supercell, and likely results from the maximum in vertical velocity that is produced by the merger, shown in Fig. 4.8c. As this heavy precipitation descends, it drives the strong downdrafts illustrated in Fig. 4.17a-c, transporting the high-momentum rear-inflow-jet air to the surface in the process. Once there, these strong winds accelerate the gust front eastward as discussed by Mahoney et al. (2009), producing the gust front surge seen in Fig. 4.9.
b. Low-level vorticity evolution

Another unique feature of the MERGER simulation identified in Section 3 is a persistent region of strong low-level (0-3 km AGL) vertical vorticity associated with the merged supercell (Fig. 4.18). In this section we will focus our attention on the development and evolution of the low-level (approximately 0-1 km AGL) vertical vorticity field. The development of low-level rotation was a key component of the observed merger evolution (Chapter 3) and determining the source of this vorticity is a logical first step toward better understanding the role that squall line-supercell mergers may play in tornadogenesis.

The low-level vertical vorticity evolution in the MERGER simulation is summarized in Fig. 4.19 and the time series in Fig. 4.20a. For the purposes of this discussion, plots at 500 m AGL will be used for consistency, however similar features are evident during this period throughout the depth from 0-3 km AGL. Prior to the merger, a localized area of vertical vorticity is associated with the supercell (Fig. 4.19a), and this feature changes little prior to the beginning of the merger at t = 265 minutes (Fig. 4.19a-c). Shortly after the merger begins, vertical vorticity values rapidly intensified (Fig. 4.20a) and the area of these large, positive values grows in size (Fig. 4.19c-e). This region of enhanced low-level vorticity remains present for close to 30 minutes following the merger (e.g. Figs. 4.19c-h, and 4.20a) before beginning to decline in intensity at t = 295 minutes (Fig. 4.19i). For the remainder of the simulation, the low-level vertical vorticity field is characterized by a large region of predominantly cyclonic vertical vorticity along the north end of the squall line.

The onset of the rapid intensification of vertical vorticity as the merger occurs is strong circumstantial evidence that the merger is a primary cause for this increase. Furthermore, as shown by the SUPE simulation, in the absence of the merger, the simulated supercell does not develop such a large area of strong low-level vertical vorticity during this same time window.
As will be shown shortly, it appears that a combination of convergence of pre-existing vertical vorticity along the squall line and the production of new vertical vorticity through the tilting of horizontal vorticity in the squall line’s cold pool are central to the rapid growth and intensification of the low-level vertical vorticity. Past research has shown that both of these processes can be important to the development of mesoscale vortices. Cram et al. (2002) found that convergence of vertical vorticity was central to the development of a line-end vortex in a simulated squall line and Finley et al. (2001) obtained a similar result for the intensification of the mesocyclone in a simulated HP supercell. Tilting of baroclinically-generated vorticity has also been shown in past studies to be important in the generation of a wide variety of different mesoscale vortices including squall line meso-vortices (e.g. Trapp and Weisman 2003; Atkins and St. Laurent 2009b), line-end vortices associated with bow echoes (e.g. Weisman 1993; Davis and Weisman 1994) and low-level mesocyclones in supercells (e.g. Rotunno and Klemp 1985).

To more fully understand the development of low-level rotation in the merger simulation, a vorticity budget was calculated using the flux form of the vorticity equation:

$$\frac{\partial}{\partial t} (\rho_0 \zeta) = - \frac{\partial}{\partial t} (\rho_0 u \zeta) - \frac{\partial}{\partial t} (\rho_0 v \zeta) - \frac{\partial}{\partial z} \zeta \frac{\partial}{\partial z} (\rho_0 w) - \rho_0 \left[ \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} \right].$$

In this equation term A represents the rate of change of vertical vorticity, terms B and C represent the horizontal flux divergence of vorticity, and terms D and E represent the stretching of vertical vorticity and the generation of vorticity through tilting, respectively. This simplified budget neglects changes to vertical vorticity due to vertical advection. These forcing terms were integrated over a 12 km x 12 km area that follows the low-level vorticity maximum initially associated with the isolated supercell, and eventually with the merged system. As shown in the time-series in Fig. 4.22a the vertical vorticity within the selected region increases rapidly.

(Fig. 4.21a-f).
beginning just before the merger time, and continues to increase for nearly 30 minutes following the merger. During this period of most rapid intensification (approximately \( t = 260-275 \) minutes, Fig. 4.22a) there are positive contributions from the tilting, stretching, and east-west flux divergence terms, with the tilting term being most significant (Fig. 4.22b). Following the merger, the stretching and tilting term each dominate at different points in the evolution. We will now evaluate each of these forcing terms in turn.

The flux divergence term provides a positive contribution to vorticity growth throughout most of the time period before and after the merger (Fig. 4.22b). Its largest contribution comes right around merger time (between \( t=260 \) and \( t= 270 \), Fig. 4.22b), likely corresponding to the merging of the vorticity maximum associated with the supercell with several maxima present along the squall line. This accounts not only for part of the increase in magnitude of vorticity, but the growth in the size of the vortex during this period as well. Qualitatively, this is evident in Fig. 4.19a-c. At \( t = 255 \) minutes (Fig. 4.19a) there was a concentrated region of vertical vorticity associated with the supercell, and several pockets of enhanced cyclonic vertical vorticity along the squall line’s gust front. As the merger proceeds Fig. 4.19b-d) these maxima appear to coalesce into a larger single region of vertical vorticity by \( t = 270 \) minutes (Fig. 4.19d). Such behavior has been shown to produce stronger, and larger vortices in a number of past studies (e.g. Marquis et al. 2007; Finley et al. 2001; Cram et al. 2002).

