Case study of the extreme high-shear/low-CAPE, strongly-forced Tennessee Valley QLCS of 12 February 2020

Adam Weiner

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CASE STUDY OF THE EXTREME
HIGH-SHEAR/LOW-CAPE, STRONGLY-FORCED
TENNESSEE VALLEY QLCS OF 12 FEBRUARY 2020

by

ADAM WEINER

A THESIS

Submitted in partial fulfillment of the requirements
for the degree of Master of Science
in
The Department of Atmospheric and Earth Science
to
The School of Graduate Studies
of
The University of Alabama in Huntsville

HUNTSVILLE, ALABAMA
2022
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THESIS APPROVAL FORM

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ABSTRACT

School of Graduate Studies
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Degree Master of Science College/Dept. Science/Atmospheric and
in Atmospheric Science Earth Science

Name of Candidate Adam Weiner

Title Case Study of the Extreme High-Shear, Low-CAPE
Strongly-Forced Tennessee Valley QLCS of 12 February 2020

High-shear, low-CAPE (HSLC) environments present unique operational forecasting challenges during the cool season across the southeastern US. Existing literature has sought to characterize the climatology of HSLC and Quasi-Linear Convective System (QLCS) environments as well as study bow echoes and mesovortices under moderate to large CAPE regimes. This case study investigates a cool season mid-Tennessee Valley severe weather event, where 200-300 J kg$^{-1}$ of CAPE and 0-1 km shear of 25 m s$^{-1}$ along with strong low-level forcing resulted in a severe weather event which was difficult to anticipate. Findings of this research suggest that bow echoes and mesovortices depend primarily on the strength and depth of low-level shear, with weak CAPE values not precluding their development or severe weather production.

Abstract Approval: Committee Chair
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I want to thank my friends and colleagues at the University of Alabama in Huntsville (UAH) for their support, discussions, and assistance throughout this process. In particular, I want to thank Huntir Cramer, Michael Graham, Clara Hochmuth, Kalitta Kauffman, Anthony Lyza, Dean Meyer, Preston Pangle, Dustin Phillips, Emily Wisinski, and all other individuals who participated in data collection on 12 February 2020 - this research would not have been possible without you. Coding assistance and discussions with Dean Meyer, Preston Pangle, Nick Perlaky, and Melissa Gonzalez-Fuentes were invaluable to the completion of this research. I also want to thank my committee - Dr. Kevin Knupp, Dr. Lawrence Carey, and Dr. John Mecikalski for their expertise and contributions to this research. In particular, I want to sincerely thank Dr. Knupp for his patience and fruitful discussions throughout this process as the research foci developed and evolved for this case study. I also want to sincerely thank my alma mater, Millersville University, and my professors (Dr. Clark, Dr. DeCaria, Dr. Sikora, Dr. Yalda, Mr. Hörst) for contributing substantially to my foundation in meteorology as they truly prepared me for success at UAH and beyond. Finally, I want to thank my parents and personal friends for their endless support of my endeavors throughout college and always believing in my abilities.
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CHAPTER 1

INTRODUCTION

An organized and persistent collection of deep, moist convection with a contiguous precipitation area extending at least 100 km in the horizontal defines the basic mesoscale convective system (MCS; American Meteorological Society, 2012). Quasi-linear convective systems (hereafter, QLCSs) are a type of MCS in which the long-axis dimension of the convective area greatly exceeds the short-axis dimension (Trapp, Tessendorf, et al., 2005; Smith, Thompson, et al., 2012). Early researchers described these types of systems as squall lines and documented their characteristically sharp pressure jumps, temperature drops, wind direction shifts, and wind speed increases, many of which occurred either along or ahead of a synoptic cold front. One hypothesis of squall line formation advanced by Tepper (1950) involved a “pressure jump line” induced by the temporary acceleration and deceleration of a cold front. This would result in a hydraulic jump during the acceleration which would foster the development of thunderstorms, followed by a faster-moving rarefaction wave during the deceleration. The pressure jump (and its associated squall line) would propagate away from the cold front as a gravity wave and the rarefaction wave would eventually overtake the pressure jump and cause the decay of the squall line. This idea was
challenged by Newton (1950), which suggested that squall lines typically originate
due to forcing over a cold frontal surface, then propagate off the cold front as a result
of the vertical mixing of strong horizontal momentum from aloft downward within
convective downdrafts. The resulting circulation within the convection would yield a
divergence pattern which placed the upper-level divergence maximum just ahead of
the low-level convergence maximum at the surface, thus maintaining a surface trough
of low pressure just ahead of the cold pool and supporting the continued development
of convection. Others have investigated the development of narrow cold frontal rain-
bands (Carbone, 1982; Moncrieff, 1989), which fit the general definition of a QLCS,
but are usually tied to intense, often slabular, low-level forcing along a synoptic cold
front (Heymsfield and Schotz, 1985; James, Fritsch, et al., 2005; Dial et al., 2010).
These ideas, among others cited within each study, present an array of conceptual
models of squall lines.

Since these early studies, a variety of characteristics associated with squall
lines beyond the fundamental changes observed across them have been investigated,
including the presence of horizontal wave structures within the line (Nolen, 1959),
the location of stratiform precipitation relative to the system (Parker and Johnson,
2000), the evolution of squall lines into bow echoes (Johns, 1993; Przybylinski, 1995;
Weisman, 2001; James, Markowski, et al., 2006), and the development of mesoscale
vortices (see Schenkman and Xue, 2016 for a complete review). In addition, the rela-
tionship between squall line longevity and the balance of horizontal vorticity induced
by strong vertical wind shear with baroclinically-generated horizontal vorticity in-
duced by the system’s cold pool, also known as RKW theory (Rotunno et al., 1988),
has received much attention in the literature as well (Weisman and Rotunno, 2004; Stensrud et al., 2005; Bryan et al., 2006; Parker, 2010).

Besides formation mechanisms and evolution patterns, researchers, especially within the last couple of decades, have sought to better understand the structures within QLCSs which are known to produce damaging winds and tornadoes. Documented heavily throughout the literature is the association of these life-threatening severe weather hazards with bow echoes and MVs, which may exist as multiple sub-system-scale features embedded within a larger QLCS or are the large-scale features which comprise the QLCS (Przybylinski, 1995; Weisman and Trapp, 2003; Trapp and Weisman, 2003; Wakimoto, Murphey, Nester, et al., 2006; Wakimoto, Murphey, Davis, et al., 2006; Wheatley, Trapp, and Atkins, 2006; Atkins and St. Laurent, 2009b; Atkins and St. Laurent, 2009a; Davis and Parker, 2014; Xu et al., 2015; Parker, Borchardt, et al., 2020). Furthermore, climatological studies have found that QLCSs often serve as the primary cause of severe weather during the cool season across the southeastern United States, with the threat zone gradually shifting northward into the summer months as the mean polar jet stream position migrates northward (Burke and Schultz, 2004; Trapp, Tessendorf, et al., 2005; Smith, Guyer, et al., 2008; Smith, Thompson, et al., 2012; Guyer and Dean, 2010; Sherburn and Parker, 2014; Ashley et al., 2019; Haberlie and Ashley, 2019).

The subject of this research is a mid-February QLCS which tracked across the Tennessee Valley supported by an environment of high-shear and low-CAPE (hereafter, HSLC), or one featuring strong vertical wind shear and weak buoyancy. HSLC environments have received increasing attention in the literature as of late (Dean
and Schneider, 2012; Davis and Parker, 2014; Sherburn, Parker, et al., 2016; King et al., 2017; Childs et al., 2018; Sherburn and Parker, 2019; Lovell and Parker, 2022), likely due to their frequent occurrence and common association with severe weather in the United States described in the aforementioned climatological studies. Considering the strong mid-upper-level dynamics often associated with the environments in which these systems form, strong forcing for ascent helps to initiate and maintain convection. Forcing is often supplied by strong cold fronts, which are typically tied to potent upper-level troughs that result in strong synoptic-scale forcing for ascent along or ahead of the front (Ashley et al., 2019; Celiński-Mysław et al., 2020). In pre-frontal squall line cases, pre-frontal troughs (Schultz, 2005) yield enough low-level convergence to initiate convection which organizes into the primary squall line well-ahead of the synoptic cold front. HSLC environments bring inherently lower predictability than high-CAPE, high-shear environments (Dean and Schneider, 2012; Sherburn, Parker, et al., 2016) and a propensity for rapid destabilization to occur on small temporal and spatial scales which may be difficult for high-resolution models to resolve accurately (Cohen, Cavallo, Coniglio, and Brooks, 2015; King et al., 2017). Case studies in the past have largely focused on mesovortexgenesis within QLCSs in moderate to high CAPE regimes, as these features are often associated with tornado development (McAvoy et al., 2000; Grumm and Glazewski, 2004; Conrad and Knupp, 2019; McDonald and Weiss, 2021; Lyza et al., 2021).

In the present case study, the QLCS of interest started out as a shallow, narrow band of convection reminiscent of a narrow cold-frontal rainband with largely slabular ascent similar to that described in Carbone (1982). As the QLCS approached north-
western Alabama, it gradually intensified into a squall line which featured distinctly cellular characteristics in its southern portion, slabular characteristics in its northern portion (James, Fritsch, et al., 2005), and a transition zone in between. In tandem with this change in character, updrafts strengthened, echo tops rose, and structures known to produce severe weather developed within the transition zone, including a single mesovortex (hereafter, MV) which was associated with thirteen wind damage reports and two weak (<EF-2) tornadoes without any severe weather warnings in effect during the first hour of severe weather production. The weak tornadoes corresponded in time with the presence of a weak echo hole (WEH) between 1-3 km above ground but were surveyed just south of the WEH position. A literature review of WEHs finds that a vast majority of studies investigated this feature as a structure within supercell hook echoes corresponding to the location of a tornado vortex, usually at very close range to the radar (Fujita, 1958; Fujita, 1965; Fujita, 1981; Wakimoto and Martner, 1992; Wakimoto and Liu, 1998; Bluestein and Pazmany, 2000; Wurman and Gill, 2000; Burgess et al., 2002; Wakimoto, Murphey, Dowell, et al., 2003; Schultz, 2014) while minimal attention has been given to their appearance in MVs (Atkins, Arnott, et al., 2004). Simply considering the extended period of time (> 2 hours) during which a WEH was detectable by radar within the MV, it is clear that this WEH was not associated with a tornado, but may instead be a consequence of the strong circulation (> $10^{-2}s^{-1}$) present within the MV. The MV in this case was not associated with a bow echo and was ultimately discernible for more than three hours while it tracked over 180 miles (294 km) from northwestern Alabama through southern Middle Tennessee, producing occasional swaths of tree,
powerline, and structural damage exclusively on its southern side. This single MV exceeds the spatial and temporal values of long-track (often tornadic) MVs which typically occur within much more volatile environments (Knupp et al., 2014; Lyza et al., 2021) and warrants an in-depth investigation. In addition to the MV, a trio of closely-spaced meso-γ-scale (Orlanski, 1975) bow echo segments also produced occasional wind damage south of the MV. When the northernmost bow echo segment merged with the southern side of the MV, the most widespread (in terms of areal extent) wind damage took place. The QLCS transitioned back to a slabular state after a few hours and with this transition came the end of the severe weather threat.

The primary goal of this work is to investigate the environmental conditions, the severe-weather-producing elements, and the overall evolution of this QLCS, with a focus on the kinematics. Since the case study of a tornadic narrow cold-frontal rainband (Carbone, 1983), the author is unaware of any detailed and strictly observational case studies which have specifically investigated MVs occurring within cool season HSLC QLCSs, although a very recent publication (Lovell and Parker, 2022) has employed a case study approach, albeit using numerical modeling. Comparing the findings of this research with peer-reviewed literature will help characterize this case and bring to light a situation in which extreme values of vertical wind shear (0-1 km shear vector magnitude $\sim 25$ m s$^{-1}$; 0-1 km storm-relative helicity $\sim 500$ m$^2$ s$^{-2}$) were able to overcome weak CAPE (MLCAPE $\sim 200$ J kg$^{-1}$) and yield a notable event where severe weather was known to be possible, but still failed to be anticipated in the operational setting. Specific science objectives include: 1) defining the environmental conditions and forcing mechanism(s) for this QLCS; 2) examining the evolution
characteristics of this QLCS prior to severe weather production; 3) characterizing the kinematic properties of the MV and bow echoes observed and comparing these observations in HSLC conditions with the published literature, which primarily examined these features in non-HSLC conditions; and 4) quantifying the rapid changes observed in the pre-QLCS environment to understand its role in conditioning the pre-storm environment. Chapter 2 consists of an overview of relevant modeling and observational studies. The data and methodology applied to this case study, as well as the limitation of these methods, are discussed in Chapter 3. Chapter 4 supplies a more in-depth overview of the case itself and the environmental setup. Results concerning the development, evolution, and severe weather production within the system are provided in Chapter 5. Finally, a summary of this research and recommendations for future work are described in Chapter 6.
CHAPTER 2

BACKGROUND

Observational case studies of weather events which occur within marginal environments present an excellent opportunity to evaluate the applicability of broader observational and numerical modeling research to these fringe situations. In this case study, a quasi-linear convective system (QLCS) occurring within rapidly-evolving high-shear, low-CAPE (HSLC) conditions will be the subject, with an emphasis on the long-track MV and meso-γ-scale bow echoes which produced a majority of the severe weather. Research addressing this collection of topics will serve as the primary focus of this chapter. In addition, given that the QLCS in this case was not forced along a synoptic cold front and was consequently removed from a majority of the upper-level divergence associated with 300 mb troughing over the middle of the country, context on pre-frontal forcing mechanisms will be provided as well. Climatological research on squall lines and linear storms will be presented first, followed by studies on HSLC environments and finally the formation and evolution of bow echoes and MVs.

Climatological studies of severe weather produced by linear storms from the late 1980s through the early 2010s covered relatively “short” time frames in their
analyses, ranging from three to eleven years. This may be attributed to the labor-intensive nature of manually gathering, examining, and classifying severe weather events based on radar data. Burke and Schultz (2004) manually examined archived national radar mosaics for fifty-one bow echoes which occurred during the cool seasons (October through April) of 1997-98 through 2000-01. The authors found that a vast majority of observed bow echoes occurred in a region stretching from the southern and central plains eastward through the southeast and mid-south states. Monthly distributions showed the top three months (accounting for 84% of the total) for bow echo formation were February through April. Environmental conditions associated with these bow echoes included strong vertical wind shear, with a mean 0-2.5 km (0-5 km) shear magnitude of 14 m s\(^{-1}\) (23 m s\(^{-1}\)), as well as a mean most-unstable CAPE (MUCAPE) of 1366 J kg\(^{-1}\). The relative lack of low-CAPE (<500 J kg\(^{-1}\)) observations may be related to the 300 km distance and 300-minute threshold employed to retrieve “proximity” soundings. However, the authors noted that two long-lived severe bow echoes featured CAPE values in the range of 76-522 J kg\(^{-1}\) and these primarily occurred in the southeastern US. The synoptic pattern conducive to bow echo development in the Gulf Coast states included a mean 500 mb trough axis over the middle of the contiguous United States with southwesterly flow over the southeast states.

Smith, Guyer, et al. (2008) manually analyzed radar data to classify convective mode for severe weather reports which occurred during the cool season (October through March) between 1995 and 2006 within the Ohio and Tennessee river valleys. QLCSs accounted for 62% of the 5967 total severe weather reports identified during
the study period, with 80% of all severe wind reports associated with QLCSs. Among the tornadoes produced by QLCSs, 79% of these were either F0 or F1. Environmental conditions associated with nineteen tornadic QLCSs included a median MLCAPE (SBCAPE) of 534 J kg$^{-1}$ (236 J kg$^{-1}$) along with a median 0-1 km (0-3 km) shear magnitude of 19.5 m s$^{-1}$ (23.2 m s$^{-1}$). Although sample sizes were small and this precludes a generalized comparison with other studies, an indication that weaker CAPE and stronger low-level shear can still support tornadic QLCSs is expressed. Larger studies of tornadoes across the contiguous United States (Trapp, Tessendorf, *et al.*, 2005; Guyer and Dean, 2010; Smith, Thompson, *et al.*, 2012) all support the findings described above, namely that a majority of tornadoes which occur within weak CAPE environments develop during the cool season, are associated with QLCS storm modes, and are typically most prevalent within the southeastern United States. Furthermore, when QLCSs do produce tornadoes, they are primarily (E)F-0 or (E)F-1 intensity while supercell storm modes account for a majority of (E)F-2+ tornadoes.

Ashley *et al.* (2019) performed the largest climatological study of QLCSs to date by examining an unprecedented twenty-two-year period from 1996 to 2017 using automated machine learning methods developed by Haberlie and Ashley (2019). Radar mosaic images were segmented, mesoscale convective systems (MCS) were classified, and the images containing an MCS were passed through a convolutional neural network to return a probability that the identified systems were QLCSs. These were detected as regions where radar reflectivity factor equaled or exceeded 40 dBZ over a length of at least 100 km which were at least three times as long as they were wide. Only probabilities of 95% or greater were accepted and then severe weather
reports were attributed to each QLCS using a 20 km buffer. Key findings from the 3064 identified QLCSs include a maximum in mean annual QLCS occurrence over the mid-Mississippi and western Tennessee river valleys, a gradual increase in mean monthly QLCS occurrence throughout the cool season across the southeastern US which peaks in April, and a clear pattern of QLCS initiation during the late afternoon and evening hours. In terms of severe weather hazards, QLCSs contributed a substantial number of severe weather reports in the southeast US, particularly during the cool season. Tornadoes produced by QLCSs across the study domain were primarily (76% of all QLCS-attributable tornadoes) in the range of EF-0 to EF-2 strength. During the cool season, QLCS tornadoes were found to occur in greater proportions relative to other storm modes and accounted for a majority of reported tornadoes. Severe wind reports attributable to QLCSs were most common across the study domain during the cool season as well (accounting for a mean 58% of all severe wind reports in February during the study period), with a distinct maximum in the southeast US. Observed temporal trends follow similar patterns for QLCS tornadoes and severe wind, with a clear maximum in reports during the late afternoon and evening hours.

Given the QLCS in this case study developed during the mid-late afternoon of a day in mid-February, this event took place during a climatologically-favorable time of day and during a time of year in which QLCS occurrence is approaching its maximum. Subsequently, these aspects of the event align with the expectation for southeast US QLCSs so far. However, consideration must now be given to the environmental setup
as the conditions in which this QLCS occurred were rather extreme relative to HSLC environments described in the literature.

