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COLD-SEASON SEVERE QLCS EVENTS OVER NORTH AL:
CLIMATOLOGY, CLOUD, AND BOUNDARY LAYER CHARACTERISTICS

by

CHRISTOPHER A. LISAUCKIS

A THESIS

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

HUNTSVILLE, ALABAMA
2018
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Submitted by Chris Lisaukis in partial fulfillment of the requirements for the degree of Master of Science in Atmospheric Science and accepted on behalf of the Faculty of the School of Graduate Studies by the thesis committee.

We, the undersigned members of the Graduate Faculty of The University of Alabama in Huntsville, certify that we have advised and/or supervised the candidate on the work described in this thesis. We further certify that we have reviewed the thesis manuscript and approve it in partial fulfillment of the requirements for the degree of Master of Science in Atmospheric Science.

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ABSTRACT
The School of Graduate Studies
The University of Alabama in Huntsville

Degree: Master of Science  College/Dept.: Science/Atmospheric Science

Name of Candidate: Chris Lissauckis
Title: Cold-Season Severe QLCS Events Over North AL: Climatology, Cloud, and Boundary Layer Characteristics

Stratocumulus clouds, hypothesized to control boundary layer processes during cold season tornado events, are investigated. Associated observed boundary layer profiles of wind, low clouds (e.g., cloud base height, cloud fraction), water vapor, and boundary layer cloud depth near tornadogenesis events over northern Alabama and nearby areas of the Southeastern United States are presented. Observed stratocumulus cloud fraction for QLCS’s is 97%, and 83% for supercells. Furthermore, stratocumulus clouds first occurred an average of 200 and 90 minutes prior to tornadogenesis for QLCS’s and supercells respectively. Mean LCL heights for all QLCS cases were found to be 659 m for QLCS, and 649 m for supercell cases. Both the subcloud boundary layer and surface layer are often statically stable, unlike the dry adiabatic profile in classical cloud topped mixed layers. Thus, large bulk shear magnitudes of 14 and 18 m/s occur within the respective 0-0.5 km and 0-1.0 km layers.

Abstract Approval:  Committee Chair
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iv
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CHAPTER 1

INTRODUCTION

Severe storms during the southeastern United States cool season (November 1-March 31) occur under a broader range of conditions than observed throughout many regions of the country. Accurately forecasting such events can be a difficult task, as many tornadoes have occurred unexpectedly, within environments that would seem to suggest little to no tornadic activity. This study explores southeastern USA cool season tornado climatology and the storm environment using a suite of UAH and NOAA instrumentation, which serves to not only reveal interesting spatiotemporal patterns of tornado occurrence, but also offers clues concerning the formation of cool season tornadoes which are often referred to as high shear low CAPE (HSLC, Sherburn and Parker 2012) events in which convective available potential energy is <500 J kg$^{-1}$ and bulk 0-6 km shear exceeds 18 m s$^{-1}$. A southeast cool season storm mode climatology is also necessary to gain an understanding of the types of tornado producing thunderstorms and their frequency during the cool season.

Vertical wind shear is highly important in tornado development, especially in cool season months when tornado formation depends on shear to a much greater extent, as opposed to what is typically considered an ideal balance of kinematic and thermodynamic variables. The investigation of boundary layer kinematic parameters such as bulk wind
shear, storm relative helicity, and 10 m wind are believed to offer potentially significant contributions to cool season tornado forecasting. Moreover, stratocumulus cloud behavior is hypothesized to be linked to patterns of thermal and moisture advection, and boundary layer processes, which are not fully understood for cool season HSLC events. The National Weather Service operates ceilometers, located at each ASOS (Automated Surface Observation Station) location, that provide observations of cloud base height and cloud fraction.

The primary goals of this study are to characterize the following thermodynamic and kinematic characteristics of cloud-topped boundary layers associated with cool season QLCS and supercell tornadoes: cloud behavior, impact on stability (day vs. night), associated wind profiles, and storm relative helicity (SRH). A secondary goal is to provide data that is useful for evaluation of numerical model parameterizations. Questions surrounding the presence of a cloud topped mixed layer in many cool season tornado events are specifically of concern, as this can dictate the degree of mixing within the boundary layer (Cohen et al. 2015, 2017). Furthermore, the intensity of turbulent mixing affects the magnitude of SRH by partially controlling wind profiles within the surface layer. In the presence of strong mixing, bulk wind shear is decreased by mixing, which increases the surface wind and decreases wind above the surface layer (within the turbulent boundary layer). However, with the cessation of surface heating, an increase in wind shear is often observed, which can play an important role in nocturnal tornado events. The WSR-88D Velocity Azimuth Display (VAD) wind profile product is used herein to determine low-level bulk wind shear and SRH.

Research questions for this investigation include the following:
1) What is the distribution of cloud base heights, and values of cloud fraction associated with the clouds atop or within the boundary layer during cold season severe weather and tornado events?

2) What is the stability class of the subcloud boundary layer? Is this boundary layer a classical cloud-topped mixed layer?