Stretching also contributes positively throughout the time period of interest (Fig. 4.22b), and dominates the vorticity budget for approximately 20 minutes following the merger during a period of maximized low-level vertical velocity (e.g. \( t = 275- 300 \) min., Fig. 4.20b). This increased vertical velocity appears to result from a superposition of the updrafts initially associated with the supercell and squall line updraft. Prior to the merger, two comparatively weak regions of vertical velocity were present associated with the supercell and the squall line (Fig. 4.23a). As the merger begins these two updrafts merge (Fig. 4.23b) leading to a con-
siderably stronger low-level updraft immediately following the merger. This strong-updraft is co-located with a broad region of vertical vorticity, creating conditions favorable for vorticity growth through tilting given the low-level vertical velocity gradient (e.g. Fig. 4.23c). As the system continues to evolve, the gust front and its associated updraft continue eastward away from the region of strongest vertical vorticity (Fig. 4.23d), which coincides with a decline in vorticity growth via stretching (Fig. 4.22b). An overall decline in vertical motion during this time (Fig. 4.20b) likely also contributes to the diminished stretching.

The evolution of the tilting term during appears rather complex during the course of the merger. Initially, positive tilting (t = 260 - 274 min., Fig. 4.22b) occurs as the low-level updraft associated with the supercell begins to interact with a region of strong horizontal vorticity along the squall line’s gust front (Fig. 4.24a). At this point, the horizontal vorticity appears to be largely streamwise, as it generates a large area of primarily positive tilting (Fig. 4.24b), leading to the overall positive contribution. Following the merger, however, as the low-level updraft penetrates deeper in the pool of horizontal vorticity (Fig. 4.24c) tilting continues, however now the vorticity appears more crosswise in nature, as the updraft produces couplets of positive and negative tilting (Fig. 4.24c). During this period (t= 275 - 285 min. in Fig. 4.22b) any positive tilting is offset by equally strong negative tilting, and the overall contribution is negative. Finally, approximately 20 minutes following the merger, the contribution from tilting becomes positive again (t= 285 - 300 min. in Fig. 4.22b), as a large region of positive tilting suggests that the horizontal vorticity has once again become streamwise (Fig. 4.24f). It is unclear whether the transitions from streamwise to cross-wise vorticity and back are the result of something systematic, or merely a reflection of the noisy, three-dimensional nature of the horizontal vorticity field following the merger.

During the course of this evolution, the circulation associated with this area of enhanced vorticity grows in size as well, as evident in the wind vectors in Fig. 4.25a-d. Initially, the
circulation evident in the wind field is approximately 5 km in diameter (Fig. 4.25a), focused on the small region of vertical vorticity that was initially associated with the supercell (approximately \( x = 225 \) km, \( y = 170 \) km). As the merger ensues, this circulation grows in size (Fig. 4.25b-c), reaching a diameter that encompasses most of the northern end of the squall line by \( t = 320 \) min. (Fig. 4.25d). This is consistent with the circulation having evolved into a line-end vortex at the north end of the bow echo, as discussed by Weisman (1993), and is the natural result of a finite line of convection in an environment with Coriolis forcing due to the convergence of planetary vorticity (e.g. Davis and Weisman 1994; Skamarock et al. 1994). In a sense, the merger is responsible for the development of this feature as well, seeing as the merger caused the weakening of the original squall line resulting in a new line-end farther south.

To summarize, the low-level vorticity evolution follows a complex path from low-level supercell mesocyclone to bow echo line end vortex. As the merger begins, smaller vortices present along the squall line merge with the low-level mesocyclone associated with the supercell to produce a larger region of strong vertical vorticity. At the same time, an enhanced region of vertical velocity, oriented orthogonal to the low-level horizontal vorticity field further enhances low-level vertical vorticity via tilting. As the updrafts initially associated with the supercell and squall line gust front merge, a region of strong low-level vertical velocities develop leading to continued vorticity growth via stretching. Later, as this low-level updraft weakens, the contribution owing to stretching decreases, however strong vertical vorticity is maintained through positive tilting. Finally, as the merged system evolves into a bow echo, the low-level vorticity field is dominated by a cyclonic line-end vortex associated with the bow echo. Overall, this evolution appears consistent with observations of storm rotation presented in Chapter 3. Specifically, the development of strong low-level rotation was a common feature in just about every case, and the circulations in many of the weakly forced environment cases
also appeared to develop broader low-level circulations post-merger, similar to the line-end vortex in the present simulation. This gives us confidence that the evolution presented in this section is a reasonable explanation of the processes at work in our observed cases.

5. Sensitivity tests

Throughout the course of performing the observational analysis presented in Chapter 3, we noted nuances between otherwise similar cases that we hypothesized may be the result of variations in storm scale structures such as the location of the merger along the squall line. This may be particularly useful in delineating between the hybrid and embedded bowing archetypes, as both of these occurred in similar environments with a major difference being that the embedded evolution results more often from mergers near the center of the squall line. In this section we will address the role of merger location, and also touch on the sensitivity to storm maturity and the application of the base-state substitution method. This is intended as a first step in beginning to assess the sensitivity of these types of merger events to storm-scale details, and serves to motivate a more complete testing of these and additional parameter spaces (such as sensitivity to the background wind profile) in the future.

a. Sensitivity to merger location

To investigate the role that merger location plays, three additional simulations were run wherein the bubble used to initiate the supercell was moved progressively further south as shown in Fig. 4.26a-d. This resulted in bubbles placed at y=150, y=130, and y=90 km which will be referred to as the Y150, Y130, and Y90 runs respectively (for reference in the MERGER simulation the bubble was placed at y=170 km).
As the location of the merger moves south, it has a progressively smaller impact on the squall line in terms of weakening north of the merger and bow echo development south of the merger, two of the main characteristics identified in the MERGER simulation. In the Y150 simulation, the post-merger bow echo appears fairly similar to that in the MERGER simulation (c.f. Fig. 4.27a and b), just shifted further south. There is also a general weakening trend north of the merger, although it is not nearly as pronounced as in the MERGER simulation, with a large area of > 40 dBZ simulated radar reflectivity remaining north of y =150 km. The Y130 simulation, produces a more dramatic departure from the MERGER simulation, with weaker bowing, and the squall line being largely sustained north of the merger location (Fig. 4.27c). Finally, in the Y90 simulation, there is no clear bowing associated with the merger, and the squall line is largely unchanged north of the merger (Fig. 4.27d).