Large-scale environmental studies of HSLC conditions have received considerable attention during the last decade (Guyer and Dean, 2010; Thompson et al., 2012; Sherburn and Parker, 2014; Sherburn, Parker, et al., 2016). Thompson et al. (2012) studied the environments of thunderstorms which produced tornadoes and significant severe weather between the years of 2003 and 2011 using hourly objective analyses produced and archived at the National Weather Service’s Storm Prediction Center. Overall, QLCS tornado events featured considerably lower MLCAPE values compared to non-QLCS storm modes with 45% of all QLCS tornado events occurring with MLCAPE under 500 J kg$^{-1}$. Vertical wind shear, particularly in the 0-1 km layer, showed a clear distinction between nontornadic and significantly-tornadic QLCSs, with median values of 21 kts versus 40 kts, respectively, and 90th percentile values of 42 kts versus 60 kts, respectively. In terms of seasonality, the authors found that vertical wind shear values associated with wintertime QLCS tornadoes were similar to that observed in supercell tornado environments, but a majority of wintertime QLCS tornado events occurred with MLCAPE below 400 J kg$^{-1}$. While these efforts helped to identify a tendency for QLCS tornadoes to occur primarily during the cool season and in considerably weaker CAPE environments than other storm modes, the main focus of this study was to distinguish nonsevere from significantly-severe environments for QLCS versus non-QLCS storm modes. As a result, this study left a knowledge gap regarding the subtleties of severe weather occurring strictly within HSLC environments.
The efforts of Guyer and Dean (2010), Sherburn and Parker (2014), and Sherburn, Parker, *et al.* (2016) represent the vast majority of literature specifically investigating HSLC severe weather environments in the contiguous United States. Aside from the climatological aspect described previously, Guyer and Dean (2010) also examined environmental parameters associated with weak CAPE (MLCAPE ≤ 500 J kg\(^{-1}\)) tornadoes during the period 2003-2009, including 2587 tornadoes. Remarkably, 1410 (54.5\%) of these tornadoes were associated with MLCAPE less than or equal to 250 J kg\(^{-1}\). Measures of vertical wind shear were found to be similar or stronger in magnitude compared to tornadic environments where MLCAPE exceeded 500 J kg\(^{-1}\), which agrees with findings by Thompson *et al.* (2012). In terms of thermodynamic parameters, weak CAPE tornado environments featured an MLCAPE (SBCAPE) median of 230 (388) J kg\(^{-1}\), 0-3 km and 700-500 mb lapse rate medians of 5.8 K km\(^{-1}\) and 6.0 K km\(^{-1}\), respectively, and surface temperature (dewpoint) medians of 68°F (20°C) and 62°F (16.7°C). While cooler conditions and weaker lapse rates are common in these weak CAPE environments, greater surface relative humidity, lower lifting condensation level heights, and similar amounts of total precipitable water were found between weak CAPE and stronger CAPE environments, which can help maintain conditions supportive of tornado formation regardless of the shallow nature of convection.

Studies addressing the subtleties of HSLC environments across the contiguous United States over the period 2006-2011 were undertaken in earnest by Sherburn and Parker (2014) and Sherburn, Parker, *et al.* (2016) with the former emphasizing significantly-severe HSLC environments while the latter focused on differentiating
nonsevere from severe HSLC environments. In both studies, HSLC conditions were defined for SBCAPE less than or equal to 500 J kg\(^{-1}\), MUCAPE less than or equal to 1000 J kg\(^{-1}\), and 0-6 km bulk wind difference of at least 18 m s\(^{-1}\). In terms of spatial distribution Sherburn and Parker (2014) found that significant tornado and wind reports were most common from the mid-Mississippi Valley through the southeast states during the cool season. When evaluating which environmental parameters distinguished most clearly the significant severe weather events from nonsevere events in the southeast, the authors found that 0-3 km lapse rate, 700-500 mb lapse rate, and 0-3 km bulk wind difference performed the best overall, and when normalized and combined into a metric called the severe hazards in environments with reduced buoyancy (SHERB) parameter, yielded the highest skill in separating significantly-severe from nonsevere environments. This SHERB parameter, and more specifically its 0-3 km bulk wind difference counterpart (SHERBS3), will help provide context in terms of the rapid evolution of the environment ahead of the QLCS in section 5.4. It is anticipated that SHERBS3 parameter should indicate an environment supportive of HSLC severe weather in this case due to strong low-level shear, but the low- and mid-level lapse rates may be a limiting factor due to antecedent cool and stable conditions early in the day.

Whereas the former study identified key ingredients of vertical atmospheric profiles within significant severe HSLC environments, Sherburn, Parker, \textit{et al.} (2016) expanded on this to identify composite synoptic environments which were typically associated with HSLC severe weather using the North American Regional Reanalysis (NARR; Mesinger \textit{et al.}, 2006). Compositing 997 NARR analyses for HSLC events
in the southeast revealed that typical synoptic setups featured strong synoptic-scale forcing for ascent driven by 300 mb troughing over the center of the country and a potent 500 mb vorticity maximum centered over Missouri. Surface low pressure centered over the western Ohio Valley drives strong low-level warm air advection out of the Gulf of Mexico which results in a tongue of enhanced 0-3 km CAPE and steeper 0-3 km lapse rates compared to nonsevere (null) cases. The SHERB parameter using 0-3 km bulk wind difference discussed earlier showed a clear maximum over the Tennessee Valley at its designed threshold value of 1 and testing of a variety of environmental parameters verified its high skill over various other combinations of variables and composite indices. However, the authors also found a distinct signal when examining the release of potential instability as a contributing factor to severe HSLC events in the southeast when compared to null events and created an even more skillful modified SHERB (MOSH) which considers the product of upward motion and vertical equivalent potential temperature gradient. Overall, the literature on HSLC environments have identified dynamic, strongly-forced environments featuring strong vertical wind shear and enhanced warm air advection, which yields relatively narrow pre-frontal zones of enhanced CAPE, as conducive to severe weather production. With these factors in mind, an opportunity exists to examine the relevance of these parameters within a case study for a severe weather event which featured environmental conditions towards the extreme ends of the HSLC spectrum (MLCAPE $\sim 200$ J kg$^{-1}$; 0-1 km shear magnitude $\sim 25$ m s$^{-1}$).

While the typical synoptic conditions described above were present on the day of this case study, peculiarities arise when considering the specific forcing mechanism.
The narrow line of showers which would become the QLCS in this case originated from a remnant surface trough/wind shift which originally extended north from a relatively weak surface cyclone (\(\sim 1009\) mb at 12 UTC, 12 February 2020) in the northwestern Gulf of Mexico. The weak surface cyclone and lack of a defined cold front at the time of QLCS formation complicates the argument that synoptic-scale forcing for ascent directly initiated this QLCS. In fact, as it will be shown in section 5.1, a surface cold pool was directly responsible for the formation of a shallow line of convection which would ultimately become the QLCS, not a surface cold front.

King et al. (2017) studied the rapid evolution of cool-season HSLC environments using real-data simulations of eleven severe events and six nonsevere events. To examine each case, the North American model’s 12 km analyses served as input to the Advanced Research version of the Weather Research and Forecasting model (WRF-ARW; Skamarock et al., 2008) with a 9 km horizontal grid spacing and a one-way nested grid featuring 3 km horizontal grid spacing to resolve convection explicitly. An SBCAPE threshold of up to 1000 J kg\(^{-1}\) and 0-3 km shear magnitude of at least 18 m s\(^{-1}\) served as the criteria for HSLC in this study. Results showed that 0-1 km shear remained generally steady during the three hours leading up to the arrival of convection, with nearly all severe cases featuring a 0-1 km shear magnitude exceeding 20 m s\(^{-1}\). Meanwhile, SBCAPE values increased in all cases, often by a couple to several hundred J kg\(^{-1}\), but it was found that severe events featured a greater maximum SBCAPE than nonsevere events. When evaluating the contributing factors to destabilization, surface moistening occurred in all cases while surface warming occurred primarily in severe events and steady or cooling temperatures oc-
curred in nonsevere events. Importantly, all simulations produced substantial cloud cover exceeding 80% prior to convection arriving, indicating that strong warm air advection was the primary source of surface warming in these HSLC environments, not solar radiation or sensible heat flux. Lastly, the authors expressed that the release of potential instability may be an important mechanism for severe HSLC environments, particularly when strong forcing for ascent is occurring. Thus, this research identified that rapid (≤ 3 hours) destabilization is a key characteristic of HSLC environments with surface warming by advection being a top contributor to SBCAPE rises in severe environments while surface moistening is common across all HSLC situations. This information is relevant in this case study as rapid destabilization is hypothesized to have occurred ahead of the QLCS and strong linear forcing along a gust front likely resulted in the release of potential instability. This will receive more attention in section 5.4.

The final primary topic of interest in this case study is the development of, and severe weather production within, bow echoes and MVs. Since the single MV in this case study was not a bookend vortex and not a transient appendage produced by horizontal shearing instability (e.g., Conrad and Knupp, 2019), these particular types of MVs will not receive attention in this section. The bow echoes produced in this case were meso-γ-scale structures which translated northeastward along the QLCS as opposed to being a larger-scale structural element of the QLCS, so these smaller-scale bow echoes will also be a focal point in this section. Weisman (2001) provides a thorough review of bow echo literature from the late 20th century, which was heavily rooted in the efforts of Tetsuya Fujita. Primary findings by Fujita (1978) included a
general lifecycle of bow echo formation from an initial strong/tall echo to a bow echo to a comma-shaped echo with a cyclonic mid-level vortex at its head, although the true dynamical explanations behind this evolution would require additional research over the years. Sometimes the cyclonic vortex can last for multiple days in the form of a mesoscale convective vortex (MCV; Bartels and Maddox, 1991; Davis and Weisman, 1994; Davis, Atkins, et al., 2004). Later researchers (Weisman, 2001; Markowski and Richardson, 2010) would find that bow echoes can exist on a variety of scales and are typically produced when an updraft tilts rearward, allowing the latent heat of condensation to produce hydrostatically lower pressure above the surface cold pool and within the stratiform precipitation. This low pressure area strengthens over time and subsequently drives a rear-to-front flow that descends into the apex of the bow beneath the front-to-rear airstream which feeds the updraft. Damaging winds often occur where this rear-inflow jet reaches the surface, although this may occur in a narrow zone just behind the gust front or can occur well-behind the gust front.

Mesovortices differ from the MCV mentioned above in both spatial and temporal scale, with the former typically lasting on the order of minutes to hours and featuring a diameter of order one to ten kilometers while the latter can exist for days and cover tens to hundreds of kilometers. Several formation mechanisms have been proposed in the literature and each will receive attention next. Weisman and Trapp (2003) and Trapp and Weisman (2003) performed idealized numerical simulations of QLCSs to study the formation of low-level (up to 1 km above ground) meso-γ-scale (order 2-20 km diameter; Orlanski, 1975) MVs within a variety of vertical shear regimes. Key findings include a clear tendency for deeper, more intense,
and longer-lasting MVs to develop primarily to the north of embedded bow echoes within QLCSs when vertical shear magnitudes reach or exceed 20 m s\(^{-1}\) over the lowest 2.5-5 km above ground. The authors directly related these observations to the tendency for stronger low-level shear to yield more intense, upright, and deeper updrafts at the leading edge of the QLCS \(\text{e.g., Rotunno et al., 1988}\). Generation of a low-level vortex couplet was determined to occur through the tilting of crosswise baroclinically-generated horizontal vorticity within a convective downdraft, resulting in cyclonic (anticyclonic) vertical vorticity south (north) of the downdraft (see Fig. 2.1). Stretching of omnipresent positive planetary vorticity was also found to be a crucial factor in strengthening the cyclonic vortex and weakening the anticyclonic vortex in these simulations. The downward-directed vertical perturbation pressure gradient dynamically-induced by intense low-level circulations within simulated MVs tended to suppress any updraft at the system’s leading edge, thus causing a kink in the QLCS structure. Strong horizontal winds (which were found in some simulations to exceed that observed at the apex of bow echoes) typically found on the right flank of MVs (relative to their direction of motion) were determined to result from a combination of the high perturbation pressure induced by the cold pool of the QLCS in tandem with the low perturbation pressure at the center of the MV, thus creating a zone where parcels experience a rapid acceleration due to the enhanced horizontal pressure gradient (see Fig. 2.2).

In light of the recommendations for future work provided by Trapp and Weisman (2003), Wheatley, Trapp, and Atkins (2006) performed an observational analysis of five bow echo events which occurred during the Bow Echo and Mesoscale Convec-
Figure 2.1: Schematic for the proposed formation of cyclonic-anticyclonic vortex pairs within a QLCS. Adapted from Trapp and Weisman (2003) Fig. 23.
Figure 2.2: Schematic for the proposed impact of vertical perturbation pressure and the rear inflow jet on the production of damaging winds within MVs and bow echoes embedded within QLCS. Adapted from Trapp and Weisman (2003) Fig. 24.
tive Vortex Experiment (BAMEX) to compare wind damage locations in relation to bow echo apexes and MVs with the aforementioned results of numerical modeling experiments. These individual events verified the modeling experiments, both in terms of the location of wind damage corresponding to where descending rear-inflow spreads laterally at the surface at the apex of bow echoes and in terms of MVs producing more intense damage than bow echoes within the same convective system. Additional observational investigations by Wakimoto, Murphey, Nester, et al. (2006) hypothesized that intense surface winds produced by MVs were not strictly a result of the enhanced horizontal pressure gradient between the cold pool and the MV, as this would result in damaging surface winds terminating nearer to its center of circulation, where the negative dynamic pressure perturbation is greatest. Instead, the authors found that these intense surface winds were produced by a superposition of the horizontal pressure gradient with the flow in which the MV was embedded, thus causing the strongest winds to occur where translational and rotational motions were in the same direction. In addition, while the MV formation mechanism was determined to be a downdraft (e.g. Trapp and Weisman, 2003), its source was not from a mature thunderstorm, but rather stemmed from subsidence on the periphery of a developing cell near the leading edge of the bow echo’s outflow boundary (see Fig. 2.3).

Atkins and St. Laurent (2009a) and Atkins and St. Laurent (2009b) further investigated bow echo MVs using numerical simulations of a bow echo observed during BAMEX with similar attention paid to the 0-2.5 km (low-level) and 0-5 km (deep-layer) shear magnitudes as Trapp and Weisman (2003). Key findings included that stronger MVs occurred when the horizontal vorticity induced by the low-level
Figure 2.3: Schematic for the proposed development of MVs within QLCSs. Adapted from Wakimoto, Murphey, Nester, et al. (2006) Fig. 12.
vertical wind shear balanced that produced baroclinically by the cold pool, which yielded stronger and more vertical updrafts that resulted in greater vertical vorticity stretching within MVs (e.g. Rotunno et al., 1988. Overall, the strength and number of MVs decreased as the magnitude of either low-level or deep-layer shear decreased while increasing values of Coriolis parameter yielded both more intense and a greater number of MVs. Lastly, it was suggested that the most damaging surface winds may occur when the southern periphery of an eastward-moving MV overlaps with a descending rear-inflow jet feeding into the rear of a bow echo (see Fig. 2.4), leveraging the same mechanisms described by Wakimoto, Murphey, Davis, et al. (2006). In terms of MV genesis mechanisms, the authors proposed two new explanations. In cases where only a cyclonic MV forms without an anticyclonic partner, air parcels descending within a convective downdraft which acquired baroclinically-generated horizontal vorticity were tilted upward and stretched as a result of an updraft along the gust front edge (see Fig. 2.5). Where a cyclonic-anticyclonic vortex pair forms, it was determined that baroclinically-generated horizontal vorticity along the gust front edge was tilted upward by an updraft maximum caused by a localized acceleration of the gust front (see Fig. 2.6). The acceleration of the gust front resulted from an enhanced convective downdraft, the source of which was not discussed.

In general, the aforementioned authors proposed that the production of damaging winds on the right-flank of MVs is due to a combination of the horizontal perturbation pressure gradient acceleration between the MV center and the surface cold pool of the QLCS, the rotational and translational motion of the MV, and in some cases, the overlap of these with the descending rear inflow jet feeding into the back of
**Figure 2.4:** Schematic for the proposed damaging wind production mechanism in QLCS MVs. Adapted from Atkins and St. Laurent (2009a) Fig. 14.
Figure 2.5: Schematic for the proposed development of cyclonic-only MVs within QLCSs. Adapted from Atkins and St. Laurent (2009b) Fig. 15.
**Figure 2.6:** Schematic for the proposed development mechanism of a pair of cyclonic-anticyclonic MVs within QLCSs. Adapted from Atkins and St. Laurent (2009b) Fig. 16.
a bow echo. The reader should remain cognizant of these contributing factors when reading section 5.3, particularly since the experiments conducted and observations examined by these authors were for moderate to large CAPE regimes (> 2000 J kg\(^{-1}\)). Since Evans and Doswell (2001) showed that derecho environments can include very low CAPE, additional modeling and case study work is needed to determine if there are differences in the formation and damaging wind mechanisms under low-CAPE regimes.