3) Is the subcloud layer coupled to the cloud layer? If it is a fully coupled cloud-topped mixed layer, then the subcloud lapse rate should be dry adiabatic (mixed) and turbulent.

4) What are the characteristics of the wind profile within the cloud topped mixed layer from the surface through the cloud layer, and above?

Answers to these and other questions will further advance understanding of cool season severe storm and tornado events.
CHAPTER 2

BACKGROUND

2.1 Characteristics of the HSLC Cool Season Environment

Researchers who have studied tornadic thunderstorms that occur primarily during the southeastern United States cool season have highlighted the need for the meteorological community to perform additional study. Furthermore, Sherburn and Parker (2016, hereafter SP16) discussed the importance of mid-level lapse rates and the ability to recognize certain synoptic meteorological patterns to improve HSLC forecasts. Also mentioned by SP16 was the tendency for surface temperatures to warm during the several hours before frontal passage during tornadic events which contributes to buoyancy of the surface based parcel and the buildup of conditional instability. Such buoyancy tends to be greater during the daytime due to insolation, resulting in greater lower and mid-level lapse rates.

SP16 also observed that the peak shear vector magnitude occurred during the more stable overnight hours due to a reduction in the vertical mixing which reduces surface layer wind flow and initiates the inertial oscillation (Van de Wiel et al. 2010) in which winds above the surface layer accelerate during the nocturnal period. This results in a nocturnal increase in both deep layer shear and storm relative helicity. SP16 discussed at
length the ability of potential instability associated with an elevated mixed layer (EML) to increase mid-level lapse rates through evaporative cooling and large scale forcing for ascent. During cool season HSLC events, it is believed that the release of potential instability is one of the primary means by which such events occur. SP16 also mentioned that composite parameters offer very little assistance in forecasting HSLC tornado events primarily because they occur in an environment characterized by CAPE values of \( \leq 500 \) J kg\(^{-1}\). This emphasizes the need for forecasters to pay close attention to the specific combination of parameters that are often observed during HSLC tornadoes.

Cool season QLCS tornadoes within the southern Mississippi Valley were studied by Rogers et al. (2016), who found that nocturnal tornadoes are typically QLCS related and are associated with very pronounced and rapidly moving longwave troughs. Although the majority of cool season QLCS’s produce weak tornadoes, it should be noted that strong and occasionally intense (EF3) tornadoes can be associated with QLCS’s, mostly in the southeast. Broyles et al. (2002) found that the southeastern United States cool season features the greatest height falls and inherently the strongest upper level flow of any region and season within the U.S. It should be noted that strong height falls are not always necessary for tornadic thunderstorm development; however, southeast cool season QLCS tornadoes certainly tend to occur most often in environments featuring such height patterns.

Jackson and Brown (2009) investigated the relationship between the level of free convection (LFC) and lifted condensation level (LCL) and remind us that a low LCL does not necessarily imply a low LFC due to potential capping inversions and weak low-level lapse rates. Strong capping inversions in the southeast cool season are infrequent,
and therefore the height separating the LCL and LFC is often small, especially when compared to the severe storm environment of the Great Plains. Jackson and Brown (2009) also note that 0-3 km CAPE offered very little discrimination between tornadic and non-tornadic environments in the southeast.

Craven and Brooks (2004, hereafter referred to as CB04) determined that high values of storm relative helicity (SRH) may produce high false alarm ratios, and should be used with care during the tornado forecasting process. This is especially true when SRH may not capture details of the wind profile, a principle discussed at length by Kerr and Darkow (1996) who mention the importance of monitoring the evolution of storm relative flow as opposed to being bound by the strict confines of storm relative helicity. Environments where high values of speed shear exist in the absence of substantial directional shear are examples, as this situation results in an environment dominated by crosswise vorticity, which could only be augmented by right moving supercells or strong vorticity tilting by QLCS cold pools. CB09 discussed the possibility of potentially using bulk shear in the 0-1 km layer as a discriminator between tornadic and non-tornadic storms, as this provides forecasters a variable that tends to vary less spatiotemporally and does not require an estimated storm motion vector, as is required to compute SRH. Tornadoes within the HSLC environment often occur in the presence of 0-1 km shear of at least 15 m s\(^{-1}\), and in some cases values approach 28-31 m s\(^{-1}\) (SP16). Additionally, QLCS propagation can be faster than suggested by the background mean wind due to dynamical cold pool effects (Corfidi 2003) which can result in SRH values which are significantly underestimated.

Lawrence (2005) developed equation 1 for estimating the LCL using only surface
temperature (T) and dewpoint (Td):

\[ \text{LCL} = 125(T - T_d). \]  \hspace{1cm} (1)

Lawrence reports that this equation is accurate to within 5% when used in the presence of a fully mixed (i.e., nearly constant potential temperature and mixing ratio) boundary layer. Significant errors in using this equation can arise in cases when the boundary layer is not well mixed and is often signaled when a discrepancy exists between observed cloud base height and the cloud base height estimate produced from the Lawrence equation.