The apparent maintenance of the squall line north of the merger location in the Y150, Y130 and Y90 simulations appears to result from the squall line’s cold pool remaining strong throughout the merger. Recalling the discussion from the previous section, in the MERGER simulation, the weakening of the squall line north of the merger appears to be a direct result of a diminished cold pool intensity (e.g. Fig. 4.28a). As the merger is moved south, however, this weakening is not as pronounced in the Y150 simulation (Fig. 4.28b) and is largely absent in the Y130 and Y90 simulations (Fig. 4.28c, d). This is likely because as the merger is shifted south, it occurs closer to the strongest (i.e. coldest, deepest) part of the squall line’s cold pool. As a result, the squall line’s cold pool is less apt to be influenced by the weaker outflow associated with the supercell. This suggests that the weakening trend observed at the north end of the squall line in the MERGER simulation is at least partially a function of the merger occurring close to the north end of the line.

The location of the merger also has an influence on the production of strong surface winds and low-level vertical vorticity in the merged system. Once again, the Y150 simulation behaves
similarly to the MERGER simulation, with the merged system producing a large region of strong surface winds and low-level vertical vorticity (c.f. Fig. 4.29a, c and b, d). However, the as the merger is shifted south both of these features decline compared to the MERGER simulation with the Y130 and Y90 simulations producing isolated pockets of severe surface winds (Figs. 4.30b and 4.31b) and short-lived low-level vorticity tracks (Figs. 4.30d and 4.31d). The changes in vorticity evolution are also evident in the time series shown in Fig. 4.32a-d. The Y150 simulation produces a peak in vertical vorticity following the merger that is much longer lived than that in the MERGER simulation (c.f. Fig. 4.32a-b), albeit slightly weaker in magnitude, consistent with the longer low-level vorticity track in Fig. 4.29d. Further south both the Y130 and Y90 simulation also produce increases in vertical vorticity following the merger (Fig. 4.32c, d), however these are comparatively weak and/or short-lived. In fact, in the Y90 simulation the strongest vertical vorticity occurs with the isolated supercell prior to the merger (t = -30 - -10 min., Fig. 4.32d), perhaps owing to a later merger time in this simulation.

The apparent north-south decline in vertical vorticity seen in these simulations is not all that surprising, given the larger-scale vorticity structure associated with squall lines. As shown by Weisman (1993) mature bow echoes are characterized by counter-rotating bookend vortices, with a cyclonic vortex at the north end of the line and an anti-cyclone vortex at the south end. In environments with Coriolis forcing, such as the present simulations, the northern cyclonic will dominate due to the convergence of planetary vorticity (e.g. Weisman 1993; Davis and Weisman 1994; Skamarock et al. 1994). In the present simulations, the mergers that occur toward the north end of the line (e.g. MERGER, Y150) appear to benefit from this enhanced cyclonic vorticity, resulting in the large values of low-level vertical vorticity post-merger (Fig. 4.32a, b). Meanwhile, the diminished vertical vorticity observed with the mergers further south (e.g. Y130, Y90, Fig. 4.32c, d), would appear consistent with an environment characterized by weakly anti-cyclonic vertical vorticity, as is found at the south end of a mature squall line.
Thus, the location of the merger along the squall line appears to play an important role in the ultimate squall line evolution. For mergers north of the center point of the squall line (e.g. MERGER, Y150) more pronounced bowing occurs and the squall line weakens more considerably north of the merger. These locations also appear to favor the development of strong low-level vertical vorticity and severe surface winds. For mergers that occur near or south of the center-point of the squall line (e.g. Y130, Y90), the impact of the merger is much more localized and the squall line is better maintained north of the line. Additionally, these situations are characterized by weaker, shorter lived increases in low-level vertical vorticity, and more localized damaging surface winds. All told, these results suggest that merger location may be an important delineator between the different archetypes identified in Chapter 3. In particular, the embedded bowing structure was typically observed following a merger near the center of a line, and often produced more of a localized impact on the squall line, as seen here in the Y130 and Y90 simulations. This indicates that merger location needs to taken into account along with the background environment for future studies trying to determine the factors that delineate between the system-scale, hybrid and embedded bowing archetypes.

b. Additional sensitivity tests

In addition to evaluating the role of merger location, the sensitivity to storm maturity at merger time, and the impacts of reversing the base-state substitution (i.e. initiating the supercell in a supercell environment and then replacing it with the squall line environment and triggering the squall line) were also evaluated. However, in both of these cases the supercells weakened prior to the merger, thus the results will just be touched upon briefly.

To test the impacts of storm maturity, the location of the isolated bubble to initiate the supercell was shifted east by 40 km (to x = 220, hereinafter the "X220" simulation). This
resulted in a delay of the merger by approximately 1 hour, with the intent being that both systems would be more mature prior to merging. However, by approximately \( t = 300 \) minutes the supercell began to weaken and became more multicellular in organization (Fig. 4.33a-c), coincident with a decline in low-level vertical vorticity (Fig. 4.34a-c). It is not entirely clear what causes this weakening, although it appears to be preceded by the development of several new cells between the squall line and supercell (Fig. 4.33a). These cells appear to merge with the supercell (Fig. 4.33b-c), which appears to produce a surge in outflow (not shown). This increased outflow may account for the shift to more of a multicellular behavior. The merger between this multicellular structure and squall line still appears to facilitate enhanced bow echo structures (Fig. 4.33d-f), and some localized meso-vortex structures south of the merger (Fig. 4.34d-f), however the strong, large area of strong low-level vertical vorticity observed in the MERGER simulation does not materialize.