Aside from the primary MV formation mechanisms described above, other researchers have investigated the importance of heterogeneities in the pre-storm environment to the production of MVs and bow echoes. Wheatley and Trapp (2008) found through numerical simulations of real QLCS cases which occurred during warm and cool seasons that meso-\(\gamma\)-scale and meso-\(\beta\)-scale (20-200 km; Orlanski, 1975) heterogeneities resulting from the presence of airmass boundaries had no distinct impact on the development of MVs, with the environmental conditions alone supporting their development before encountering any boundaries. James, Fritsch, \textit{et al.} (2005) studied the influence of water vapor mixing ratio on the development of bow echoes within QLCSs using numerical simulations. The authors found that drier conditions produced stronger cold pools which yielded a more two-dimensional system with slabular lifting along the cold pool’s edge while greater moisture content in the low- and mid-levels produced weaker cold pools which allowed for more three-dimensional cellular elements to dominate in the QLCS’s structure. Xu \textit{et al.} (2015) performed numerical simulations for a severe bow echo event which occurred in the central US during the warm season with an emphasis on the role of frictionally-generated horizontal vorticity.
on the genesis of MVs in this case. The authors found that while the low-level vertical vorticity of the MVs stemmed from the tilting of baroclinically-generated horizontal vorticity into the vertical via updrafts or downdrafts, near-surface horizontal vorticity induced by friction in combination with convergence induced by the descending rear-inflow jet also contributed significantly to vertical vorticity in the lowest 300 m above ground. In summary, these studies highlight the robustness of MV formation primarily resulting from a favorable combination of strong environmental wind shear and linear storm organization while external factors such as boundaries and friction play a considerably smaller role in their development. With relevant background information on the important topics for this case study established, a review of the data and methods applied to conduct this case study is provided in the next chapter.
Radar data collected by several National Weather Service radars and the University of Alabama in Huntsville’s (UAH) Advanced Radar for Meteorological and Operational Research (ARMOR; UAH, 2022d) as well as their Mobile Alabama X-band (MAX; UAH, 2022c) radar were utilized in this research. In addition, Automated Surface Observing System (ASOS; NOAA/NWS, 2022) data from numerous airports across the lower Mississippi Valley and Tennessee Valley provided surface observations to diagnose and quantify changes in the environment. Furthermore, UAH had deployed its collection of research vehicles, including those associated with the Mobile Atmospheric Profiling Network (MAPNet; UAH, 2022b) on this day. Thus, a variety of profiling instruments collected kinematic and thermodynamic profiles across north Alabama throughout this event as well. Personal weather stations (Weather Underground, 2022) located across southern Middle Tennessee were also examined to provide a more complete picture of the pre-storm environment. More details on each of these data sources are provided below.
3.1 Radar Data

Much of the radar data analyzed in this case study came from Weather Surveillance Radar – 1988 Doppler (WSR-88D) S-band radars located in the lower Mississippi Valley and central Tennessee Valley (see Fig. 3.1). These data were sourced from the Next Generation Weather Radar (NEXRAD) Level-II archive available from the National Centers for Environmental Information (NOAA/NWS Radar Operations Center, 2022). In addition, data collected by the ARMOR, which is located at Huntsville International Airport, and the MAX radar, which was located near Tanner, AL for this case, will be interrogated during the analysis. ARMOR and MAX data are available from an archive server located at UAH and can be obtained by contacting the principal investigator or field manager (UAH, 2022a). Selected specifications related to each type of radar are provided in Table 3.1. Only the base radar products collected by each radar are of primary interest in this study, including reflectivity, radial velocity, spectrum width, differential reflectivity ($Z_{dr}$), copolar correlation coefficient (CC), differential propagation phase ($\Phi_{dp}$), and specific differential phase ($K_{dp}$).

The WSR-88D radar data considered in this study were observed between 12 UTC on 12 February 2020 and 04 UTC on 13 February 2020 using volume coverage patterns (VCP) 212 (for convection) or 215 (for general scanning in precipitation). VCP 212 uses elevation angles 0.5°, 0.9°, 1.3°, 1.8°, 2.4°, 3.1°, 4.0°, 5.1°, 6.4°, 8.0°, 10.0°, 12.5°, 15.6°, and 19.5°, which is normally completed in under five minutes while VCP 215 uses the same elevation angles up to 10°, then scans at 12.0°, 14.0°, 16.7°, and 19.5°, which is normally completed in about six minutes. The radars at
Figure 3.1: Map of fixed radar locations utilized in this study. All radars shown are WSR-88D NEXRADs except for ARMOR, which is a C-band research radar operated by UAH.
Table 3.1: Selected specifications related to each type of radar used in this study. WSR-88D specifications are maintained by the Radar Operations Center (NOAA/NWS Radar Operations Center, 2022). The University of Alabama in Huntsville maintains ARMOR and MAX specifications (UAH, 2022d)

<table>
<thead>
<tr>
<th>Radar</th>
<th>WSR-88D</th>
<th>ARMOR</th>
<th>MAX</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location(s)</td>
<td>Various</td>
<td>Huntsville Intl. Airport</td>
<td>Tanner, AL</td>
</tr>
<tr>
<td>Band</td>
<td>S-band</td>
<td>C-band</td>
<td>X-band</td>
</tr>
<tr>
<td>Wavelength</td>
<td>10-11 cm</td>
<td>5.3 cm</td>
<td>3.2 cm</td>
</tr>
<tr>
<td>Peak Power</td>
<td>750 kW</td>
<td>350 kW</td>
<td>250 kW</td>
</tr>
<tr>
<td>Pulse Widths</td>
<td>1.57,4.57 µs</td>
<td>0.4,0.8,1.0,2.0 µs</td>
<td>0.4,0.8,1.0,2.0 µs</td>
</tr>
<tr>
<td>Polarization</td>
<td>Dual</td>
<td>Dual</td>
<td>Dual</td>
</tr>
<tr>
<td>Beamwidth</td>
<td>0.925°</td>
<td>1.1°</td>
<td>0.95°</td>
</tr>
<tr>
<td>Gate Spacing</td>
<td>250 m</td>
<td>250 m</td>
<td>125 m</td>
</tr>
<tr>
<td>First Sidelobe</td>
<td>-29 dB</td>
<td>-30 dB</td>
<td>-31 dB</td>
</tr>
</tbody>
</table>

Hytop, AL (KHTX); Nashville, TN (KOHX); Columbus, MS (KGWX); Jackson, MS (KDGX); Shreveport, LA (KSHV); and Fort Polk, LA (KPOE) all utilized VCP 212 during this event when convection was present while the radars at Fort Smith, AR (KSRX); Little Rock, AR (KLZK); and Memphis, TN (KNQA) used VCP 215 as generally stratiform rainfall occurred in those locations. While the elevation angles above represent the full range of angles that may be employed in a scan strategy, the Automated Volume Scan Evaluation and Termination (AVSET), Supplemental Adaptative Intra-Volume Low-Level Scans (SAILS), and Mid-volume Rescan of Low-level Elevations (MRLE) all can affect the number and sequence of elevation angles in order to adapt to rapidly-changing weather conditions. AVSET allows for the early termination of a volume scan if minimal activity is detected whereas SAILS
and MRLE allow the radar operator to carry out additional low-level scans within one volume scan, if desired. Both AVSET and SAILS were employed by many of the radars used in this research. In addition, it is important to note that the lowest three elevation angles of VCP 212 and 215 utilize a split-cut technique which permits the usage of two separate pulse repetition frequencies (PRF) during each scan (which requires two sweeps per elevation angle), thus permitting a greater unambiguous range and greater unambiguous velocity at each of these elevations (NOAA/NWS Warning Decision Training Division, 2022b). Furthermore, these VCPs utilize a set of range-folding mitigation strategies which help to maximize the amount of data retrieved at longer ranges (NOAA/NWS Warning Decision Training Division, 2022a). Lastly, these radars feature two different azimuthal data resolutions, with the lowest three tilts of VCP 212 and 215 using super-resolution (0.5° azimuthal spacing) while the remaining upper tilts use legacy resolution (1° azimuthal spacing).

The ARMOR and MAX radars employed the same elevation angle sequence when convection was within range of each radar, specifically 0.7°, 1.3°, 2.0°, 3.1°, 4.3°, 5.5°, 6.7°, 7.8°, 9.0°, 10.2°, and 11.5°. These radars were not time-synchronized in their scanning, thus resulting in irregular time intervals between when their scan start times lined up, both with each other and with KHTX. The challenges associated with this irregularity will be discussed in section 3.6 of this chapter. ARMOR operated with a single PRF of 1200 Hz throughout this event, yielding a Nyquist velocity of 16 m s\(^{-1}\) based on the relation \(V_{\text{max}} = \pm (\text{PRF } \lambda ) / 4\), where \(\lambda\) is the wavelength of the radar in meters Rauber and Nesbitt, 2018. MAX also operated with a single PRF of 1200 Hz, yielding a Nyquist velocity of 9.5 m s\(^{-1}\). Given the very strong vertical wind
shear which was present on this day, extensive velocity folding occurred in the raw data collected by these radars.

3.2 MAPNet Data

The UAH MAPNet fully deployed its array of research vehicles for the 12 February 2020 quasi-linear convective system (QLCS) event (see Fig. 3.2), including the Mobile Integrated Profiling System (MIPS), the Rapidly-Deployable Profiling System (RaDAPS), the Mobile Doppler LiDAR and Soundings (MoDLS) system, and the aforementioned MAX dual-polarization radar. These first three vehicles feature vertical wind profilers, vertical thermodynamic profilers, as well as surface observing systems which all help to assess rapid changes in the pre-QLCS environment. In particular, the MIPS and RaDAPS include 915 MHz radar wind profilers, microwave profiling radiometers, and lidar ceilometers. For this particular event, the MoDLS included a Doppler wind lidar and balloon sounding capabilities. In addition, the Mobile Meteorological Mesonet Vehicle (M3V) collected surface weather information throughout the event while in transit and the M3V team launched balloon soundings from targeted locations. Lastly, several balloon sounding teams were out collecting data as well, collocated with RaDAPS, MoDLS, and MIPS, along with one team positioned in Falkville, AL. Two types of balloon sounding systems were employed during this event, with iMet4 radiosondes (International Met Systems, 2022) launched from the MIPS (5 total), RaDAPS (3 total), and Falkville, AL (4 total) locations while Windsond radiosondes (Sparv Embedded AB, 2015) were launched from the MoDLS location (3 total) and from M3V (3 total). Besides the plethora of instruments oper-
Figure 3.2: Map of the placement of UAH MAPNet facilities and fixed radars across north Alabama on 12 February 2020. Platforms labeled as “Mobile Profiler” feature several profiling instruments.

ated by UAH on this day, a 449 MHz radar wind profiler as well as a surface observing system operated by the NOAA Physical Sciences Laboratory was also collecting data at Courtland, AL (NOAA Physical Sciences Laboratory, 2022), collocated with the MoDLS system.

3.3 Surface Weather and Lightning Data

Evaluating the evolution of the QLCS as well as the rapid changes preceding it required examining a large number of weather stations across the Mississippi and Tennessee valleys. Automated Surface Observing System (ASOS; NOAA/NWS,
sites provide 1-min data resolution of temperature, dewpoint, pressure, wind, visibility, precipitation accumulation, and other relevant elements. However, as these systems are not found at every airport, some significant data gaps can exist (see Fig. 3.3). In order to supplement the ASOS network, particularly in southern Middle Tennessee where the nearest ASOS site with archived data northward from the Alabama border is at Nashville International Airport, personal weather stations within Weather Underground’s (Wunderground) personal weather station network (Weather Underground, 2022) were examined as well.

Lightning data were sourced from the Geostationary Lightning Mapper (GLM) onboard the latest Geostationary Operational Environmental Satellite – East (GOES-16). GLM is an optical sensor which detects subtle differences in background luminance. Due to the nature of this detection method, the GLM is at least 70% efficient in detecting lightning flashes, and may occasionally detect other luminance events such as meteors burning up in Earth’s atmosphere, especially at night (Goodman et al., 2012). Processed (level-2) lightning data were ordered and downloaded from the NOAA Comprehensive Large Array-data Stewardship System (CLASS; NOAA, 2022) under the GOES-R Series GLM L2+ Data Product (GRGLMPROD).

3.4 Methodology to Evaluate Mesoscale Variability

Given the broad region within which the pre-storm environment evolved, data from numerous surface observing stations were examined. One-minute ASOS data were read, quality-controlled, and plotted using the Python packages known as Pandas (The pandas development team, 2020; McKinney, 2010), Matplotlib (Hunter,
Figure 3.3: Locations of surface weather observing sites considered in this study. ASOS sites are located at select airports across the region, which results in fairly large spatial gaps. Personal weather stations which participate in Weather Underground’s personal weather station network were subsequently examined in southern Middle Tennessee to address the data void between Huntsville and Nashville.
2007), and Numpy (Harris et al., 2020). In addition to the fundamental state parameters such as temperature, dewpoint, pressure, and wind, derived variables such as potential temperature and its variants as well as different measures of moisture content were calculated. The more accurate formulation for potential temperature and equivalent potential temperature developed in Bolton (1980) were used while moisture content calculations such as vapor pressure, specific humidity, and mixing ratio followed equations described in Petty (2008).

Time series plots of both the one-minute ASOS and personal weather station data were examined to understand both the magnitude and rapidity of thermodynamic and kinematic changes observed at each site. More specifically, one-minute ASOS data were used to diagnose the origination of the cold pool which ultimately provided sufficient low-level forcing for the QLCS to develop. The magnitude of the cold pool within the QLCS was then tracked over time based on changes in the magnitude of temperature and moisture deficits observed as it tracked across the ASOS sites. Furthermore, observations of the magnitude of warming and moistening ahead of the QLCS provided insight into the importance of advection in conditioning the pre-storm environment over short time scales. Personal weather stations in southern Middle Tennessee were primarily used to understand how far north the warm sector and weakly-unstable air reached before the QLCS arrived.

3.5 Methodology to Examine QLCS Evolution

In addition to monitoring changes in the QLCS’s surface cold pool using surface observations, the system’s structure was also investigated using dual-polarization
radar data. GR2Analyst (Gibson Ridge Software, LLC, 2022) software was used for visualization of PPI scans as well as reconstructed cross-sections and volumetric cross sections using PPI scans. Aside from examining the overall depth of convection simply based on the height of reflectivity isosurfaces, microphysical signatures in the dual-polarization radar products helped provide insight into the strength of convective updrafts. For a review of the definitions and uses of base dual-polarization radar products, the reader is referred to Kumjian (2013a) and Kumjian (2013b). More specifically, a search for $Z_{dr}$ and $K_{dp}$ columns (Kumjian and Ryzhkov, 2008; Kumjian, Khain, et al., 2014) helped to glean whether convective updrafts were able to push above the freezing level and subsequently served as a proxy for updraft intensity (Kumjian, Ganson, et al., 2012; Kumjian, Khain, et al., 2014; Snyder et al., 2015). These, and other microphysical differences between the slabular structure at earlier times and the more cellular structure (James, Fritsch, et al., 2005) at later times helped to characterize the updraft/downdraft structure based on the types and heights of hydrometeors identified.

In addition to surface data and dual-polarization radar signatures, lightning data available from the GOES-16 GLM were used to help characterize the strength of updrafts over time as the charge separation required to produce lightning necessitates persistent collisions of ice hydrometeors in the presence of supercooled liquid water (Zipser and Lutz, 1994). GLM data downloaded from NOAA CLASS were read and plotted using the netCDF4 and Matplotlib packages written for Python. These data are categorized into events, groups, and flashes, each with their own definition. An event occurs when a single pixel exceeds the brightness threshold of the background
"noise" (during the daytime, this "noise" would be the brightness of cloud tops). A 2 ms integration window is used for finding events, so multiple optical pulses may occur within that time window in the same pixel, and thus one "event" is not necessarily one singular optical pulse. A group is defined when one or more pixels adjacent to an event also show a brightness exceeding the background noise threshold within the same 2 ms integration window. A flash is defined as a set of groups sequentially separated in time by no more than 330 ms and in space by no more than 16.5 km. These thresholds were chosen to produce results that correspond to the typical definition of a conventional lightning flash, though it is recognized that this definition has limitations. For more information, the reader is referred to (Goodman et al., 2012).

3.6 Methodology for Dual-Doppler Analysis

Prior to running dual-Doppler analyses, raw radar data collected by KHTX, the ARMOR, and the MAX radar were converted from their native file formats to CfRadial (UCAR Earth Observing Laboratory, 2022). ARMOR and MAX data were evaluated for the need to apply azimuth corrections by examining the location of point targets in comparison to the location of persistent, isolated, and strong returns on the reflectivity product. After evaluating the offset from seven independent point targets in three quadrants relative to ARMOR and from four independent point targets in three quadrants relative to MAX, azimuth corrections of 3° and 2.9° were applied to the ARMOR and MAX data, respectively. After the azimuth corrections, the data were converted into the Doppler Radar Exchange (DORADE; Lee et al., 2010) format, which splits each individual sweep comprising a volume scan into a separate
file. This was performed in order to manually edit the data using the National Center for Atmospheric Research’s solo3 software, which is contained within the larger Lidar Radar Open Software Environment (LROSE; Bell et al., 2020). Within solo3, manual filtering of noise, ground clutter, second-trip echoes, and bad rays was performed, in addition to the manual dealiasing of raw radial velocities. WSR-88D data in this case featured a Nyquist velocity between 25 and 33 m s$^{-1}$, typically requiring only one round of unfolding towards the edges of data collected at the higher tilts. Meanwhile, ARMOR and MAX data featured Nyquist velocities of 16 and 9.5 m s$^{-1}$, respectively, thus requiring two to as many as five rounds of dealiasing.

After correcting raw radar data, the DORADE files were aggregated and converted back to the CfRadial format, then gridded into a Cartesian coordinate system using PyART (Helmus and Collis, 2016). Grid specifications for the ARMOR-MAX dual-Doppler pair (18 km baseline) included an 80 km square centered on the ARMOR with 0.5 km grid point spacing in the horizontal while the vertical dimension ranged from 0 to 14 km with 0.25 km grid point spacing. For the ARMOR-KHTX dual-Doppler pair (70 km baseline), the grids featured a 200 km square centered on the ARMOR with 1 km horizontal grid point spacing while the vertical dimension ranged from 0 to 15 km with 0.25 km grid point spacing. The 0.25 km vertical grid point spacing was chosen in an effort to resolve shallow and/or fine-scale features potentially associated with severe weather production in the QLCS. However, given the 70 km baseline of the ARMOR-KHTX dual-Doppler pair, this degree of oversampling may produce fine-scale features which are not necessarily physical and could be artificial. After gridding, an advection correction was applied to each vertical level.
which shifted the data to the middle of the volume scan time based on a constant storm motion vector. This process assumed stationarity of the convection and sought to account for the rapid advection of precipitation observed in this case, which approached 30 m s\(^{-1}\) for some convective elements. It should be noted that researchers have tested a spatially-variable advection correction and found that smaller errors were produced when compared to using a constant motion vector (Shapiro, Willingham, \textit{et al.}, 2010a; Shapiro, Willingham, \textit{et al.}, 2010b; Shapiro, Gebauer, \textit{et al.}, 2021), so this is a potential source of error. However, the retrieved winds and vertical motion field did not appear to be adversely impacted by this method, at least qualitatively, so the authors accept the constant motion vector method as sufficient for the purposes of this research. After the “vertical” advection correction was applied, the gridded radar data were evaluated at 1 km to check how closely the leading edges of the QLCS lined up from each radar’s perspective. In some cases, radar data whose volume start times differed by as much as 2 min were utilized, again with the assumption of stationarity applied. In order to improve the dual-Doppler analyses produced by radar data which was offset in time, a horizontal advection correction was applied to the entire depth of one grid to align the leading edge at 1 km with that of the other grid. This would be another potential source of error since the assumption of the entire grid depth moving at the same rate in a strongly-sheared environment may not necessarily be true. However, for the purposes of this research, this method was deemed to be acceptable in order to maintain vertical continuity of the QLCS’s structure.
With the completion of all the pre-processing steps described above, the radar data was finally ready to be entered into a dual-Doppler analysis script. In this work, the Pythic Direct Data Assimilation (PyDDA; (Jackson et al., 2020; Shapiro, Potvin, et al., 2009; Potvin et al., 2012) package was employed, which uses a 3D variational framework. Significant advantages of this framework include the ability to add additional observations or model data to help improve the analysis result. This technique is also less sensitive to initial and boundary conditions compared to more traditional methods which vertically integrate the mass continuity equation either from the bottom or top of the model grids. PyDDA seeks to retrieve the 3D wind field \( V \) which minimizes the cost function \( J \) shown in Equation (3.1).

\[
J(v) = C_o J_o + C_{mass} J_{mass} + C_v J_v + C_r J_r + C_s J_s + C_{model} J_{model} + C_{point} J_{point} \tag{3.1}
\]

where subscript “o” represents the cost from observations, “mass” represents mass continuity, “v” represents vertical vorticity, “r” represents background winds from a rawinsonde, “s” represents smoothing, “model” represents model data, and “point” represents point observations from surface observing stations. More details about the process as well as its benefits and limitations are available, although the applications employed in past studies have not included QLCS events (Gao et al., 1999; Shapiro, Potvin, et al., 2009; Potvin et al., 2012; Jackson et al., 2020). Nonetheless, one key benefit of this method is the user can modify each \( C_n \), which is a weighting coefficient that is applied to each cost function component. This allows the user to adjust the importance of each component in the analysis, usually based on the uncertainty in
the data provided. For example, model data would likely receive a smaller weighting coefficient than radar observations.

The cited studies indicate that the best combination of cost function weighting coefficients requires experimentation and evaluation of the resulting upward motion field and retrieved winds. Although general guidelines exist, such as producing an analyzed upward motion field which yields values that are comparable to actual measurements or seeking cost function coefficients which yield relative magnitudes of each $C_n J_n$ that align with the relative importance of each component, there are no well-defined steps or published studies which have applied this method to a QLCS event in a high-shear, low-CAPE (HSCL) environment. In addition, with no source to directly verify the retrieved upward motion fields in this particular case, proxies such as consistency across time and space, the presence of lightning, and the presence of $Z_{dr}$ columns will serve as indicators to help verify the strength of upward motions retrieved using this analysis method. In addition, given that real data is used in this analysis and real data possesses limitations such as attenuation, sampling which may not fully include the entire storm depth, and noisy echoes or ground clutter which may mask true meteorological echoes during the filtering stage, artifacts will inevitably appear in the analysis (Shimizu, 2012). Every attempt has been made to produce dual-Doppler analyses which limit the presence of artifacts while producing realistic wind retrievals; however, this is yet another source of error to consider when interpreting the results of this research. Lastly, after the dual-Doppler analyses were completed, calculations for divergence and relative vorticity were carried out across
Table 3.2: Cost function weighting coefficients applied in each final dual-Doppler analysis based on radar pair.

<table>
<thead>
<tr>
<th>Radar Pair</th>
<th>$C_o$</th>
<th>$C_{mass}$</th>
<th>$C_s$</th>
<th>$C_v$</th>
<th>$C_{point}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>ARMOR-KHTX</td>
<td>1</td>
<td>100</td>
<td>$10^{-5}$</td>
<td>$10^3$</td>
<td>$10^{-3}$</td>
</tr>
<tr>
<td>ARMOR-MAX</td>
<td>1</td>
<td>100</td>
<td>$10^{-4}$</td>
<td>$10^{-4}$</td>
<td>$10^{-3}$</td>
</tr>
</tbody>
</table>

each grid to provide additional context for the investigation into the specific nature of the QLCS and associated elements which produced severe weather.

In total, dual-Doppler analyses were conducted for nine volume scans between KHTX and ARMOR, covering a time period of approximately 45 min while two volume scans between ARMOR and MAX were analyzed due to significant attenuation from both radars at all other times when the QLCS was located within the dual-Doppler lobes produced by these radars. For the purposes of this research, the northwestern dual-Doppler lobe produced by the ARMOR-KHTX pair and the northeastern lobe produced by the ARMOR-MAX pair will be examined. After much experimentation, the best cost function weighting coefficients which minimized artifacts and noise while retrieving reasonable estimates of the W field are shown in Table 3.2. HRRR model analysis data were tested as a constraint, but resulted in significantly poorer retrievals, likely due to the shallow nature of the convection and lack of association with larger-scale dynamic forcing mechanisms such as the 300 mb jet streak and the synoptic cold front.
CHAPTER 4

CASE OVERVIEW

The synoptic setup for this severe weather event featured a 300 mb longwave trough centered just east of the Rocky Mountains with the right-entrance region of a southwesterly 160-knot jet streak resulting in a broad region of upper-level divergence over the lower Mississippi and western Tennessee River valleys (see Fig. 4.1). At 500 mb, a shortwave trough ejected from west Texas on the morning of 12 February 2020 into the base of a larger 500 mb trough centered over the middle of the United States, resulting in enhanced forcing for ascent due to enhanced positive vorticity advection and strengthening vertical wind shear over the lower and mid-Mississippi Valley through the day. At 850 mb and 925 mb, warm air advection ramped up considerably by the end of the day, as evidenced in the official Nashville, TN soundings obtained by the National Weather Service for 12 UTC on February 12 and 00 UTC on February 13. These soundings revealed an increase in southwesterly 850 mb flow from 25 kts at 12 UTC to 70 kts at 00 UTC and an increase in 925 mb flow from easterly at 5 kts to southerly at 50 kts over the same time frame. In Huntsville, radiosondes launched from the Severe Weather Institute – Radar and Lightning Laboratories (SWIRLL) at UAH measured 925 mb flow of 21 kts at 15 UTC, which ramped up to
44 kts by 00 UTC, and 850 mb flow of 35 kts at 15 UTC, which increased to 68 kts by 00 UTC. At the surface, a cyclone initially formed along a stationary front in the northwestern Gulf of Mexico during the early morning hours of 12 February 2020. As the 500 mb shortwave tracked northeastward into northern Texas and Oklahoma, the surface low lifted northeastward and gradually strengthened, folding the stationary front into a new frontal system with a warm front extending east of the low and a cold front extending south of the low (see Figs. 4.2 and 4.3).
Figure 4.2: Surface analysis at 12 UTC on 12 February 2020. Note the weak low pressure center near Houston, TX with a synoptic warm front extending eastward along the northern Gulf Coast. Image source: NOAA/NWS/WPC.
Figure 4.3: Surface analysis at 00 UTC on 13 February 2020. Note the 1006 mb low pressure center near the AR/LA/MS border intersection with a synoptic warm front extending generally eastward across the northern Gulf Coast states. In addition, a squall line was analyzed from Middle Tennessee through northern Alabama into central Mississippi, which is the severe quasi-linear convective system (QLCS) of interest in this case study. Image source: NOAA/NWS/WPC.
Upper-level divergence and subsequent ascent in the right-entrance region of the 300 mb jet streak resulted in a large area of stratiform precipitation in northeast Texas, Oklahoma, and Arkansas within a cool and stable near-surface environment early in the day on February 12. Low-level convergence focused along a surface trough extending north from the weak low in the northwestern Gulf of Mexico resulted in the formation of a narrow north-south-oriented band of showers stretching from far eastern Texas through northwestern Louisiana into south-central Arkansas between 13-15 UTC. Through evaporative cooling, a cold pool developed beneath these showers and shifted northeastward with the mean low-level flow (quantification of the cold pool strength is described in section 5.1). As a second surface trough and its associated showers caught up with the first, the aforementioned cold pool advanced eastward and northeastward into a region of decreasing static stability near and behind the northward-advancing warm front. This resulted in the formation of a narrow low-topped line of convective showers along its leading edge between 19-20 UTC, extending from southwest Tennessee through south-central Louisiana. This convective line existed within two regimes: one north of the warm front, where a continuous band of reflectivity exceeding 40 dBZ appeared more reminiscent of a narrow cold-frontal rainband with slabular ascent (James, Fritsch, et al., 2005), and a second regime south of the warm front, where decreasing static stability yielded a broken line with numerous distinct convective elements largely embedded within a continuous band of stratiform rain (see Fig. 4.4).

Surface frontogenesis rapidly increased (see Fig. 4.5) as convective overturning within this line of showers reinforced the cold pool and new cells continued to develop
Figure 4.4: Base reflectivity PPI images at 2009 UTC on 12 February 2020 from KNQA (Memphis, TN) at (a) 0.5\(^\circ\) elevation and (b) KDGX (Jackson, MS) at 0.3\(^\circ\) elevation. Yellow text denotes county names while white text depicts towns/cities. Blue lines indicate interstates while orange lines indicate state highways. Thin green lines are county boundaries. The eastern band of reflectivity $> 40$ dBZ seen from KNQA radar corresponds to the early stages of the QLCS of interest north of the warm front. The broken band of reflectivity $> 40$ dBZ which tapers to a narrow band of reflectivity $\geq 35$ dBZ near Greenwood, MS seen from KDGX radar corresponds to the same QLCS mainly south of the warm front. Image source: GR2Analyst
along the gust front amid strengthening warm air advection. Through sustained upward motion and latent heat release in the lower to mid-troposphere attendant to the convective line, the surface trough deepened such that its surface pressure minimum and horizontal temperature gradient magnitude soon exceeded that of the original cold front. This development of a “secondary” cold front preceding the primary synoptic cold front bears similarities to numerical simulations carried out by Hoskins et al. (1984). Although the perturbation introduced into the pre-frontal environment in the aforementioned study was a surface-based warm anomaly, the development of a secondary cold front which takes on characteristics similar to, or of greater magnitude than, that of the synoptic cold front appears to match the effects observed in the fourth case described in their section 4. However, since vertical wind profiler observations (not shown) identified distinct southerly low-level jets preceding both the surface trough and the synoptic cold front, it is also possible that frontogenesis via horizontal shearing as described in section 3f of Schultz (2005) may have also contributed to the development of this secondary cold front and its associated QLCS. Forward propagation of the gust front aided by downward mixing of very strong low-level flow in convective downdrafts likely caused this pre-frontal trough, and the developing QLCS, to advance increasingly away from the synoptic cold front.

The QLCS gradually strengthened as it moved across Mississippi and developed a line-echo wave pattern (LEWP; Nolen, 1959) structure as increasingly unstable air advected northward in tandem with the northward-advancing warm front. A MV and several bow echo segments became evident on PPI images from the KGWX radar as the QLCS neared and crossed into northwestern Alabama. At 2229 UTC, severe
Figure 4.5: Surface frontogenesis analysis at 21 UTC on 12 February 2020 from the SPC mesoanalysis. A region of intense surface frontogenesis stretches from central Louisiana through western Mississippi into western Tennessee which corresponds to the early stages of the QLCS of interest for this case study. Image source: NOAA/NWS/SPC mesoanalysis archive.
weather reports first occurred in far northwest Alabama with trees and power lines downed by strong winds associated with the MV. This subsequently produced another fourteen reports of severe weather, including a 76 mph wind gust and two weak (≤ EF-2) tornadoes over the next thirty minutes in far northwestern Alabama before any severe weather warnings were issued by the National Weather Service in Huntsville, AL. Meanwhile, to the south of this MV, a trio of bow echo segments rippled up the line and produced initially sporadic wind damage reports. The MV produced multiple swaths of extensive wind damage, particularly when the northernmost bow echo merged with the southern flank of the MV between 00-01 UTC, likely due to the close proximity of dynamically-lowered perturbation pressure in the center of the MV to the mesohigh associated with the convection just south of the MV, leveraging the mechanisms described previously (Trapp and Weisman, 2003; Wakimoto, Murphey, Davis, et al., 2006; Atkins and St. Laurent, 2009a). After this period of widespread damaging winds, the MV continued to produce sporadic severe wind damage at times while it tracked into east-central Tennessee and encountered increasingly stable air. As the bow echo segments tracked into southern Middle Tennessee around 00 UTC, the QLCS began to transition back into a more slabular convective line with a nearly-continuous and linear band of reflectivity greater than 40 dBZ. With this transition came a significant reduction in the severe threat as only one report of wind damage was observed in northeastern Alabama (see Fig. 4.6).

Lightning data collected by the GOES-16 GLM (see Fig. 4.7) reveals that despite the low-CAPE conditions, especially with northward extent, a considerable amount of lightning still occurred, supporting the idea that convective updrafts were
Figure 4.6: Map of scanning radar locations, severe wind and tornado report locations, as well as manually-determined bow echo tracks and the MV track. Storm reports data source: NOAA/NCEI.

able to reach well-above the freezing level and bring ice hydrometeors into contact for periods long enough to promote charge separation, particularly when severe weather production was ongoing over northern Alabama and southern Middle Tennessee. In the next chapter, detailed results will be presented with section 5.1 focusing on the development of the QLCS, section 5.2 delving into the evolution of the QLCS based on microphysical signatures observable by dual-polarized radars, section 5.3 exploring dual-Doppler analyses of the MV and severe bow echoes, section 5.4 quantifying rapid changes in the pre-storm environment as well as the QLCS’s cold pool strength, and section 5.5 exploring the operational forecasters’ thought processes leading up to this event.
Figure 4.7: GOES-16 GLM lightning events, groups, and flashes overlaid with the manually-determined MV track (based on the location of a weak echo hole) and bow echo tracks. A distinctly higher concentration of lightning flashes is clearly associated with the bow echoes, indicating stronger and deeper updrafts. GOES-16 GLM data source: NOAA CLASS.
CHAPTER 5

RESULTS

This section delves into a variety of results stemming from observational investigations into the quasi-linear convective system (QLCS) development and evolution, as well as the kinematics of the severe-weather-producing elements. Section 5.1 details the QLCS origin, section 5.2 follows the QLCS evolution on an approximately one-hour basis, section 5.3 covers dual-Doppler analysis results, section 5.4 examines mesoscale variability at stationary sites, and section 5.5 provides insight into the predictability of this event based on forecast discussions issued by several National Weather Service offices impacted by or adjacent to the severe threat on 12 February 2020.

5.1 QLCS Development

The severe QLCS of interest tracked across the Tennessee Valley during the afternoon and evening hours of 12 February 2020. However, its origination can be traced back to a slow-moving band of showers which developed around 14 UTC within a low-level convergence zone attendant to a surface trough extending from northwest Louisiana through far southeast Texas (see Fig. 5.1). This surface trough was sup-
ported by the fringes of weak upper-level divergence associated with the right-entrance region of a 300 mb jet streak and was only producing scattered showers along it. Evaporative cooling beneath these persistent showers caused an initially weak cold pool to develop (see Fig. 5.2). As warm air advection strengthened with the slow northward advancement of a warm front and upper-level divergence associated with the 300 mb jet streak gradually lifted northward, the band of showers lengthened towards the northeast. The cold pool subsequently grew in the along-line dimension while a second cold pool associated with another surface trough to the west caught up with the first. The forward motion of the first cold pool increased, and starting around 19 UTC, a narrow line of showers began developing along the leading edge of the first cold pool. By 20 UTC, this initially broken line had become a nearly-continuous band of convective showers extending from southwest Tennessee through western Mississippi into central Louisiana (see Fig. 5.3). As a result of the rapid linear organization of the convection, downdrafts and subsequent evaporative cooling produced a narrow but intense cold pool (see Fig. 5.2) which quickly tightened the temperature gradient behind the surface trough and essentially produced a relatively shallow mesoscale cold front which lied ahead of the synoptic cold front and extended well north of the synoptic warm front. The frontogenesis process associated with inhomogeneities in prefrontal air described by Schultz (2005) is proposed as the most likely mechanism behind the formation of this pre-frontal trough, as it consistently remained ahead of any notable upper-level divergence associated with the 300 mb jet streak. The thermal gradient produced by the convection was further enhanced by rapid advection of warm, moist air from the Gulf of Mexico, which is one of the
key factors in the production of severe weather by this QLCS later in its life. This line of convection would ultimately become a severe QLCS upon entering northwest Alabama. The evolution of this band of showers into the severe QLCS is explored in the next section.

5.2 Evolution of QLCS from a Microphysical Perspective using Dual-Polarized Radars

Early in its life, the northern portion of the QLCS was distinctly slabular in nature (James, Fritsch, et al., 2005), featuring a contiguous band of radar reflectivity ($Z_h > 35$ dBZ at 1 km above radar level (ARL) that was over 100 km in length and up to 8 km wide, as seen in the northern half of figure 5.4. Embedded areas of $Z_h > 50$ dBZ likely represented locally enhanced updrafts which augmented collision-coalescence processes at low levels in the line, as values of $Z_{dr}$ and $K_{dp}$ were also elevated in these areas relative to the rest of the line. Further south, the QLCS featured a more cellular structure marked by a rougher leading edge, much greater variation in the across-line dimension of the convective region ($Z_h > 35$ dBZ), and local enhancements in $Z_h$ with magnitudes similar to or greater than that observed in the slabular portion. At 21 UTC, the Weather Prediction Center’s surface analysis (Fig. 5.5) included a squall line over northern Mississippi bisected by a warm front, which indicates that this QLCS existed within two separate regimes demarcated by the warm front’s position. As the QLCS approached the western Alabama border, overall $Z_h$ values increased at low levels in the slabular portion of the line, with many
Figure 5.1: Radar reflectivity at S-band from the WSR-88D located at Shreveport, LA (KSHV) at 1433 UTC on 12 February 2020. A broken band of showers extended from far eastern Texas through northwestern Louisiana into southwestern Arkansas. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
Figure 5.2: Time series plots of $\theta_e$ derived from 1-min ASOS sites located near (a) El Dorado, AR (KELD); (b) Monroe, LA (KMLU); and (c) Greenville, MS (KGLH) covering the time frame 14 UTC through 21 UTC on 12 February 2020. Around 1445 UTC, a minor 1 K drop was observed as the band of showers shown in Fig. 5.1 moved into El Dorado, AR. As the band of showers filled in, the precipitation area to the west caught up to this band, and evaporative cooling continued, a more robust cold pool developed. Around 1815 UTC, a much more substantial drop in $\theta_e$ of 18 K occurred at Monroe, LA, albeit over an hour-long period. Just after 1915 UTC at Greenville, MS, an even larger and much sharper drop in $\theta_e$ was observed as a narrow band of convection developed along the leading edge of the cold pool. This narrow band of convection (see Fig. 5.3) ultimately became the severe QLCS that is the subject of this research.
Figure 5.3: Radar reflectivity at S-band at lowest tilts from WSR-88Ds located at (a) Memphis, TN (KNQA) and (b) Jackson, MS (KDGX) around 2009 UTC showing the initial development of a narrow band of convection along the leading edge of a cold pool. This narrow band of showers ultimately became the severe QLCS of interest. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
local enhancements into the 50-60 dBZ range becoming apparent below 2 km ARL from the lower tilts of the KGWX radar.