The supercell meets a similar fate in a simulation where the base-state substitution method is reversed. In this test, the simulation is started with the supercell wind profile, and a warm bubble so that an isolated supercell is the initial mode. The supercell is then allowed to mature for 2 hours, at which point the wind profile is replaced with the weaker-shear squall line wind profile from the original BASE simulation, and the squall line is triggered with the warm line thermal. The introduction of the weaker shear profile leads to a very rapid weakening of the supercell (e.g. Fig. 4.35a-b), and a squall line structure forms along the supercell’s outflow boundary (e.g. Fig. 4.35b-d). This squall line appears to dominate the simulation (Fig. 4.35b-d) with the remnants of the squall line triggered by the line thermal eventually merging with its southern end (Fig. 4.35d-f). While a low-level meso-vortex appears to develop where these two lines intersect (not shown), the simulation does not capture the desired squall line-supercell merger, and there is no large area of cyclonic vorticity associated with this merger. This does, however, underscore the apparent importance of having a favorable supercell environment to
sustain the supercell during the course of a merger event.

While both of these sensitivity simulations failed to capture the desired merger behavior, the concepts that motivated them (storm maturity, and the effects of a weaker shear environment on the merged system) remain of interest. We plan to revisit these topics in future studies as we evaluate a wider parameter space including different background environments and microphysics schemes. In particular, it could be informative to run the “reverse base-state substitution” configuration using a less dramatic change in wind profiles, such as moving from a strong-shear supercell hodograph to a weaker shear supercell hodograph. This may provide some useful additional insight as to the role that the background environment plays in the post-merger evolution.

6. Conclusions and future work

A series of idealized model simulations capture many of the salient features observed in the weakly-forced, system-scale bowing squall line-supercell merger cases presented in Chapter 3. These simulations demonstrated that while the development of bow echo organization was driven by the background environment used in these simulations, the merger exerts considerable influence on the details of the system evolution. Specifically, the merger influences the nature and location of the bowing, causing a more pronounced bow located farther south in the MERGER simulation. The merger also appears to influence the sensible weather produced by the squall line, locally promoting increased rainfall, stronger surface winds, and enhanced low-level vertical vorticity. Furthermore, the merger exerts a considerable influence on the structure of the squall line’s cold pool and low-level vorticity field. As the merger begins, the squall line’s cold pool is locally weakened as it interacts with outflow from the pre-line supercell. As a result, the supercell becomes the new leading edge of the line and continues to ingest environ-
mental air. The remnant supercell persists as a zone of enhanced updraft, and thus enhanced precipitation, for some time. Post merger, intense downdrafts associated with the enhanced rainfall produce strong surface winds that cause an eastward surge in the merged system’s gust front, generating a more pronounced bow echo than was seen in the NOMERGER simulation. The merger also engenders the development of strong low-level vertical vorticity through the tilting of horizontal vorticity present in the squall line’s cold pool. The primary mechanism for this tilting appears to be a localized region of enhanced low-level (e.g. 1 km AGL) vertical velocities associated with the remnant supercell. Eventually this low-level circulation evolves into a broad line-end vortex structure as the bow echo emerges post-merger.

The simulations presented in this chapter provide some insight into the physical processes responsible for the observed behaviors documented in Chapter 3. Of particular interest is the finding that the outflow associated with the supercell effectively weakens the cold pool of the squall line prior to the merger, precluding the squall line’s outflow from “undercutting” the supercell’s updraft and removing its source of inflow. Past authors (e.g. Wolf 1998) had commented that the observation of supercells being sustained during the merger was an “un-expected” result, owing to the assumption that the squall line’s outflow would overwhelm the isolated storm. Our results provide an explanation for why this does not occur. Additionally, our results help to further explain the observation of Goodman and Knupp (1993) that the squall line’s gust front was “distorted” by the supercell as a merger took place. The authors had speculated that “the supercell’s meso high blocked the eastward advance of the squall line gust front”, which would be in line with the results of our simulations.

The production of enhanced low-level vertical vorticity coincident with the onset of the merger is consistent with the observations presented in Chapter 3, and with past studies that have observed tornadogenesis occurring around the time of a squall line-supercell merger (e.g. Goodman and Knupp 1993; Wolf et al. 1996; Sabones et al. 1996; Wolf 1998). While the
500 m horizontal grid spacing used in the present simulations is insufficient to resolve actual tornadoes, the presence of strong vertical vorticity at the lowest model level (e.g. Fig. 4.10c) at the very least suggests a condition favorable for tornado development. Interestingly, if this low-level vorticity is primarily generated through the tilting process outlined above, then it is entirely possible that any storm merger involving a squall line may produce enhanced low-level vertical vorticity, not just an interaction with a supercell. The assumption would be that most storm mergers could cause a localized gradient in $w$ along the zone of otherwise slabular ascent, promoting tilting of horizontal vorticity (which is common in squall line cold pools). This is of interest as conversations with forecasters have revealed that small scale circulations and tornadoes are sometimes observed with mergers between squall lines and ordinary cell thunderstorms, not just supercells. It stands to reason that a merger with a supercell would be more favorable for the development of low-level rotation, owing to stronger vertical motions and pre-existing low-level vertical vorticity, but the effect of non-supercell mergers may deserve additional attention in a future study.