Severe weather began as the QLCS pushed into far northwestern Alabama in association with the early stages of a long-lived mesovortex (MV). This MV and a trio of closely-spaced bow echo segments embedded within the slabular-to-cellular transition zone south of the MV all produced instances of wind damage as the QLCS moved across northern Alabama and southern Middle Tennessee. Severe weather primarily occurred near and south of the MV, which shows that the most favorable conditions for high-shear, low-CAPE (HSLC) severe weather were restricted to the transition zone between the slabular and cellular modes. South of these features, a few isolated instances of severe wind occurred within the distinctly cellular portion, but this was mainly associated with one bow echo. The cellular portion of the QLCS appeared to gradually advance northward along the line, marked by distinct breaks in the $Z_h$ field along with numerous local peaks in reflectivity at or above 60 dBZ translating northward while the QLCS as a whole slowly trekked eastward. Interestingly, as the cellular portion of the QLCS pushed through north central Alabama and into southern Middle Tennessee, the slabular mode appeared to return from south to north with one contiguous region of $Z_h$ in the 40-50 dBZ range and locally higher values mostly in the 50-56 dBZ range tracking across northeast Alabama. Exploration of the microphysical changes this QLCS underwent throughout its life will help to characterize changes in the strength of the system over time. An observational analysis of the evolution of the slabular and cellular convective regions follows below on an hourly basis, using dual-polarization radar data available from several WSR-88Ds located...
Figure 5.4: 1 km ARL reflectivity at 2042 UTC on 12 February 2020 composited from three WSR-88Ds (KDGX, KGWX, KNQA) and gridded onto a Cartesian grid with 1 km horizontal and 0.25 km vertical grid spacing. The colormap ticks were chosen to emphasize the incipient QLCS, which is located at the leading edge of the broader swath of radar returns encompassing the top-left half of this figure. The leading convective line is forced by a cold pool which provided strong low-level forcing for ascent. Data source: NOAA/NCEI NEXRAD Level II archive.
Figure 5.5: 21 UTC surface analysis from 12 February 2020 produced by the Weather Prediction Center. The analyst identified the incipient QLCS as a squall line across Mississippi. The squall line was bisected by a warm front in this analysis and this will remain the case throughout the life cycle of this QLCS. Image source: NOAA/NWS/WPC.
across the southeastern United States, as well as the ARMOR located at Huntsville International Airport.

5.2.1 Slabular Structure Analysis

At 2029 UTC, the incipient QLCS was located east of Memphis, TN and featured a distinctly slabular appearance. PPI imagery (Fig. 5.6) showed widely-spaced areas of locally-enhanced $Z_h, Z_{dr}$, and $K_{dp}$ relative to the majority of the line. A vertical cross-section reconstructed from PPI scans through one of these enhanced areas (not shown) reveals local maxima of $Z_{dr}$ in the 1-2 dB range and $K_{dp}$ in the 1-2.3° km$^{-1}$ range below the melting level, which implies larger mean drop diameters and a higher concentration of drops relative to the rest of the QLCS, respectively. With the HRRR analyzed sounding from Memphis, TN at 20 UTC (not shown) indicating an environmental $0^\circ$C level around 3.5 km above ground level (AGL) and the vertical cross-section depicting virtually all radar returns below 4 km ARL, it is clear that there were no $Z_{dr}$ or $K_{dp}$ columns (Kumjian and Ryzhkov, 2008; Kumjian, Khain, et al., 2014) present at this time. Nevertheless, areas of larger mean drop diameters and higher concentrations of drops imply that locally stronger updrafts were present, which would enhance condensational growth of droplets and thereby contribute to faster growth by collection.

Gradual strengthening of the QLCS occurred through 2130 UTC, as suggested by an overall increase in $Z_h, Z_{dr}$, and $K_{dp}$ values seen by the KGWX radar (see Fig. 5.7). At 21 UTC, the HRRR analyzed sounding for Tupelo, MS approximated the freezing level at around 3.7 km AGL. Vertical cross sections (not shown) indicated
Figure 5.6: 0.5° PPI imagery from KNQA at 2029 UTC on 12 February 2020. Base radar products are shown, including (a) reflectivity, (b) differential reflectivity, (c) correlation coefficient, and (d) specific differential phase. White letters are town or city names while yellow letters are county names. From the PPI perspective, this slice through the system ranges from around 0.5 km ARL near the TN/MS border to around 0.9 km near the bottom of each panel, assuming standard refraction. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
that a vast majority of radar returns were below 4 km AGL with no distinct columns of enhanced $Z_{dr}$ or $K_{dp}$ above the freezing level. Several sections of the line exhibited local maxima in $Z_h$ of 60-65 dBZ, which typically corresponded to $Z_{dr} > 3$ dB and $K_{dp} > 3^{°}$ km$^{-1}$ at a height of around 1.1-1.4 km ARL. In some cases, these values overlapped with CC $\geq 0.97$, suggesting a relatively high concentration of large raindrops.

By 2230 UTC, overall $Z_h$, $Z_{dr}$, and $K_{dp}$ values had increased throughout the line, and severe wind production had begun near Waterloo, AL. This was associated with the early stages of a mesovortex (MV) which would dominate as the primary severe weather producer throughout this event. Figure 5.8 depicts a portion of the
QLCS at 2230 UTC from the KGWX radar, with $Z_h > 50$ dBZ common throughout the line along with many zones where $Z_{dr} > 3$ dB corresponds to $K_{dp}$ values in the 3-4° km$^{-1}$ range and values of $CC > 0.96$, indicating that mainly large, oblate raindrops were present in these areas. The HRRR analyzed sounding from 22 UTC for Tupelo, MS indicated that the freezing level was just below 4 km. A vertical cross-section through the line created using a reconstruction of individual PPI scans (Fig. 5.9) showed a noticeably taller system with $Z_h > 30$ dBZ reaching near 5 km ARL and reduced CC around the freezing level. A proposed $Z_{dr}$ column is circled in the cross-sections, where a contiguous zone of $Z_{dr}$ in the 1.75-2 dB range reaches up to around 6 km ARL overlaid with $CC > 0.9$, and positive $K_{dp}$ which decreases to near 0 above the freezing level. Also note the considerable reflectivity overhang which suggests strong updrafts were sustaining large raindrops and possibly small hail at this time.

Moving forward to 2330 UTC, the QLCS had produced fifteen reports of severe winds and/or wind damage by this point in time, including a 76 mph wind gust and two weak (≤ EF-1) tornadoes, without any severe weather warnings in effect. The slabular reflectivity appearance continued across Tennessee and into northwest Alabama with a continuous band of $Z_h > 35$ dBZ stretching from the TN/KY border southward to near Phil Campbell, AL. Enhancements of $Z_h$ into the 50-55 dBZ range were observed throughout the QLCS across Tennessee from the lowest tilt on the Nashville, TN radar (not shown). In northwest Alabama and southern Middle Tennessee, a nearly-continuous band of $Z_h > 50$ dBZ was visible from the lowest tilts of the KHTX and KGWX radars extending north from Phil Campbell, AL through
Figure 5.8: 0.5° PPI imagery from KGWX at 2229 UTC on 12 February 2020. Products shown and map labels as in figure 5.6. From the PPI perspective, this slice through the system ranges from around 2.4 km ARL north of the TN/MS border to around 1 km in central Pontotoc County, assuming standard refraction. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
Figure 5.9: Reconstructed cross-section from KGWX radar at 2230 UTC on 12 February 2020 perpendicular to the QLCS. Products shown as in Fig. 5.6. This cross-section is taken at approximately 64 km from 3° azimuth, resulting in the cross-section bottom starting around 0.8 km ARL based on standard refraction. The white ellipse corresponds to the proposed location of a $Z_{dr}$ column. Note that approximately 3 min elapsed between the lowest elevation angle and the top of the echo overhang region near 5 km altitude. Therefore, the extent of the overhang region is exaggerated somewhat by the forward motion of the QLCS. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
Figure 5.10: Base reflectivity PPI images at 2330 UTC on 12 February 2020 from the lowest elevation angles of (a) KHTX and (b) KGWX radars with three bow echoes and a long-lived mesovortex annotated. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.

Spring Hill, TN. Three bow echoes as well as the MV were apparent in this zone of elevated $Z_h$ (see Fig. 5.10).

The official 00 UTC sounding retrieved from Nashville, TN (launched at 23 UTC) found a freezing level near 3.5 km AGL while the sounding launched from the Severe Weather Institute - Radar and Lightning Laboratories (SWIRLL) at UAH at 23 UTC found a freezing level around 3.8-4.0 km AGL, exemplifying the north-south gradient in temperatures observed across the warm front that was generally located around the TN/AL border at this time. Vertical cross-sections taken around 2330 UTC from the KOHX radar reconstructed using PPI scans indicated an increased vertical depth to the system with highly-correlated ($CC \geq 0.97$) $Z_h$ in the 10-20 dBZ
range reaching up to around 8 km ARL while $Z_h \geq 30$ dBZ reached up to around 5 km ARL. Further south, the KHTX radar observed an important distinction in system depth, with a shallower echo top (similar to that observed from KOHX) north of the MV compared to a deeper system with more convective vigor to the south of the MV. For example, a reconstructed cross-section taken through the QLCS just south of the MV near the AL/TN border south of Minor Hill, TN (see Fig. 5.11) showed $Z_h$ up to 47 dBZ reaching into the 6-7 km AGL layer while $Z_h > 50$ dBZ reached up to around 5 km AGL. Within these regions of enhanced $Z_h$ near and above the freezing level, $Z_{\text{dr}}$ of 1-2.25 dB, $K_{\text{dp}}$ of 1-4.4° km\(^{-1}\), and CC generally above 0.97 suggest that both $Z_{\text{dr}}$ and $K_{\text{dp}}$ columns (Kumjian and Ryzhkov, 2008; Kumjian, Khain, et al., 2014) were present, indicating that stronger updrafts were lofting large liquid hydrometeors well-above the environmental freezing level and these updrafts were sustaining relatively high concentrations of liquid hydrometeors above the freezing level (Kumjian, Ganson, et al., 2012; Kumjian, 2013b). As a result, precipitation loading as well as evaporative cooling and melting likely contributed to locally stronger downdrafts in the convective precipitation region, yielding the sporadic wind damage reports observed outside of the MV. Reconstructed cross-sections through the three bow echoes (not shown) depict a similar pattern of enhanced positive $Z_{\text{dr}}$ and $K_{\text{dp}}$ values reaching similar heights ARL and above the freezing level as that observed just south of the MV, further supporting the idea that an environment favoring stronger updrafts existed south of the MV.

A majority of severe weather occurred within the transition zone between the relatively shallow slabular portion of the QLCS and the relatively deeper cellular
Figure 5.11: Reconstructed cross-section from KHTX radar at 2331 UTC on 12 February 2020 taken perpendicular to the QLCS just south of Minor Hill, TN. Products shown as in Fig. 5.6. This cross-section is taken at approximately 107 km from 275° azimuth, resulting in the cross-section bottom starting around 1.3 km ARL based on standard refraction. The white ellipse corresponds to the proposed location of $Z_{dr}$ and $K_{dp}$ columns. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.
portion of the QLCS. It is within this zone that meso-\(\gamma\)-scale structures that are well-known to produce severe weather (bow echoes and mesovortices) were concentrated. An intriguing finding is that the trio of bow echoes can be traced back an hour and half and featured a depth of up to around 5 km ARL, as viewed from the KGWX radar at 2202 UTC, but they did not produce severe weather until moving into northern Alabama. This yields the question, what microphysical differences existed in the generally sub-severe cellular portion of the QLCS south of the bow echoes?

### 5.2.2 Cellular Structure Analysis

From the KGWX radar at 2230 UTC on 12 February 2020, the QLCS appearance transitions from a generally smooth, narrow, and continuous band of \(Z_h > 35\) dBZ with subtle LEWP structures (Nolen, 1959) to a more ragged structure reminiscent of lobe and cleft instability (Markowski and Richardson, 2010) starting near Fulton, MS (see Fig. 5.8). Interestingly, while the general orientation of the QLCS became increasingly parallel to the 0–6 km environmental shear vector with southward extent, individual bow echoes tended to orient themselves with their apex nearly orthogonal to the deep-layer shear vector, and individual cells tended to remain separated from their neighbors. James, Markowski, et al. (2006) examined the impact of water vapor content on QLCSs and the development of bow echoes. Results from the study suggest that the transition in structure observed in this case may be due to a change in the ambient water vapor content across the warm front, with lower water vapor content to the north supporting stronger cold pools which yield greater low-level forcing at the cold pool edge and greater water vapor content to
the south supporting weaker cold pools which yield weaker low-level forcing. Under this hypothesis, the severe bow echoes existed in a zone of intermediate water vapor content, where only some downdrafts became enhanced by evaporative cooling and caused the QLCS to tilt upshear locally, thus producing small rear inflow jets.

At 2330 UTC from the KGWX radar’s perspective, the QLCS featured a cellular reflectivity structure south of Phil Campbell, AL, with multiple bow echoes and convective cells having distinct separation ($Z_h < 35$ dBZ between $Z_h$ maxima) along one boundary. The HRRR analyzed sounding from 23 UTC at Columbus, MS showed a freezing level near 4 km AGL. Reconstructed vertical cross-sections through the cells (not shown) revealed a general maximum height in the 40 dBZ isosurface at around 5 km ARL, with 10-20 dBZ reflectivity reaching into the 6-7 km ARL layer. The most intense portion of the QLCS in this cellular region was located within a bow echo just south of Hamilton, AL, where the 40 dBZ isosurface reached up to around 8 km ARL and 56 dBZ reflectivity reached up to around 6 km ARL. In this area, $CC > 0.98$, $Z_{dr}$ varying around 0 dB, and $K_{dp}$ also varying around $0^\circ$ km$^{-1}$ all suggest graupel and/or small hail were present above the freezing level. $Z_{dr} > 3$ dB was primarily found at or below 5 km ARL in this cellular portion of the QLCS, indicating that mixed rain and melting hail/graupel were contributing to large $Z_{dr}$ below the freezing level. Thus, in addition to evaporative cooling, melting hydrometeors and precipitation drag likely contributed to locally stronger downdrafts in this area, especially considering that reports of tree and power line damage occurred within the 2325-2330 UTC time frame in association with this bow echo. Given the
freezing level near 4 km AGL, it is proposed that convective updrafts were able to sustain relatively shallow $Z_{dr}$ columns at this time.

By 0004 UTC on 13 February 2020, the QLCS was actively producing severe weather in northern Alabama and southern Middle Tennessee. The 00 UTC HRRR analyzed sounding from Huntsville International Airport and the 00 UTC radiosonde launched from SWIRLL both showed a freezing level near 3.6 km AGL. From the ARMOR, a vertical cross-section taken through one of the small bow echoes (see Figs. 5.12 and 5.13) south of the MV showed a strong updraft with an accompanying bounded weak echo region (BWER; Kumjian, 2013b) reaching up to 5 km ARL. Enhanced $Z_{dr}$ of around 3.5 dB reached just above 5 km ARL and values in the range of 1-2 dB were observed up to the top of the scan around 6.8 km ARL, which is proposed to be a $Z_{dr}$ column. Generally high CC values greater than 0.93 were found in the $Z_{dr}$ column, and given the C-band wavelength of ARMOR, large oblate raindrops were likely present at this time. Given the presence of a BWER, precipitation loading in combination with evaporative cooling and melting hydrometeors likely contributed to locally stronger downdrafts in the system, yielding the sporadic areas of wind damage observed outside of the MV.

At 0003 UTC, a long-track MV that originated in far northwest Alabama was moving into far southern Middle Tennessee. As viewed from KHTX, an arc-like feature of enhanced $Z_{dr}$ values was observed wrapping around the MV (see Fig. 5.14). This was most prominent on the lowest tilt and surrounded a weak echo hole (WEH) which approximately denoted the center of circulation associated with the MV. Given the values of $Z_{dr}$ in the 2-4 dB range, large oblate raindrops were likely embedded in
Figure 5.12: 1.3° PPI imagery from ARMOR at 0004 UTC on 13 February 2020. Products shown and map labels as in figure 5.6. A white line depicting the location of the vertical cross section for Fig. 5.13 is shown on all panels. From the PPI perspective, this horizontal slice through the system ranges from around 0.25 km ARL near the bottom of each panel to around 0.8 km near the top of each panel, assuming standard refraction. A much less homogeneous QLCS is observed at this time in comparison to the early stages of the QLCS. Image source: GR2Analyst. Data source: UAH ARMOR.
Figure 5.13: Reconstructed cross-section from the ARMOR at 0004 UTC on 13 February 2020 perpendicular to the QLCS through a small bow echo rippling along the QLCS during its cellular phase. Products shown as in Fig. 5.6. This cross-section is taken at approximately 35 km from 342° azimuth, so the bottom of the cross-section starts around 0.80 km ARL based on standard refraction. The white ellipse corresponds to the proposed location of a $Z_{dr}$ column. Image source: GR2Analyst. Data source: UAH ARMOR.
Figure 5.14: 0.4° PPI imagery from KHTX at 0000 UTC on 13 February 2020. Products shown and map labels as in figure 5.6. This slice through the system ranges from around 0.85 km ARL near the bottom of each panel to around 1 km ARL near the top of each panel, assuming standard refraction. An arc of enhanced $Z_{dr}$ values is seen wrapping around the MV with a local minimum visible in the center of the MV. Image source: GR2Analyst. Data source: NOAA/NCEI NEXRAD Level II archive.

The circulation and centrifuged outwards from the center (Dowell et al., 2005; Kumjian and Ryzhkov, 2008). This arc of high $Z_{dr}$ values appeared to cycle over time from a line of enhanced values along the leading edge of the QLCS to wrapping around the MV. However, this feature will not receive further attention in this research for brevity.

5.3 Dual-Doppler Analyses

Dual-Doppler analyses for nine volume scans between 2330 UTC and 0015 UTC were produced from the KHTX-ARMOR pair of radars due to the favorable
orientation of the dual-Doppler lobes relative to the bow echoes and MV which produced severe weather (see Fig. 5.15). As a reminder, given that the ARMOR and KHTX scan strategies were not time-synchronized, some dual-Doppler analyses were produced from volume scans whose start times differed by as much as 2 min, so this may impact the accuracy of the analyses and will be noted where applicable. Otherwise, the ARMOR-MAX dual-Doppler configuration resulted in two acceptable volume scans for analysis between 0015 UTC and 0020 UTC, although significant attenuation from both radars due to heavy rain precludes the accuracy of each at increasing range. This section will be organized by QLCS element with subsections for the MV and bow echo segments, as well as the for the overall QLCS structure.
5.3.1 Overall QLCS Structure

The QLCS in this case was embedded within a weakly unstable environment characterized by MLCAPE measured at a maximum of 203 J kg\(^{-1}\) while 0-1 km wind shear magnitudes were measured at around 25 m s\(^{-1}\). These extreme HSIC conditions yielded a relatively shallow QLCS whose updraft magnitudes generally peaked at around 3-4 km above ARMOR with typical max values of around 10-12 m s\(^{-1}\) south of the MV and around 8-10 m s\(^{-1}\) north of it. These findings verify the proposed differences in updraft character noted in the previous section, namely that convective updrafts were stronger (weaker) south (north) of the MV. The updrafts south of the MV were also more narrow and focused near the leading edge of the QLCS (see Fig. 5.16), where the system’s cold pool forced the weakly unstable air upward, exhausting virtually all convective instability in the process. To the north of the MV, updrafts were considerably broader and somewhat weaker than those to the south, perhaps due to weaker instability or weaker forcing for ascent along the QLCS’s leading edge, or both. This distinct difference in updraft area as well as the horizontal updraft gradient persisted throughout all 45 min of analyses.

It is recognized that the individual point soundings launched ahead of the QLCS may not necessarily represent the true inflow air, particularly when rapid advection can lead to CAPE rises up until the QLCS gust front arrives (King et al., 2017). As a result, it is possible that CAPE values in the true pre-convective inflow air may have been closer to 300 J kg\(^{-1}\) or more. Nevertheless, when considering that updraft magnitudes typically reach about half of their theoretical maximums (based
Figure 5.16: 3 km CAPPI imagery at 2346 UTC with reflectivity factor from KHTX as the background color fill for both plots and colored contours for (a) vertical vorticity and (b) vertical velocity (W) with upward (red) and downward (blue) motion plotted. Ground-relative wind vectors are shown as well. The vertical vorticity field denotes where the wind shift occurs, and the vertical velocity field clearly shows the location of convective updrafts near the leading edge of the QLCS.
on $w_{\text{max}} = \sqrt{2\text{CAPE}}$; Markowski and Richardson, 2010) due to a variety of limiting factors, a CAPE range of 200-300 J kg$^{-1}$ would yield an expected maximum updraft magnitude range of around 10-12 m s$^{-1}$, which is retrieved in these dual-Doppler analyses. In addition, when considering that freezing level heights (see Table 5.3 in section 5.4) were around 3.6 km AGL just ahead of the QLCS, updrafts maximizing at around 3-4 km and extending well above the freezing level would support the persistent collision of ice hydrometeors in the presence of supercooled liquid water, thus promoting the production of lightning (Zipser and Lutz, 1994), which was observed in this case from the GOES-16 GLM instrument (see Fig. 5.17).

Over time, the relative locations, heights of maxima, and area of the updrafts south of the MV remains remarkably consistent, thus boosting the authors’ confidence
in the accuracy of the retrieved wind fields. Starting with the 0008 UTC dual-Doppler analysis, attenuation resulting from the ARMOR’s viewing angle becoming increasingly parallel to the QLCS generally degrades the results; however, the placement and strength of the updrafts outside of the zone of attenuation is generally maintained (see Fig. 5.18).

5.3.2 Severe Wind Production in the Mesovortex

The area covered by the northwestern ARMOR-KHTX dual-Doppler lobe allows for an investigation into the long-track MV which was responsible for the most concentrated and most significant (reported) damage produced by this QLCS. However, it would be negligent to not recognize the limitations of local storm reports, which depend on a variety of human factors that influence the accuracy and depiction of severe weather (Trapp, Wheatley, et al., 2006). Given the length of the track of this MV, not all periods of severe wind production were captured. Furthermore, given that the location of the MV when it first developed was well-outside the northwest lobe, the formation mechanism will not be analyzed in this work. Primary attention will be given to the kinematic structure of the MV near the times of severe weather production which occurred well-within the northwestern lobe, including the vertical vorticity, storm-relative motion, and updraft-downdraft magnitudes and heights.

The earliest analysis time is around 2331 UTC, which is just a few minutes following a surveyed EF-0 tornado that occurred near the town of Five Points, TN. From the single-Doppler perspective of KHTX, a notable increase in ground-relative wind speed was observed on the lowest tilt (0.5°) at around 112 km from the radar,
Figure 5.18: 1km CAPPI imagery around 0015 UTC on 13 February 2020 with reflectivity from KHTX as the background color fill along with ground-relative motion vectors and vertical velocity (W) in color contours. (a) Dual-Doppler analysis without masking data where ARMOR has attenuation. (b) Dual-Doppler analysis performed with masking of data where ARMOR has attenuation.
which yields an approximate beam height of 1.5 km above the radar site’s elevation. Leading up to the 2328 UTC start time of the tornado based on the official survey (NWS Nashville, 2020), a local maximum in wind at 2326 UTC of 51.5 kts increased to 68 kts at 2328 UTC.

For the purposes of this analysis, cross-sections will be centered on the low-level downdraft maximum south-southeast of the MV’s vorticity maximum, as this is nearest to where severe weather occurred. In addition, storm-relative motion vectors shown in these figures are based on the best estimate for the MV’s motion, which is approximately 28.8 m s\(^{-1}\) from 236° (or \(u=24\) m s\(^{-1}\), \(v=16\) m s\(^{-1}\)). From this view, a relative offset between the vertical vorticity maximum and the strongest downdraft at 1 km above ARMOR can be observed with a -5 m s\(^{-1}\) contour found to the south-southeast of the vorticity maximum (see Fig. 5.19). The magnitude of the vorticity maximum within the MV at this time is around \(1.6 \times 10^{-2}\) s\(^{-1}\) below the 1 km analysis level (not shown).

The north-south cross-section taken through the downdraft maximum (Fig. 5.20) depicts strong storm-relative inflow from the south feeding into an updraft which peaks in magnitude at around 9-10 m s\(^{-1}\) at an altitude of 3-4 km above ARMOR. A clear overturning circulation is centered around 4 km above ARMOR with a strong and fairly narrow downdraft on the southern periphery of the MV, where a secondary peak in vertical vorticity is noted. When examining the east-west cross-section taken through the downdraft maximum (Fig. 5.20), one finds a much different storm-relative flow structure. As expected by Trapp and Weisman (2003), the low-level vorticity maximum associated with the MV is presumed to generate a
Figure 5.19: 1km CAPPI imagery around 2331 UTC on 12 February 2020 with reflectivity from KHTX as the background color fill along with storm-relative motion vectors and (a) vertical vorticity and (b) vertical velocity (W) in color contours. Dashed black lines represent the extent and location of cross-sections taken through the low-level downdraft maximum south of the MV.
strong downward-directed vertical perturbation pressure gradient acceleration which produces a persistent downdraft over time that prevents updrafts from forming at the leading edge of the QLCS nearest to the MV. It should be noted that persistent and widespread lightning activity was observed by the GOES-16 GLM south of the MV (see Fig. 4.7), and with the observed updraft values above the freezing level nearing the threshold values suggested by Zipser and Lutz (1994) to support considerable lightning activity in continental thunderstorms, it is considered likely that the retrieved updraft magnitudes are reasonably close to reality.
Figure 5.21: X-Z cross-section through the low-level downdraft maximum around 2331 UTC with the same background and colored contours for vertical vorticity as in Fig. 5.19, but also included colored contours representing W with red contours for upward motion and blue contours for downward motion. This view is looking at the QLCS from the south. Storm-relative motion vectors are plotted using the same vector as in Fig. 5.19. From this viewpoint, the strong downdraft is clearly offset from the vorticity maximum towards the east. Within this storm-relative reference frame, it can be seen that the flow structure through the system is considerably different, with no clearly-defined updraft observed at the true leading edge of the QLCS relative to its direction of motion.
The next time period of interest is 2346-2353 UTC, which leads up to and includes the extensive uprooting of dozens of hardwood trees approximately 10 km east of Pulaski in central Giles County, TN. 1 km CAPPI imagery and cross-sections from both 2346 UTC and 2351 UTC will be used in this analysis to observe changes in the MV structure during this period of severe weather production. This view depicts a 1 m s\(^{-1}\) downdraft contour to the southeast of the low-level vorticity maximum while a 1 m s\(^{-1}\) updraft contour appears to wrap around the vorticity center (see Figs. 5.22 and 5.23). The vorticity maximum associated with the MV peaks near \(1.4 \times 10^{-2} \text{s}^{-1}\) near the lowest analysis level. Between the two times, an apparent distortion of the vorticity field into a more elliptical shape occurs while the downdraft contour seems to lag behind its original position relative to the vorticity field extending southward from the MV. Reports of downed trees were timed at 2345 and 2348 UTC with the most extensive damage occurring around 2353 UTC.

North-south cross-sections taken through the low-level downdraft maximum (Figs. 5.24 and 5.25) depict a similar storm-relative flow structure to the first set of cross-sections. Differences in the positioning of the 1 km downdraft contour relative to the vorticity maximum result in noticeably different vertical velocity magnitudes. At 2346 UTC, an updraft maximum of 5-6 m s\(^{-1}\) is centered in altitude around 3 km while the adjacent downdraft features a similar magnitude at a similar height. Relative to the vertical vorticity maximum, this downdraft remains displaced to the southern periphery of the MV circulation. Vertical vorticity peaks at just over \(1.4 \times 10^{-2} \text{s}^{-1}\) in the updraft zone while the peak in vertical vorticity related to the southern flank of the MV is around \(0.6-0.7 \times 10^{-2} \text{s}^{-1}\). At 2351 UTC, the 1 km downdraft contour
Figure 5.22: As in Fig. 5.19, but around 2346 UTC on 12 February 2020.
Figure 5.23: As in Fig. 5.19, but around 2351 UTC on 12 February 2020.
Figure 5.24: As in Fig. 5.20, but around 2346 UTC on 12 February 2020. The downdraft maximum is displaced towards the southern flank of the MV’s vertical vorticity field with a maximum value between $0.6-0.8 \times 10^{-2} s^{-1}$ observed in this cross-section. Notice the considerably stronger vertical vorticity maximum within the QLCS’s updraft zone well-south of the MV, with a maximum value exceeding $1.6 \times 10^{-2} s^{-1}$.

shifts to a position more directly south of the MV’s vertical vorticity maximum, which corresponds closest in time to the production of extensive damage to hardwood trees.

In figure 5.25, a maximum updraft of 9 m s$^{-1}$ centered near 4 km altitude shows a considerably steeper slope than in figure 5.24, although a similar downdraft maximum of 5-6 m s$^{-1}$ is observed in both. The vertical vorticity maximum in the updraft zone at 2351 UTC peaks at a similar value to that observed at 2346 UTC, whereas the maximum value observed in relation to the MV is $1.2 \times 10^{-2} s^{-1}$ with the downdraft penetrating the MV on its southern flank.

The final time period of interest is 0003-0008 UTC, as damage to a barn was reported around 0006 UTC approximately 10 km southwest of Petersburg, TN. It
Figure 5.25: As in Fig. 5.20, but around 2351 UTC on 12 February 2020. The downdraft maximum is displaced towards the southern flank of the MV’s vertical vorticity field with a maximum value near $1.2 \times 10^{-2} \text{s}^{-1}$ observed in this cross-section. Notice the continuation of strong vertical vorticity within the QLCS’s updraft zone well-south of the MV, with a maximum value exceeding $1.6 \times 10^{-2} \text{s}^{-1}$.
should be noted that the viewing angle from ARMOR became increasingly parallel to the QLCS after 00 UTC, resulting in attenuation at C-band which may have a negative impact on the quality and accuracy of the dual-Doppler analyses. As a result, artifacts may be affecting the retrieved W field, especially at 0008 UTC. Aside from this caveat, the 1 km CAPPI imagery (see Figs. 5.26 and 5.27) between the two times once again depicts a structural change to the vertical vorticity field of the MV, with an apparently more elliptical shape appearing just after severe wind production, which also corresponds to the low-level downdraft maximum lagging behind the gust front in a similar fashion to that observed between 2346 UTC and 2351 UTC. In addition, the -3 m s\(^{-1}\) contour appeared at 0008 UTC in the 1 km CAPPI imagery where only a -1 m s\(^{-1}\) contour was originally found at 0003 UTC. Vertical vorticity magnitudes are generally weaker in these analyses, but this may be a side effect of the nearly QLCS-parallel viewing angle from ARMOR which results in attenuation at C-band with increasing range. Nonetheless, the MV is still clearly visible with maximum magnitudes of 0.8 \(\times 10^{-2}\) s\(^{-1}\) at 0003 UTC and 1.0 \(\times 10^{-2}\) s\(^{-1}\) at 0008 UTC at 1 km altitude.

North-south cross-sections through the 1 km downdraft maximum show a 9-10 m s\(^{-1}\) updraft maximum in the range of 2-3 km altitude at 0003 UTC and 3-4 km altitude at 0008 UTC. When comparing figures 5.26 and 5.27, it is observed once again that the low-level downdraft maximum appears to lag behind the gust front, thus resulting in the north-south cross-sections intercepting different parts of the QLCS. Nonetheless, vertical vorticity maxima at the gust front reach up to around 1.0 \(\times 10^{-2}\) s\(^{-1}\) at both times while vorticity associated with the MV circulation peaks
Figure 5.26: As in Fig. 5.19, but around 0003 UTC on 13 February 2020.
Figure 5.27: As in Fig. 5.19, but around 0008 UTC on 13 February 2020.
in the range of $0.6-0.7 \times 10^{-2}\text{s}^{-1}$ at both times. Also, a descent in the downdraft contours is visible between 0003 UTC and 0008 UTC with the 5 m s$^{-1}$ downdraft contour reaching below 2 km at 0008 UTC. The downdraft at both times appears to be generally centered on the low-level vorticity maximum, but it is clear from figures 5.28 and 5.29 that this is still displaced towards the south side of the MV’s center of circulation.

Unfortunately, the most expansive wind damage area produced by the MV took place between 0015-0030 UTC, during which radar data from the ARMOR was entirely attenuated over the MV, thus precluding accurate dual-Doppler wind retrievals in this region. Radar data solely from KHTX reveals that this corresponded in time to the merging of an independent meso-$\gamma$-scale bow echo segment with the southern flank of the MV, thus creating a superposition of not only the rotational

**Figure 5.28:** As in Fig. 5.20, but around 0003 UTC on 13 February 2020.
and translational motion of the MV, but also of the descending rear-inflow jet feeding into the back of the bow echo. This mechanism of extensive severe wind production was proposed by Atkins and St. Laurent (2009b) and occurs when a bow echo merges with the right flank of an MV relative to its direction of motion, thus combining forces related to the horizontal perturbation pressure gradient between the MV and the cold pool with the forward motion of the QLCS, the rotational motion of the MV, and the hydrostatically-induced jet which drives rear-to-front flow into the apex of the bow echo.

In addition to the known mechanisms of severe wind production within MVs and bow echoes, it appears that a descending downdraft core may have also played some role in promoting severe winds. Given that a horizontal perturbation pressure gradient is well-known to occur between the MV's vorticity maximum and the cold
pool (Trapp and Weisman, 2003), it is proposed here that a downdraft impinging on this horizontal gradient from above may further tighten the perturbation pressure gradient due to the combination of high perturbation pressure resulting from the buoyancy force below the downdraft with the high perturbation pressure hydrostatically induced by the cold pool. It is also possible, and perhaps more likely, that the downward mixing of high-momentum air from even just 1 km above ground combined with the aforementioned forces to produce this expansive swath of damaging winds. Further investigations into the perturbation pressure field would be required to support or negate these hypotheses.

5.3.3 Kinematics of the Bow Echo Segments

A trio of closely-spaced meso-γ-scale bow echo segments moved rapidly northward along the QLCS, producing occasional wind damage. These segments were associated with enhanced updrafts and vertical vorticity at 5 km above ARMOR, as seen in figure 5.30. To obtain insight into the flow structure within these segments, X-Z and Y-Z cross-sections will be taken through the northernmost bow echo at two times: 2337 UTC and 2358 UTC.

The northernmost bow echo segment first entered the analysis domain at the 2337 UTC analysis time. It is clear to see its location based on the arc in upward motion and the local maximum in vertical vorticity observed at 3 km above ARMOR in figure 5.31. It should be noted that the 1 km gridding used in this analysis may smooth over finer-scale features associated with the bow echo segment and also that X-Z and Y-Z cross sections do not represent views that are perpendicular to the bow
Figure 5.30: As in Fig. 5.19, but at 5 km above ARMOR and around 2358 UTC on 12 February 2020. Notice the three distinct updraft maxima collocated with the three vertical vorticity maxima associated with the bow echo segments.
echo’s motion vector, which is estimated to be 236° at 35 m s\(^{-1}\) \((u=29\) m s\(^{-1}\), \(v=19.5\) m s\(^{-1}\)). Nonetheless, cross-sections through the updraft maximum, which is located nearest to the apex of the bow echo, are shown in figures 5.32 and 5.33 to provide a visualization for the updraft tilt.