Finally, the primary simulations presented in this chapter were run using idealized background conditions that fall between the “weakly forced” and ”strongly forced” environments detailed in Chapter 3, and generally capture the behaviors associated with the “system-scale bowing” evolution. To better understand the role that the background environment plays in governing the observed post-merger storm organizations (e.g. including the ”hybrid” and ”embedded bowing” archetypes discussed in Chapter 3) an additional study is planned to look at a wider range of background environments. This would entail using both more realistic wind profiles created from the composite RUC environments, as well as a larger parameter-space type study using idealized wind profiles to systematically test different shear profiles. These tests should also include several different merger configurations, as the results of current sensitivity tests show that merger location plays an important role in the ultimate system organization. In
addition, case study simulations of several “strongly forced” cases are also under way to better understand merger evolution within this environment.
Figure 4.1: Skew-T log-P diagram of temperature, dew point temperature and lifted parcel path, wind profiles and hodographs (kts) for (a) base-state squall line environment and (b) the supercell environment added 3 hours into the simulation. Color coding on the hodograph denotes various height layers as follows: green: 0-1 km, blue: 1-2 km, red: 2-3 km, yellow 3-6 km, and black 6km- model top. Wind barbs are in knots with a half barb = 5 knots, full barb = 10 knots, and flag = 50 knots.
Figure 4.2: Simulated radar reflectivity (dBZ), 1 km AGL winds (m s\(^{-1}\)), and -1 K surface potential temperature perturbation between t = 125 and 245 minutes for (a)-(e) a squall line simulation using the initial base-state environment and (f)-(j) a squall line simulation wherein the environment is modified after 3 hours, as discussed in the text.
FIGURE 4.3: Simulated radar reflectivity (dBZ, shaded as shown) and 1 km AGL winds (m s\(^{-1}\), representative vector at lower right of panels e and f) at (a) 200 (b) 245 (c) 265 (d) 280 (e) 300 and (f) 320 minutes into the NOMERGER simulation.
As in Fig. 4.3, but for the SUPE simulation. Note that the times labels reflect storm initiation at $t = 180$ min., to ease comparison with the MERGER simulation. Also, a smaller plotting window is used to zoom in on the supercell.
MERGER simulation
sim. radar ref. and 3 km AGL winds