Cross-sections through the updraft maximum at 3 km altitude (Fig. 5.30) reveal the upshear-tilted updraft structure which is known to support the development of rear-inflow jets due to latent heating and the subsequent hydrostatic lowering of pressure above the surface cold pool (Weisman, 2001; Markowski and Richardson, 2010). In figure 5.32, it is shown that the updraft in this bow echo peaks at just over 11 m s\(^{-1}\) around 3 km altitude at 2337 UTC while vertical vorticity peaks at 1.4-1.5 \(x10^{-2}\) s\(^{-1}\) at around the same level, albeit displaced slightly further west than the updraft maximum. Although the storm-relative motion vectors depict an entirely front-to-rear flow within the bow echo, this is likely due to the cross-section itself not being parallel to the bow echo segment’s motion vector, which should otherwise indicate some degree of rear-to-front flow beneath the upshear-tilted updraft. In figure 5.33, a north-south cross section shows a much more gradual updraft over the system’s cold pool with a clear overturning circulation centered between 3-4 km above ARMOR. Storm-relative motion vectors from this view depict strong low-level flow towards the south in the convective precipitation zone, exceeding 20 m s\(^{-1}\) near the northern edge of the updraft zone.

As the bow echo segment tracked north-northeastward towards the MV, the arc of enhanced upward motion originally seen at 2337 UTC near the leading edge of the bow echo largely disappears (see Fig. 5.34) and a more vertically-erect updraft
Figure 5.31: As in Fig. 5.19, but at 3 km above ARMOR and around 2337 UTC and using a storm motion vector of $U=29 \text{ m s}^{-1}$, $V=19.5 \text{ m s}^{-1}$. Notice the arc of enhanced upward motion extending northwestward from the intersection of the dashed black lines.
Figure 5.32: X-Z cross-section around 2337 UTC on 12 February 2020 taken through the updraft maximum at 3 km altitude. Plotting conventions are the same as in Fig. 5.20. Notice the updraft feeding into a zone of reflectivity overhang with a maximum magnitude of just over 11 m s$^{-1}$ around 3 km altitude. Vertical vorticity peaks at 1.4-1.5 $\times$10$^{-2}$s$^{-1}$ just west of the updraft core.
Figure 5.33: As in Fig. 5.20, but around 2337 UTC on 12 February 2020 and taken through the updraft maximum at 3 km as shown in Fig. 5.30. Notice the broad overturning circulation centered around 3-4 km altitude with an updraft maximum just over 11 m s\(^{-1}\) located around 3 km altitude. Vertical vorticity peaks at 1.4-1.5 \(\times10^{-2}\) s\(^{-1}\).
structure takes shape (see Fig. 5.35). However, the relative placement of the updraft maximum at 3 km remains near the apex of the bow echo and the updraft structure maintains some degree of upshear tilt, continuing to support a rear-inflow jet into the back of the bow echo. Other updraft and vertical vorticity maxima associated with the other two bow echoes can be seen in figure 5.34.

Vertical cross-sections through the bow echo at 2358 UTC reveal the continuation of a focused updraft region which suspends hydrometeors, as evidenced by the echo overhang feature observed in the reflectivity color fill. With a freezing level measured by multiple observed soundings averaging around 3.6 km at 00 UTC (see Table 5.2), the peak updraft of just over 11 m s\(^{-1}\) around 4-5 km altitude shown in figure 5.35 generally meets the thresholds found by Zipser and Lutz (1994) to support considerable lightning activity in continental thunderstorms, which was observed by the GOES-16 GLM (see Fig. 4.7) in this case. The occurrence of lightning through-
Figure 5.35: As in Fig. 5.32, but around 2358 UTC on 12 February 2020. Notice the more vertically-oriented updraft compared to Fig. 5.32 with a peak vertical velocity of just over 11 m s\(^{-1}\) centered around 4-5 km altitude.

out this event serves as a proxy for updraft strength which supports the observed vertical velocities retrieved by PyDDA. Figure 5.36 depicts an upshear-tilted updraft zone which is physically consistent with theories for bow echo structure (Weisman, 2001).

5.3.4 ARMOR-MAX Dual-Doppler Results

The ARMOR-MAX dual-Doppler domain featured a short baseline of around 18 km, resulting in the ability to retrieve winds with a higher spatial resolution than what is feasible for the ARMOR-KHTX dual-Doppler domain. However, due to the maximum elevation angle of 11.5° used by both ARMOR and MAX, substantial data voids exist within the dual-Doppler lobes (see Fig. 5.37), which limits the accuracy of the wind retrievals. Nonetheless, at 0015 UTC, a portion of the QLCS which exhibited
Figure 5.36: As in Fig. 5.33, but around 2358 UTC on 12 February 2020. Note the rearward-tilted updraft zone over the bow echo’s cold pool.

A hail signature was captured near the edge of the northeastern dual-Doppler lobe and will be briefly examined below. A storm-relative motion vector of 236° at 35 m s\(^{-1}\) (u=29 m s\(^{-1}\), v=19.5 m s\(^{-1}\)) was found to closely match the motion of this portion of the line, as it was originally associated with the southernmost bow echo segment in the trio of bow echoes described previously, so the storm-relative vectors have this motion subtracted out.

At 0015 UTC, a zone of enhanced reflectivity (60-65 dBZ), reduced CC (0.85-0.95), and depressed Z\(_{\text{dr}}\) (-1 to 2 dB) at around 0.6 km above the KHTX radar elevation (not shown) was located just north of Harvest, AL, indicating the presence of rain and melting hail likely in the range of 1-2 cm reaching the ground in this location. The 3 km CAPPI image (Fig. 5.38) shows considerably stronger low-level updrafts than those retrieved by the ARMOR-KHTX dual-Doppler analyses. Very
Figure 5.37: 5 km CAPPI from ARMOR-MAX dual-Doppler analysis around 0015 UTC on 13 February 2020 with reflectivity from ARMOR as the background color fill and ground-relative motion vectors as well as (a) vertical vorticity contours and (b) vertical motion contours. Notice the large data void covering much of the dual-Doppler lobes, thus precluding accurate wind retrievals due to a lack of data up to storm top.
strong vertical vorticity maxima are also noted in this zone, although it is possible that these are erroneously strong given that data from the MAX radar were attenuated before reaching the front of the QLCS. Figure 5.39 depicts an east-west cross section through the 3 km updraft maximum and shows a significant echo overhang feature which extends more than 5 km ahead of the convective precipitation zone ($\geq 40$ dBZ). The vertical vorticity field indicates that the wind shift line associated with the system’s cold pool extended away from the convective precipitation zone and was located about midway into the echo overhang region. The updraft maximum in this analysis is just over 13 m s$^{-1}$ around an altitude of 2.5-3.5 km, which is somewhat lower in height than what was shown by the ARMOR-KHTX analyses, but stronger in magnitude. Given the data void above approximately 7 km, spurious retrievals are likely causing the excessive updrafts and downdrafts present above 5 km altitude. As a result, the retrieved W field, while generally placed in physically consistent locations with respect to the QLCS, is deemed to be too contaminated to make inferences.

5.4 Boundary Layer Variability – Quantifying Rapid Changes in the Environment and in QLCS Cold Pool Strength

ASOS sites from the lower Mississippi Valley through the Tennessee Valley provided high-quality, 1-min resolution measurements of thermodynamic and kinematic variables. For simplicity, temperature and moisture variables were combined to derive equivalent potential temperature ($\theta_e$) and will be shown in the plots below. Interestingly, the observed $\theta_e$ drop which occurred across the QLCS was much greater earlier in its life than when severe weather was occurring (see Figs. 5.40 and 5.41).
Figure 5.38: As in Fig. 5.37, but instead at 3 km above ARMOR, with storm-relative motion vectors, and dashed black lines denoting cross-sections taken through the updraft maximum.
Figure 5.39: X-Z cross-section taken along the dashed black line shown in Fig. 5.38 with the same background color fill, storm motion vectors, and vertical vorticity contours as in Fig. 5.38. W contours follow the same convention as in Fig. 5.20.

This was likely caused by entrainment of a distinct dry layer starting near 3 km above ground level observed in 12 UTC soundings (see Fig. 5.42) from sites south and east of the large stratiform precipitation area seen across Texas into Arkansas in figure 5.1. With the development of convection along the leading edge of the cold pool described earlier, strong convective overturning was likely able to bring down considerably lower values of $\theta_e$ during the early stages of the QLCS, when the pre-storm environment was relatively pristine. Later on, while dry layers near and above 3 km were still observed on soundings launched across northern Alabama, and these may have played some role in severe wind production, it is proposed that the increasingly moist inflow air as well as layer lifting induced ahead of the QLCS helped to moisten the dry layer and result in relatively smaller $\theta_e$ drops.
Figure 5.40: Time series plots of equivalent potential temperature at (a) Monticello, AR (KLLQ), (b) Greenville, MS (KGLH), and Greenwood, MS (KGWO). Notice the increasing magnitude and steepness of the drop-off in $\theta_e$ over time, representing an increasingly strong cold pool.
Figure 5.41: As in Fig. 5.40, but for (a) Tupelo, MS (KTUP), (b) Muscle Shoals, AL (KMSL), and (c) Huntsville, AL (KHSV) and through 02 UTC on 13 February 2020. Notice the comparatively weaker magnitude of the decline in $\theta_e$ relative to the sites in Fig. 5.40.
Figure 5.42: Atmospheric sounding and derived variables from a radiosonde launched from Jackson, MS around 11 UTC for the 12 UTC hour. Similar dry layers starting near 3 km were observed in adjacent 12 UTC soundings launched from Lake Charles, LA; New Orleans, LA; and Birmingham, AL. Image source: NOAA/NWS/SPC.
Table 5.1: 500 mb heights measured by iMet radiosondes launched by UAH teams during this event. N/A denotes "not available" and indicates where a sounding was either not launched at a particular site or whose maximum height failed to reach the 500 mb level.

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>Huntsville, AL</th>
<th>Decatur, AL</th>
<th>Falkville, AL</th>
</tr>
</thead>
<tbody>
<tr>
<td>1500</td>
<td>5576.9</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>1800</td>
<td>5561.1</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2100</td>
<td>N/A</td>
<td>5585.5</td>
<td>5573.0</td>
</tr>
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<td>2200</td>
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<td>N/A</td>
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Surface pressure changes and 500 mb height falls can help one infer the dynamic characteristics of the environment. On 12 February 2020, subtle 500 mb height falls of around 5 dam (see Table 5.1) and surface pressure falls of 9-10 mb over 6-8 hours (see Figs. 5.43 and 5.44) were observed. The low-level mass response yielded increases in $\theta_e$ that approached or exceeded 30 K over 6 hours in some locations, including at KMSL and KHSV. Pressure rises across the QLCS gust front were mainly in the 2-4 mb range, with a generally steady trend after QLCS passage until the synoptic cold front arrived as much as 9 hours later.

CAPE values derived from observed soundings launched across north Alabama (not shown) demonstrated steep rises within the 3 hours preceding QLCS arrival, with changes ranging from a 73 (114) J kg$^{-1}$ rise in MLCAPE (MUCAPE) within 3 hours at Falkville, AL to a 196 (292) J kg$^{-1}$ rise in MLCAPE (MUCAPE) within 2 hours at SWIRLL. These observations verify the findings by King et al. (2017) that rapid rises
**Figure 5.43**: As in Fig. 5.40, but for station pressure.

**Figure 5.44**: As in Fig. 5.41, but for station pressure.
in CAPE can and often do occur in cool-season pre-storm environments. In addition to advection contributing to these rapid changes, layer lifting ahead of a QLCS can also contribute to the release of potential instability, which was also observed in this case and is described below.

Freezing level height is important because it delineates the warm cloud depth, which would be most supportive of collision-coalescence growth of liquid hydrometeors, and also affects the ability for hail aloft to reach the surface. Although severe (1”; 2.54 cm) hail was not reported during this event, hail signatures in KHTX dual-polarization radar products were observed at its lowest tilt from around 2330 UTC through 0015 UTC in northern Alabama primarily within the trio of bow echoes. This occurred where enhanced reflectivity maxima exceeding 65 dBZ corresponded with lowered CC values into the 0.85-0.95 range, and $Z_{dr}$ mostly in the -1 dB to 1 dB range. Table 5.2 shows a listing of freezing level heights based on radiosondes launched by UAH teams in three separate locations over north-central Alabama. While freezing levels rose within the warm advection regime well-ahead of the QLCS, a significant drop in freezing level height was noted between 23 UTC and 00 UTC, with a 526 m drop at Huntsville, 546 m drop at Decatur, and a 420 m drop at Falkville. This substantial cooling mostly in the 600-650 mb layer at these sites took place in the hour preceding QLCS arrival, which may have occurred as a result of height falls and/or layer lifting due to ascent induced by the QLCS. Given that height falls at each location were quite small based on Table 5.1 and no precipitation occurred anywhere near these these sites, which would have caused latent cooling in this layer, it is considered
Table 5.2: As in Table 5.1, but for freezing level heights. Heights were linearly extrapolated from the nearest two measured levels with a temperature near 0°C. Only the Huntsville 23 UTC sounding exhibited multiple heights where temperatures crossed the 0°C threshold. In this case, the highest level where the threshold was crossed was taken to be the final freezing level.

<table>
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<th>Falkville, AL</th>
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<td>N/A</td>
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</tr>
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<td>N/A</td>
</tr>
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</tr>
<tr>
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</table>

probable that layer lifting ahead of the QLCS accounted for this substantial reduction in freezing level height.

To complete the thermodynamic analysis, recall that Sherburn and Parker (2014) found low-level lapse rate, mid-level lapse rate, and 0-3 km shear vector magnitude as three key ingredients which showed some of the highest skill in identifying significant severe HSLC environments. The authors combined these ingredients and normalized their values into the SHERBS3 parameter, which has a threshold value of 1. Table 5.3 provides values of SHERBS3 from the three locations previously described and these values clearly indicate that conditions were more than supportive for the development of severe HSLC convection. It should be noted that the strength of the 0-3 km shear vector contributed most greatly to raising these values above 1, as all sites observed greater than 30 m s\(^{-1}\) of shear from 22 UTC onward. Nonetheless, low-level lapse rates also exceeded the set threshold value of 5.2 K km\(^{-1}\) in the 1-2
Table 5.3: As in Table 5.1, but for SHERBS3 parameter.

<table>
<thead>
<tr>
<th>Time (UTC)</th>
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<th>Decatur, AL</th>
<th>Falkville, AL</th>
</tr>
</thead>
<tbody>
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<td>N/A</td>
</tr>
<tr>
<td>1800</td>
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<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2100</td>
<td>N/A</td>
<td>1.090</td>
<td>1.180</td>
</tr>
<tr>
<td>2200</td>
<td>1.164</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2300</td>
<td>N/A</td>
<td>1.373</td>
<td>1.536</td>
</tr>
<tr>
<td>0000</td>
<td>1.774</td>
<td>1.588</td>
<td>1.462</td>
</tr>
</tbody>
</table>

hours preceding QLCS arrival, with all sites reporting 6-6.5 K km\(^{-1}\) in their 23 UTC and 00 UTC soundings. Thus, it is evident that rapid evolution of the pre-QLCS environment took place, which contributed to the forecast challenge of anticipating severe convection. The next subsection will provide a review of forecast discussions written by NWS forecasters from Huntsville and surrounding forecast offices to provide insight into the thought processes and understanding of the potential for severe weather preceding this event.

Kinematic changes are also important to note as part of the environmental evolution as the arrival of a low-level jet often serves to enhance both low-level shear and warm advection ahead of a QLCS. In figure 5.45, a time series of vertical wind profiles from the 915 MHz wind profiler on MIPS depicts very strong low-level shear peaking within the 2 hours preceding the arrival of the QLCS, with wind magnitudes reaching up to 60 kts at the 1 km level and up to 70 kts near 1.2 km. These measurements help to independently verify the strong low-level shear measured by radiosondes, which averaged around 25 m s\(^{-1}\) in the 0-1 km layer just ahead of the QLCS. A time series
of wind and virtual potential temperature measured at Courtland, AL (CTD) can be seen in figure 5.46, with similarly strong low-level shear arriving within a couple hours of the QLCS. In addition, rapid thermodynamic changes were observed at CTD, with near-surface virtual potential temperature values increasing from the low 280s K around 14 UTC to near 300 K by 23 UTC. However, the persistently positive $\theta_v$ lapse rate over time through the lowest 1.4 km indicates a statically stable low-level environment over CTD.

To help exemplify the relevance of rapidly-changing environments in HSLC events, a couple of Wunderground sites within Middle Tennessee will receive brief attention. Figure 5.47 depicts selected surface observations from the town of Centerville, TN in Hickman County. A surface temperature and dewpoint rise of nearly 10°F (5.6°C) was observed in the hour preceding QLCS arrival, likely due to the northward-lifting of the stationary front first noted on the 21 UTC surface analysis from the Weather Prediction Center. Figure 5.48 shows surface measurements from approximately 12 km northwest of Lawrenceburg, TN. This site included solar radiation measurements, which help to identify periods of cloudiness. The most significant warming and moistening occurred not in alignment with any diurnal trend, but later in the day when the warm front apparently reached the site and southerly flow took over, which corresponds approximately to when solar radiation dropped to around 50 W m-2. A rise in temperature and dewpoint of approximately 8°F (4.4°C) occurred over 45 mins between 1:45 PM CST and 2:30 PM CST and an overall temperature rise of 17°F (9.4°C) occurred in the 4 hours leading up to QLCS arrival.
Figure 5.45: Time series of vertical wind profiles up to around 3.3 km above ground level at SWIRLL from the 915 MHz wind profiler on the MIPS platform. Note the increase in wind magnitude observed in the few hours leading up to the arrival of the QLCS, marked by the distinct wind shift around 0040 UTC. Data source: UAH MAPNet.
Figure 5.46: Time series of vertical wind profiles measured between 12 UTC on 12 February 2020 and 00 UTC on 13 February 2020 at Courtland, AL, which is approximately 65 km west of where the MIPS was located. Time increases from right to left on this plot. Color fill represents values of virtual equivalent potential temperature. Notice both the strong low-level wind shear as well as the substantial modification of the low-level thermodynamic profile resulting primarily from warm air advection ahead of the QLCS. Data source: NOAA Physical Sciences Laboratory, Courtland site.
Figure 5.47: Time series (in CST) of temperature, wind and gust speed, and wind direction at Wunderground site KTNCENTE3, which is located in the town of Centerville, TN in Hickman County. Wind direction data is considered to be suspect due to a general lack of easterly component north of the stationary front and lack of southerly component when the warm air arrived. Image source: Wunderground.
Figure 5.48: Time series (in CST) of temperature and dewpoint, wind and gust speed, wind direction, precipitation accumulation, and solar radiation at Wunderground site KTNLAWRE20, which is located approximately 12 km northwest of Lawrenceburg, TN. Image source: Wunderground.
5.5 NWS Forecaster Insight into the Predictability of the QLCS

Area forecast discussions (AFDs) written by National Weather Service forecasters are typically updated once every three to six hours throughout any given day and provide insight into the thought process behind the forecasts they issue. As a result, these discussions provide a snapshot in time of the interpretation of model guidance and the anticipated outcome of an upcoming weather event, which can change substantially with the next update as new data becomes available. A synthesis of forecast discussions issued by forecast offices in Memphis, TN; Jackson, MS; Huntsville, AL; and Nashville, TN will allow the reader to better understand the forecast challenges associated with this event and how the forecasters’ expected outcomes evolved over time.

From the early morning forecast packages typically issued around 3-4 AM local time, forecasters were aware of the highly-amplified pattern which would lead to a rapidly evolving environment ahead of an eastward-advancing synoptic cold front. However, some forecasters expected the synoptic warm front to lift through their forecast zones much more quickly than what was actually observed, particularly in the Huntsville and Nashville forecast offices. Forecasters in these offices both expected the warm front to reach the AL/TN border by midday on 12 February 2020 and continue northward through the afternoon and evening. All forecasters acknowledged that severe weather was possible along or just ahead of the cold front and that instability would be a limiting factor while very strong wind shear would be supportive of severe weather.
Mid-late morning forecast discussions maintained mentions about the severe weather threat, with the Memphis AFD stating that the HRRR model was among the most aggressive model guidance in showing MLCAPE values of 200-500 J kg$^{-1}$ reaching into northeastern Mississippi ahead of the cold front. The Huntsville AFD noted that model guidance depicted “frontal convection” developing over Mississippi around 21 UTC, with this convection not moving into Alabama until 00 UTC at the earliest. Thus, the forecasters were aware of a HSLC setup that would likely feature convection forced by a linear lifting mechanism, but questions remained with regard to the timing and magnitude of destabilization ahead of the modeled cold front.

With the afternoon forecast packages typically issued between 3-4 PM local time, forecasters in the Memphis office first noticed the narrow band of convection which initiated near the Mississippi River. They also recognized that model guidance was lagging behind observations in terms of the convection and noted that their southeastern zones were most likely to see severe weather where MLCAPE values of 200-400 J kg$^{-1}$ would combine with very strong 0-1 km shear and helicity to support a damaging wind and weak tornado threat. The Jackson forecasters anticipated convective development to remain confined to the cold front and used the term “squall line” for the first time in their discussion. Strong shear was deemed to be sufficient for severe weather production, but a shear vector that was expected to be mostly parallel to the initiating boundary limited the expectations to a linear complex of storms with damaging winds and possibly a tornado due to anticipated 0-1 km helicity of 200-400 m$^2$s$^{-2}$. Huntsville forecasters noted the slow northward progress of the warm front in their mid-afternoon update with 20 UTC temperatures across their coverage area in
the upper 50s to low 60s. The mention of a surface cold front and a 60-70 kt low-level jet suggested that strong low-level shear and rapid advection would play a role in the severe weather threat. In terms of the timing of severe weather, high-resolution models showed a convective line reaching the AL/MS border by 00-01 UTC, although the writer noted that the current pace of convection in Mississippi would bring the threat to northwest Alabama by 23-00 UTC. An isolated tornado threat was also mentioned by the writer as effective storm-relative helicity values were projected to reach up to 400 m²s⁻². Lastly, the Nashville forecasters acknowledged that a lack of instability and “marginal shear” would preclude a “significant severe weather event”, but also noted the presence of a narrow band of convection across western Tennessee.

As seen earlier in the results, the QLCS ultimately arrived in northwest Alabama around 2230 UTC, produced severe weather mainly between 2230 UTC and 0030 UTC, then pushed through much of northern Alabama by 0100 UTC. It was not associated with a synoptic cold front and was largely supported by strong low-level forcing along a convectively-reinforced cold pool. The Huntsville forecasters acknowledged that the QLCS was not associated with a cold front in their evening forecast update. Later in the night, a separate line of convection would develop along the synoptic cold front, but only several sporadic reports of severe weather would occur across east central Alabama in association with this line, including a brief tornado.
CHAPTER 6

SUMMARY AND DISCUSSION

This work featured a detailed observational case study of a cool-season severe weather event occurring under extreme high-shear, low-CAPE (HSLC) conditions using a combination of dual-polarized radars, surface observing systems, vertical profiling instruments, and lightning data. The original forcing mechanism of the quasi-linear convective system (QLCS) was determined to stem from a surface cold pool produced by a persistent band of showers which was supported by weak upper-level divergence in the right-entrance region of a 300 mb jet streak. Shallow convection developed along the leading edge of this cold pool across western Mississippi into northeast Louisiana and tracked rapidly eastward, outpacing the upper-level divergence and the developing cold front. This differs from the conventional findings of Sherburn, Parker, et al. (2016), which showed that strong upper-level forcing for ascent was commonly associated with QLCS formation and maintenance. Meso-γ-scale bow echo segments and an MV developed along the QLCS as weakly unstable air combined with strong low-level shear. Upon entering northwest Alabama, severe wind production began in association with the MV. Swaths of severe winds would continue to occur on the southern flank of the MV with approximately 15-minute
repeat intervals from 2230 UTC through around 0030 UTC. Two weak tornadoes, a hurricane-force wind gust, and numerous instances of damaging winds would lead to countless downed trees, power lines, and structural damage along the MV’s path. Bow echoes also shifted northeastward along the QLCS, producing occasional patches of wind damage south of the MV. When the northernmost bow echo merged with the southern flank of the MV, the most extensive wind damage occurred.

The examination of this HSLC QLCS from a microphysical perspective using dual-polarized radars represents the first documented case which applied these methods to infer changes in QLCS structure. This investigation revealed that the initial convective line was reminiscent of a narrow cold frontal rainband, with a continuous band of shallow convection largely remaining below the freezing level, representing slabular ascent (James, Fritsch, et al., 2005). Over time, updrafts within the early QLCS deepened, producing stronger reflectivity factor primarily driven by increased mean drop diameters. With further deepening of the convective updrafts, as evidenced by increased heights in reflectivity isosurfaces, as well as the appearance of \( Z_{dr} \) and \( K_{dp} \) columns, the intense low-level shear began to have a greater influence on the organization of the convection. The QLCS became more cellular in nature as the development of meso-\( \gamma \)-scale features typically associated with severe weather took place. These features originated from the southern reaches of the QLCS, where greater instability resided, and gradually shifted northeastward along the line as it approached northwestern Alabama. An MV developed separately from these features and began producing severe wind in far northwestern Alabama around 2230 UTC. South of this MV, strong updrafts maintained a persistent reflectivity overhang which occasionally
exhibited bounded weak-echo region (BWER) characteristics, particularly in the updrafts which fed into the bow echo segments. Hail signatures also appeared within the bow echo segments, presumably due to the enhanced updrafts, which were able to sustain hail growth amidst a shallow freezing level height of around 3.5 km. After a couple of hours, the vast majority of severe weather had occurred and a return to the more slabular convective structure observed previously in Mississippi took place over northeastern Alabama.

Dual-Doppler analyses of this QLCS allowed for a more detailed investigation into the kinematics of the system, with an emphasis on characterizing the strength of vertical motions and elucidating the storm-relative kinematic structure of the QLCS and its severe bow echoes and MV. From the CAPPI imagery, narrow updrafts characterized the QLCS south of the MV, with a characteristic magnitude of 10-12 m s\(^{-1}\) centered typically around 3-4 km in altitude. One novel observation is that the MV featured a persistent downdraft of around 7-8 m s\(^{-1}\) centered at an altitude of 3-4 km which was slightly offset to the southeast of the MV’s low-level vertical vorticity center throughout the 45-min analysis time frame. The vertical structure of the downdraft resulted in its low-level center remaining southeast of the MV’s vertical vorticity maximum as well. This downdraft prevented the development of updrafts along the leading edge of the QLCS immediately east of the MV, which was expected by Trapp and Weisman (2003). Y-Z cross-sections showed that inflow air originating from the south rose over the gust front, exhausted its convective instability, then appeared to partially feed this downdraft as part of a persistent overturning circulation centered around 4 km in altitude. A second updraft zone was encountered north of the MV.
with comparatively weaker updraft maxima of 8-10 m s$^{-1}$ and broader updraft areas than south of the MV.

The MV was analyzed just after the occurrence of an EF-0 tornado (near 2330 UTC) as well as during the time frame in which extensive wind damage to hardwood trees took place (around 2346-2353 UTC) and when damage to a barn was reported (around 0003-0008 UTC). Findings from this work showed that each occurrence of severe wind was preferentially located on the right flank of the MV relative to its direction of motion, which aligns with the published literature on damaging wind production in MVs (Trapp and Weisman, 2003; Wakimoto, Murphey, Nester, et al., 2006; Atkins and St. Laurent, 2009b). Subsequently, it is probable that similar dynamical mechanisms were at play in the production of severe wind, even though these earlier studies were conducted for moderate to large CAPE regimes.

The meso-$\gamma$-scale bow echoes were investigated to better understand their kinematic structure under extreme HSLC conditions. Although cross-sections were not taken perpendicular to the apex of the northernmost bow, X-Z cross-sections found the upshear-tilted updraft structure well-known to be associated with bow echo development (Weisman, 2001). 3 km CAPPI imagery at 2337 UTC showed an arc of enhanced upward motion corresponding to the front edge of the northernmost bow echo segment and 5 km CAPPI imagery at 2358 UTC verified that local updraft maxima were clearly associated with each bow echo segment. Typical max updraft magnitudes were just over 11 m s$^{-1}$ at both of the times analyzed, but at 2337 UTC, the height of this maximum was closer to 3 km altitude whereas at 2358 UTC the maximum rose to around 5 km altitude. This rise in height corresponded to a more
vertically-erect updraft as seen in east-west cross-sections at both times, so it is possible that the cold pool associated with the bow echo weakened, which allowed for the very strong vertical shear to more equally balance the horizontal vorticity generated by the cold pool. Although not explored in detail due to attenuation from the ARMOR’s perspective, the merging of the northernmost bow echo with the southern flank of the MV resulted in widespread wind damage. Whether the more vertically-oriented updraft at 2358 UTC returned to a more upshear-tilted configuration prior to or during this merger is unknown, but this merger process has been recognized as one that can lead to widespread damaging winds (Atkins and St. Laurent, 2009b).

Overall, the MV and bow echoes observed in this case showed similarities to those observed in moderate-CAPE studies (e.g., Trapp and Weisman, 2003; Wakimoto, Murphey, Nester, et al., 2006; Atkins and St. Laurent, 2009b). The location of damaging winds associated with the MV were all preferentially located on the right flank of the MV relative to its direction of motion while more sporadic wind damage was observed near the apex of bow echoes. Despite weak MLCAPE, these structures managed to develop and persist for extended periods of time, with a single MV lasting for over 3 hours and traveling over 290 km. Inferred pressure perturbations based on the strength and persistence of the MV’s circulation combined with $\theta_e$ deficits of $\sim$10 K across the QLCS’s gust front suggest that a tightened horizontal perturbation pressure gradient existed between the MV and the QLCS’s cold pool, and this combined with the rotational and translational motion to produce damaging winds on the right flank of the MV. This suggests that MVs and bow echoes rely heavily on the presence of strong low-level shear to organize and they are not particularly sensitive to static
stability, so long as convection reaches a sufficient depth to "feel" the effects of strong low-level shear. Subsequently, when appraising the pre-storm environment for severe weather potential in similar HSLC conditions, close attention should be given to the depth of convection and depth/strength of the low-level shear.

An investigation into the mesoscale variability preceding with the passage of the QLCS was also conducted to better understand how rapid changes influenced the pre-storm environment (e.g., King et al., 2017). Time series of $\theta_e$ at sites roughly aligned perpendicular to the QLCS’s direction of motion showed $\theta_e$ rises of generally $\sim 15$-$20$ K in the 3 hours preceding QLCS arrival. Interestingly, greater drops in $\theta_e$ were found to occur within the convective precipitation region early in the life of the QLCS than later on. The significant declines in $\theta_e$ peaked with a maximum observed drop of around 28 K in less than 10 min (with a $\sim 20$ K drop occurring in 2 min) at Greenwood, MS. These $\theta_e$ drops are attributed to the vertical mixing of a deep dry layer with a base observed to start around 3 km above ground on official soundings retrieved from NWS offices in the lower Mississippi Valley (e.g., see Fig. 5.42). Once the QLCS was mature and tracking across northeast MS and northern AL, $\theta_e$ drops shrunk to a more modest $\sim 10$ K drop within 5-15 min. Later on, inferred layer lifting ahead the QLCS is proposed to contribute to moistening of the dry layers observed on soundings from northern Alabama, as freezing level heights dropped from around 4.1 km to near 3.6 km between the 23 UTC and 00 UTC soundings. At locations ahead of the QLCS, pressure decreases were generally 3-5 mb within 3 hours preceding QLCS arrival, with greater declines occurring earlier in the life of the QLCS. Across the QLCS’s gust front, pressure rises were mostly in the 2-4 mb range, with a generally
steadily until the synoptic cold front arrived as much as 9 hours later.

CAPE values derived from observed soundings launched across north Alabama demonstrated steep rises within the 3 hours preceding QLCS arrival, with increases ranging from a 73 (114) J kg\(^{-1}\) rise in MLCAPE (MUCAPE) within 3 hours at Falkville, AL to a 196 (292) J kg\(^{-1}\) rise in MLCAPE (MUCAPE) within 2 hours at the Severe Weather Institute - Radar and Lightning Laboratories (SWIRLL) at UAH. These rapid rises in CAPE observed under a HSLC regime support the numerical simulations performed by King et al. (2017). The SHERBS3 parameter (Sherburn and Parker, 2014) exceeded its design threshold of 1 for all reliable soundings obtained by UAH teams from three locations in north central Alabama within 3 hours of QLCS arrival (see Table 5.3), with the primary contributor to these values being the 0-3 km shear vector magnitude, followed by low-level lapse rates. Vertical wind profiles obtained by the MIPS platform and the NOAA Physical Sciences Laboratory atmospheric profiling site at Courtland, AL both sampled significant low-level shear with wind speeds of 60-70 kts around 1 km above ground arriving within the 2 hours preceding QLCS arrival. A brief look at two personal weather station sites participating in Wunderground’s network of citizen weather observers demonstrated the role of rapid advection in conditioning the pre-storm environment ahead of a QLCS in HSLC conditions. At the site located in Centerville, TN, an approximately 10°F (5.6°C) rise in temperature and dewpoint occurred within the one-hour period preceding QLCS arrival. At the site located west of Lawrenceburg, TN, substantial temperature and dewpoint rises occurred primarily as a result of advection behind the warm front,
with as much as an 8°F (4.4°C) rise in temperature and dewpoint occurring in 45 min, and an overall temperature rise of 17°F (9.4°C) occurring in the 4 hours leading up to QLCS arrival.

Finally, a review of forecast discussions issued by NWS offices affected by this event provided insight into the thoughts and perceptions of the severe weather potential prior to its occurrence. In summary, forecasters were aware of the anomalously amplified pattern and the potential for a HSLC severe event to occur. Model depictions of the warm front’s rapid northward progress and the development of convection along the synoptic cold front were both incorrect, although credit should be given to the lack of convection forecast guidance depicted ahead of the linear convective system that they produced. Nevertheless, the timing of convection as depicted by model guidance was multiple hours too slow as a narrow band of showers advanced away from the synoptic cold front and became the severe QLCS of interest. The extreme shear and storm-relative helicity values were recognized by most forecasters to yield a nonzero tornado threat, although limited instability precluded a more widespread outbreak of tornadoes. Even so, the long-track MV in this case lasted longer than most single meso-γ-scale MVs have been observed in some well-known high-shear, high-CAPE events (Knupp et al., 2014; Lyza et al., 2021), and subsequently produced several swaths of severe weather with a repeat interval of approximately 15 minutes over a 2-hour period. Two weak tornadoes were produced by this MV, which verified the conceptual models of the forecasters, but in the operational setting, these tornadoes were not warned, and furthermore, the first hour of severe weather was not warned. It is recognized that isolated tornado cases and severe events occurring under
marginal conditions are inherently harder to warn (Dean and Schneider, 2012; Sherburn, Parker, et al., 2016), so it is difficult to consider this event as one that should have been anticipated better, particularly when severe weather production began in a portion of Alabama which lacks low-level radar coverage, thus making it very difficult to clearly discern the presence of an MV. So, with respect to recommendations for future improvement in the operational setting, continued research and development of reliable techniques to better understand when HSCL environments are conducive to severe weather production is key to improving warning operations and keeping the public safe in these marginal environments.

Forecast models in this case were not the most skillful at properly depicting the overall evolution of the pre-storm environment (lifting the warm front northward too quickly) or the actual forcing mechanism for the QLCS (which was not the synoptic cold front), and this led to some forecasters anticipating certain outcomes which did not come to fruition. As a result, one route for future work is to perform a modeling study with similar initial conditions to those observed in this case. Parameterization schemes which are best-suited for HSCL regimes have already received attention in the literature (Cohen, Cavallo, Coniglio, and Brooks, 2015; Cohen, Cavallo, Coniglio, Brooks, and Jirak, 2017) and testing those findings in this case as well as varying resolution and input data may serve as a means to perform sensitivity testing on the ability of high-resolution models to capture shallow HSCL convection in extreme environments such as the one studied here. Additional modeling studies may be performed on the MV itself, particularly due to its long-track, long-duration, and propensity to produce damaging winds instead of tornadoes. A numerical modeling
investigation into the cause and role of the downdraft observed on the southern flank
of the MV is yet another potential route for future work on this case, as the downdraft
never fed directly into the center of the MV’s vertical vorticity maximum at low levels,
where the greatest negative perturbation pressure should have been observed.
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