Figure 4.5: As in Fig. 4.3, but for the MERGER simulation.
FIGURE 4.6: As in Fig. 4.5, but zoomed in on the merger with data every 10 minutes between $t = 255$ and $t = 310$ min.
FIGURE 4.7: Swaths of maximum vertical vorticity (s$^{-1}$, shaded as shown) at 1 km AGL (top panels), 3 km AGL (middle panels) and 6 km AGL (bottom panels) accumulated between 3 and 6 hours into the simulation for the a, c, e MERGER and b, d, f NOMERGER simulations. The fields are accumulated by taking the maximum value over time across the subset of the model domain shown. The vertical dashed line in each panel denotes the approximate X-position of the merger. Dashed ovals highlight features of interest discussed in the text.
FIGURE 4.8: As in Fig. 4.7 but for top panels: minimum surface potential temperature perturbation (K, shaded as shown). Middle panels: 1 km AGL vertical velocity (m s$^{-1}$, shaded as shown). Bottom panels: 5 km AGL vertical velocity (m s$^{-1}$, shaded as shown).
Figure 4.9: Difference plot comparing the surface potential temperature perturbation (K, shaded as shown) in the MERGER and NOMERGER simulations (MERGER-NOMERGER). The sold and dashed black contours indicate the positions of the -1 K potential temperature contours, representing the gust front location for the MERGER and NOMERGER simulations, respectively.
Figure 4.10: As in Fig. 4.7 but for maximum wind speed (top panels, m s\(^{-1}\), shaded as shown) and vertical vorticity (middle panels, s\(^{-1}\), shaded as shown) at the lowest model level, and rainfall rate (bottom panels, mm hr\(^{-1}\), shaded as shown).
Figure 4.11: Evolution of gust front lifting in the MERGER simulation. (a-e) magnitude of the surface potential temperature gradient (K km$^{-1}$, shaded as shown) and -1 K theta perturbation (dashed contour). (j-g) 1 km AGL vertical velocity (m s$^{-1}$, shaded as shown and -1 K surface theta perturbation (dashed contour). Arrows and dashed ovals denote features of interest discussed in the text.
Figure 4.12: Evolution of storm-relative inflow at between t = 230 minutes and t= 260 minutes (left-to-right) in the MERGER simulation. Top panels: plan-view of surface equivalent potential temperature (K, shaded as shown), surface potential temperature perturbation (contoured at -2 (black), -4 (dark blue), -6 (medium blue) and -8 (purple) K, and squall line-relative surface winds. Bottom panels: x-z cross sections of the same fields in (a-c) taken along the black lines shown in (a-c).
Figure 4.13: Surface potential temperature perturbation (K, shaded as shown) and surface winds greater than 35 m s$^{-1}$ (representative vector below color bar) for the (top panels) MERGER simulation and (bottom panels) NOMERGER simulation at t = 265, 275 and 295 minutes (left to right).
Figure 4.14: Time series of maximum cold pool strength, \( c (\text{m s}^{-1}) \) for the MERGER (solid black) and NOMERGER (dashed red) simulations between \( t = 190 \) and 360 minutes. The approximate merger time is denoted by the vertical dashed line. The maximum value is calculated between \( y = 120 \) and 180 km, which encompasses the region of maximum cold pool intensity in association with the post-merger bow echo in the MERGER simulation.
Figure 4.15: As in Fig. 4.13, but for downward momentum flux (shaded as shown, m² s⁻²), surface rain rate (contoured every 50 mm hr⁻¹ starting at 50), surface winds greater than 35 m s⁻¹ (representative vector below color bar), and -1 K surface potential temperature perturbation.
Figure 4.16: Vertical (x-z) cross sections of horizontal wind speed (m s$^{-1}$, shaded as shown) and $w$ (contoured at -1, -5 and -10 m s$^{-1}$) along $y = 165$ km for the (top panels) MERGER and (bottom panels) NOMERGER simulations at (left-right) $t = 270, 280, 290$ minutes..
**Figure 4.17:** As in Fig. 4.13, but showing $w$ (values < 0 shaded as shown, m s$^{-1}$), surface rain rate (contoured every 50 mm hr$^{-1}$ starting at 50) and -1 K surface potential temperature perturbation.
FIGURE 4.18: Time vs. height plot of maximum vertical vorticity (s$^{-1}$) associated with the supercell pre-merger and merged system post-merger. The vertical dashed line denotes the merger time.
Figure 4.19: 500 m AGL vertical vorticity (s$^{-1}$, shaded as shown) and 40 dBZ simulated radar reflectivity contour for the MERGER simulation at the times shown in the upper right of each panel. Black arrows in each panel identify the vorticity associated with the isolated supercell and subsequent merged system.
Figure 4.20: (a) 500 m AGL maximum vertical vorticity (s⁻¹,) and (b) 1 km AGL maximum vertical velocity for (solid black) MERGER and NOMERGER (dashed red) simulations between t = 240 and t = 325 minutes.
FIGURE 4.21: Horizontal cross-sections of vertical vorticity (s$^{-1}$, shaded as shown) and 40 dBZ radar reflectivity contour at 1 km AGL for the SUPE simulation at t = (a) 250, (b) 260, (c) 270, (d) 280, (e) 290, and (f) 300 min.
Figure 4.22: Time series of (a) vertical vorticity and (b) vertical vorticity tendency due to tilting (solid black), stretching (dashed red), east-west flux divergence (dotted green), and north-south flux divergence (alternating dashed blue) integrated over the 12 km x 12 km area described in the text.
**Figure 4.23:** Cross section of along-line averaged vertical velocity (contoured every -2.5, 2.5, 5, 10 m s\(^{-1}\)) and along-line maximum vertical vorticity (shaded as shown, s\(^{-1}\)) at t = (a) 250 (b) 265 (c) 275 (d) 295 min. Values are averaged/maxed between y = 150 and 180 km.
FIGURE 4.24: Evolution of tilting term in vorticity equation at $t = (a,b)$ 265, (c,d) 280 and (e,f) 295 min. Left panels (a,c,e): horizontal vorticity vectors and magnitude (shaded as shown, $s^{-1}$) and vertical velocity (contoured at 5, 10 m $s^{-1}$). Right panels: tilting term (shaded as shown, $s^{-2}$) and vertical velocity (contoured at 5 m $s^{-1}$).
Figure 4.25: Horizontal cross-sections of vertical vorticity (s\(^{-1}\), shaded as shown) and storm-relative wind vectors at 3 km AGL for the MERGER simulation at \(t = (a) \) 275, \(b) \) 290, \(c) \) 305 and \(d) \) 320 min.
FIGURE 4.26: Plan-view plots of simulated radar reflectivity (shaded as shown, dBZ) and integrated updraft helicity (contoured every 250 m$^2$ s$^{-2}$) at $t = 235$ min. for (a) MERGER simulation, (b) Y150 simulation, (c) Y130 simulation, (d) Y90 simulation. Black arrows identify the supercell in each panel.
Figure 4.27: As in Fig. 4.26 but at $t = 330$ min. Black arrows identify the location of the merged supercell in each panel.
Figure 4.28: As in Fig. 4.27 but showing the surface potential temperature perturbation (shaded as shown, K) and the 40 dBZ simulated radar reflectivity contour. Black arrows denote areas of comparatively weaker cold pool in the vicinity of the merger in each simulation.
FIGURE 4.29: As in Fig. 4.10 but comparing the MERGER and Y150 simulations.
FIGURE 4.30: As in Fig. 4.10 but comparing the MERGER and Y130 simulations.
FIGURE 4.31: As in Fig. 4.10 but comparing the MERGER and Y90 simulations.
**Figure 4.32:** Time-series of maximum vertical vorticity $s^{-1}$ calculated at 1 km AGL over 250 km (east-west) by 100 km (north-south) box centered on the merger for (a) MERGER simulation (b) Y150 simulation, (c) Y130 simulation, and (d) Y90 simulation (solid black line). The dashed red line in each panel represents the maximum vorticity calculated over the same region in the NOMERGER simulation.
FIGURE 4.33: As in Fig. 4.3, but for the X220 simulation at $t =$ (a) 290 (b) 305 (c) 315 (d) 330 (e) 345 and (f) 360 min.
Figure 4.34: As in Fig. 4.33, but showing 500 m AGL vertical vorticity (shaded as shown, \( \text{s}^{-1} \)) and the 40 dBZ simulated reflectivity contour. Black arrows denote the isolated and eventual merged supercell and red arrows identify squall line meso-vortices.
Figure 4.35: Simulated radar reflectivity (dBZ, shaded as shown) and -1 K surface potential temperature perturbation for the reverse base-state substitution simulation at $t = (a) 190$ (b) 210 (c) 230 (d) 250 (e) 270 and (f) 290 min.
Chapter 5

Conclusions and future work

The preceding chapters have examined two complicated, but commonly observed kinds of squall line interactions. First, in Chapter 2, it was shown that the addition of a low-level jet can result in significant changes to squall line intensity by altering storm-relative inflow and low-level lifting. More specifically it was found that:

1. The relative direction of the low-level jet plays a role in modulating storm relative inflow. A jet directed towards the squall line increases storm-relative inflow, resulting in enhanced precipitation output. Conversely, a jet directed away from the squall line decreases storm-relative inflow, resulting in diminished precipitation output. These impacts become magnified as the boundary layer stabilizes and the layer of high-theta-e inflow becomes restricted to the layer of the low-level jet.

2. The addition of the low-level jet also alters the low-level vertical wind shear profile, which has implications for gust front lifting according to the theory of Rotunno et al. (1988). In the present simulations, a jet oriented toward the squall line produced favorable above-jet and unfavorable below-jet shear, while the jet oriented away from the squall line had the opposite effect. As with the storm-relative inflow, the relative impacts
of the jet-induced shear evolved as the boundary layer stabilized. While the system was surface-based the below-jet shear had the largest influence on the simulated squall line, and as the cooling progressed, the above-jet shear became increasingly important. This produced a complex evolution in the strength of gust front lifting, as each of the jet configurations produced favorable conditions for maximized ascent at different points in the simulation.

These results provide an example of how a time-varying background environment can alter squall line behavior by directly impacting the convective scale. The processes detailed in Chapter 2 may help explain observed changes to squall line organization in cases where the effective layer of parcels feeding the squall line is changing with time. Such conditions occur either as night falls and the boundary layer stabilizes, or as the sun rises and systems become surface based. Recent observational work by Marsham et al. (2011) provides an example of just such a scenario. As an initially elevated nocturnal MCS becomes surface-based after daybreak, it reorients and intensifies. Those authors hypothesized that a change in the shear layer interacting with the squall line’s cold pool (much as described in Chapter 2) may be partially responsible for the observed evolution. The emerging picture may be operationally useful as a means of anticipating squall line evolution in situations where the effective inflow layer to an MCS is changing.

Chapters 3 and 4 examined another process that can impact squall line evolution: mergers with nearby, isolated storms. In Chapter 3 common behaviors were identified for 21 cases wherein a squall line merged with an isolated supercell thunderstorm. Among the key findings from this analysis were:

1. In just about every case the squall line evolved into a bow echo post-merger, with the details of the ultimate organization appearing to correspond to different background envi-
environments. In particular, cases characterized by weak synoptic forcing and comparatively weaker vertical wind shear tended to produce large “system-scale” bow echoes whereas cases with strong synoptic forcing and stronger vertical wind shear tended to produce smaller-scale bow echoes along a longer squall line (the “hybrid” and “embedded” bowing structures).

2. Across all of the cases investigated, the merger appeared to correspond with a general trend toward storm rotation being maximized in the low-levels post-merger, although the exact pathway to this low-level rotation appeared to vary with the background environment.

3. Based on severe weather reports, different types of severe weather tended to occur at different points in the merger process. In particular, reports of large hail tended to occur prior to the merger (presumably with the isolated supercells), tornado reports were maximized right around merger time (consistent with the enhanced low-level rotation), and damaging wind reports were most common following the merger (presumably associated with the post-merger bow echo).

These results motivated the model simulations that were presented in Chapter 4, which aimed to isolate some of the processes responsible for the observed behaviors. The key findings from these simulations were:

1. The development of bow echo structures appeared to be related to the background environment used for the simulations, although the merger did appear to influence the details of the bowing, including the exact location and extent of the bowing structure.

2. The merger appeared to be responsible for developing enhanced surface winds, low-level vertical vorticity, and heavy rainfall, as all of these features were absent from the
simulation without the merger.

3. The merger exerts a significant influence on the squall line’s cold pool, causing a local weakening of the cold pool that facilitates the short-term maintenance of the supercell structure while also causing a weakening of the squall line north of the merger. Post-merger, downward transport of horizontal momentum from the squall line’s rear-inflow jet causes the squall line’s gust front south of the merged supercell to surge eastward, producing an enhanced bow echo structure.

4. The mechanism for enhanced low-level vertical vorticity in the simulated system appears to be tilting of low-level horizontal vorticity present in the squall line’s cold pool by the merged supercell. Specifically, an isolated core of strong $w$ associated with the remnants of the supercell moves rearward into a region of large horizontal vorticity in the squall line’s cold pool. Due to its isolated nature, this updraft produces the necessary gradients in $w$ to generate low-level vertical vorticity via tilting.

The combined results of Chapters 3 and 4 illustrate the significant impacts that storm mergers can have on squall line evolution. While a number of individual cases of this phenomena have been documented in the past (e.g. Goodman and Knupp 1993; Sabones et al. 1996; Wolf et al. 1996; Wolf 1998; Sieveking and Przybylinski 2004), the results presented in Chapter 3 are the first attempt to find common features over a relatively large number of squall line-supercell merge cases. Perhaps even more significantly, the results in Chapter 4 represent the first time (to the author’s knowledge) that such a behavior has been addressed using numerical simulations, and thus the first opportunity to explore the storm-scale details that drive these types of events.

This study was motivated by an interest in improving the understanding of how on-going squall lines evolve over time. Anticipating changes to squall line intensity and/or organization has important implications for short-term (e.g. 1-3 hour) forecasts of precipitation and severe
weather. One of the common threads that emerges from the results presented in Chapters 2-4 is that the processes that drive squall line evolution often occur on scales that are not commonly resolved in with the current observing network. In the case of the low-level jet interactions presented in Chapter 2, low-level jets and squall lines are frequently observed, however data may be lacking in terms of how deep the nocturnal stable layer is at a given point during the night or over which layers the jet-induced wind shear may be the strongest. In a similar vein, squall line-supercell mergers are readily observable, however the key processes responsible for the post-merger evolution, such the details of changes to the cold pool and low-level vorticity field, and vertical momentum transports usually are not. This makes it challenging to directly apply some of these findings in an operational framework. Rather, the findings of this work are perhaps best applied as conceptual models of what may happen given an ongoing squall line and a given set of circumstances. In this sense, by providing a better understanding of the processes that drive squall line evolution in the scenarios presented, this study can help forecasters make the most of the available data when trying to anticipate how an on-going squall line may evolve.

Finally, avenues for future work exist with both aspects of this study. Regarding the low-level jet results from Chapter 2, it would be useful to apply these findings to some observed cases to evaluate how significantly these processes may alter ”real-world” squall lines. The aforementioned recent work by Marsham et al. (2011) has noted the apparent role of changing shear layers with time in altering MCS behavior, and it would be interesting to examine some examples of this occurring in conjunction with a low-level jet, as in the present simulations. As mentioned at the end of Chapter 4, there are also a number of potential avenues for future work associated with the storm merger part of this project as well. One area that shows some promise is evaluating the sensitivity of the simulated merger to features such as the background environment, merger location, and relative maturity of the squall line-supercell. All of
these features have been hypothesized to potentially be important in delineating between the observed organizational archetypes (e.g., "system-scale", "embedded" and "hybrid" bowing). Additionally, simulations are underway using a full-physics case study model configuration in order to investigate mergers in the "strongly forced" environment. Through the combination of these additional idealized sensitivity tests, and the case study simulation a more complete picture of the full spectrum of observed merger evolutions may be realized.
REFERENCES


APPENDIX
As noted in Chapter 2, the relative humidity (RH) correction described by P08 was not applied consistently in his simulations, nor was it optimally designed. This appendix discusses the nature of the problem and the impacts on the results of P08.

a. Nature of the deficiency in P08

One of the primary problems with the way that P08 instituted the RH ≤ 0.98 correction is that the technique can lead to surprisingly low RH values in the simulated systems’ post-line regions. After several hours of artificial cooling, wave-like vertical displacements began to emerge in the stable low-level flow of P08’s simulations (see, for example, the isentropes in Figs. 12 and 16 of P08). As a part of this wave-like flow branch, inflowing air parcels would first ascend several hundreds of meters, and then would subside again. The problem with the P08 RH correction is illustrated by considering an initially saturated (or nearly saturated) air parcel undergoing such a vertical excursion. In nature, when super-saturation occurs due to the parcel’s vertical displacement, cloud droplets are formed. Such droplets remain in the airstream and are subsequently evaporated during the parcel’s downward displacement, yielding a parcel whose final characteristics are very similar to those with which it began. However, if the P08 RH correction is applied, the parcel’s RH is reset to 0.98 at the apex of its upward displacement, with the subsequent removal of any excess cloud water. The parcel therefore is subsaturated
during its downward displacement, yielding a RH that may be as low as 0.80 once it has returned to its original altitude.

In short, an unintended consequence of the P08 scheme’s implementation was that the RH correction in some cases removed condensate that should have been present due to the grid-scale processes in the model. Without delving too deeply into the various iterations that led up to the original P08 RH correction, it is clear that it has undesirable side effects. As explained in Section 2, in the present revised version, the artificial cooling routine is applied after the model microphysical parameterization, allowing the model to treat moist processes associated with the resolved motions first. Then, when the artificial cooling is added, only supersaturation and condensation caused specifically by the artificial cooling are removed. Thus, in the preceding thought experiment, the grid-scale ascent will produce cloud water that remains in the low-level flow as it proceeds into the post-line region, leading to much more realistic post-line RH values.

The negative consequence of the unrealistically depressed low-level RH values in P08 is that additional low-level evaporative cooling occurs when the squall line’s precipitation falls into the sub-saturated air stream. In other words, the convective outflow ends up being somewhat stronger than it should be. There are also some small impacts on the vertical force balance because the artificial removal of (legitimately produced) cloud water removes hydrometeor loading, meaning that parcel buoyancy in the low-levels was slightly less negative than it should realistically have been; however, the hydrometeor loading in this region of the P08 squall lines was actually quite small compared to the negative thermal buoyancy.

b. Impacts on the conclusions of P08

We reran all of the P08 simulations, uniformly applying the revised (improved) cooling and RH correction code. Our newly rerun simulations differ in some details, but continue to uphold the principle conclusions summarized on p. 1339 of P08. Much of the discussion in P08
followed from his “DEEP-unlim” run, which used the same environment and cooling profile as our current study. Interested readers may compare Figs. 2.4, 2.6a, and 2.11 in the present article with Figs. 5-9 in P08 to verify just how little is changed by the revised cooling/RH scheme. Indeed, very few of the plots from the new simulations are notably different from those that appear in P08. What follows are the exceptions.

First off, in the reruns using the Parker and Johnson (2004, “PJ04”) midlatitude MCS sounding and limited cooling, the simulated squall lines are less able to continue ingesting low-level air as time passes. In our rerun, we found that the system’s outflow was somewhat weaker (owing to the revised RH correction discussed above). As a consequence, the PJ04-289K squall line was much weaker and barely ingested any low-level air after roughly t = 7:30 (Fig. A1). A similar decrease in intensity also occurred when we reran the PJ04-287K (i.e. with 14K of cooling) simulation, such that the convection dissipated after 9h (not shown).

There were also noticeable differences in the rerun of P08’s “DEEP-unif” run, which utilized the sounding from the present study but in which the -3K h⁻¹ cooling was uniformly applied to every grid point below 1 km AGL (i.e. including the cold pool, not just the environment). In our rerun of the DEEP-unif case (Fig. A2), the system remained surface-based for a longer period of time (through t = 7:30, as compared to about 6:30 in P08). The rerun of DEEP-unif also displayed a much slower forward speed in its later stages, including some evidence of slowing/stalling during the period from t = 5:30 - 7:30 (Fig. A2). These changes are attributable to two effects. First, once again the system’s cold pool density current was less cold (cf. Fig. A2 to P08’s Fig. 15) due to the revised RH correction, which enabled the system to slow (i.e. a “stalling phase” became possible). And second, because the revised RH correction lessened the amount of evaporative cooling in the vicinity of the system’s developing bore, it took much longer for the bore to develop significant amplitude and outrun the surface density current.
Most other differences were minor. In short, it appears that the physical interpretations presented by P08 are robust, but that the system speeds and cold pool temperatures were overdone in some cases due to the poorly implemented RH correction.
FIGURE A2. As in Fig. 2.4, except for our rerun of the DEEP-unif simulation. This figure is directly comparable to Fig. 15 of P08.