Rasmussen and Blanchard (1998, hereafter RB98) found that the height of the LCL, measured with radiosonde observations, was correlated with greater tornadic potential. They found that relatively high LCL heights of above 1200 m favored cool rear flank downdrafts and tornadogenesis failure, while LCL heights of 800 m or lower suggest warm and buoyant rear flank downdrafts and tornadogenesis. Being able to anticipate the evolution of the LCL height during severe storm events is an important factor in accurately assessing tornadic potential. RB98 also presented findings on bulk shear in tornadic environments, primarily that 0-6 km shear offers discrimination between supercell and non-supercell environments and 0-1 km between tornadic and non-tornadic events.

James et al. (2005) discusses how slabular convection, convective systems which are defined as an unbroken two-dimensional swath of ascent (James et al. 2005), typically form in an environment featuring a pronounced low-level jet, which enhances lower tropospheric shear to magnitudes seldom experienced in conventional tornadic environments. James et al. (2005) also indicate that slabularity is affected by buoyancy, convective inhibition, and the tropospheric wind profile, all of which influence storm
mode and evolution. It should also be mentioned that the HSLC effective layer SRH as described by Guyer and Dean (2010) is often less than total layer SRH due to the frequent occurrence of low-topped storms resulting in a reduced depth of atmospheric flow affecting the cloud-bearing layer. This further illustrates the importance of understanding the background environment, including the depth of the buoyancy profile and kinematics throughout this layer. Esterheld and Giuliano (2008) determined that the angle between the 0-500 m bulk shear vector and the 10 m storm relative inflow identifies environments dominated by low-level streamwise vorticity, which favors tornadogenesis. Critical angle values within the 60-115 degree range have been identified as being most associated with tornado events across many regions (Guyer and Hart 2012). Numerous southeast cool season HSLC significant tornadoes have occurred with an observed critical angle within the 55-70 degree range according to Guyer and Hart. Both the study by Esterheld and Giuliano (2008) as well as Guyer and Hart (2012) exemplify the importance of performing frequent wind profile hodograph analysis.

2.2 Observational or Modeling Studies of HSLC QLCS Characteristics

Cope (2004) examined an early morning mid-Atlantic QLCS that produced four weak short-lived tornadoes within an HSLC environment that lacked recognition of tornadic potential by many, if not all forecasters. Cope mentioned that the tornadoes were believed to be of non-descending type, implying that significant mid-level rotation was not evident during Doppler radar volume scans immediately preceding the tornadoes. The strongest rotational velocity ($V_{rot}$), 10.8-13.3 m s$^{-1}$, were observed on WSR-88D in the lowest 1 km of the atmosphere. Interestingly, the thunderstorms responsible for this event were low-topped and produced little to no lightning. Such shallow storms have been
observed during HSLC cool season events in the southeastern states associated with
tornado producing storms of up to F2/EF2 intensity (Guyer and Dean 2015). No tornado
warnings were issued for two of the four tornadoes in Cope’s study, which is likely due
to rapid low-level development, movement, and dissipation of the parent circulation.

Observations were made by Lane and Moore (2006) of a broken line of non-
supercell tornadic thunderstorms via the Charlotte, NC terminal Doppler weather radar
(TDWR). This study focused on a radar reflectivity signature known as the “broken-S”
and further identified it as a possible basis for tornado warning issuance. As with the
Cope (2004) study, Lane and Moore acknowledged that many non-supercell tornadoes do
not develop in the top-down sequence followed by many supercell thunderstorms
containing mesocyclones and that many times, non-supercell tornadoes are very difficult
to detect. The “broken-S” signature may offer significant discrimination and improved
warning capability due to Doppler velocity data often offering little if any assistance in
detection of such events. According to results from Cope’s January 13, 2006 case study,
it is believed that a rear-to-front descending air current may have played a significant role
in the development of the breaks within the convective line. This lends further credence
to the idea that mid-level dry air plays a significant role in the formation of the cool
season HSLC tornadic environment. Modeling studies performed by Weisman and Trapp
(2003) and Trapp and Weisman (2003) indicate downward tilting of crosswise
environmental vorticity, via sinking air currents, to be the origins of such vortices,
including those which are found in rear inflow jets of QLCS’s. Listed by Cope as
parameters to monitor for the development of such HSLC tornadic situations are 0-3 km
shear values of at least 20 m s$^{-1}$ and strong linear forcing. Latimer and Kula (2010) and
McAvo (2000) studied HSLC tornadic QLCS’s which produced tornadoes whose signatures were not immediately apparent on Doppler radar. Both studies discussed at length the importance of relying upon reflectivity signatures during tornado warning issuance in such situations.

2.3 Challenges with Numerical Simulation of the BL of HSLC Events

Investigations by Cohen et al. (2015, 2017) shed light on parameterization schemes that have difficulty reproducing the planetary boundary layer. This is especially important during severe storm events due to the capability of heat, moisture, and momentum fluxes in modifying boundary layer stability. Specifically, observations have indicated that stratocumulus clouds play a significant role in controlling boundary layer momentum fluxes (King et al. 2017). Cohen et al. discussed how the total energy-mass flux (TEMF), a hybrid local-nonlocal scheme (Angevine et al. 2010) most accurately represented environments consisting of both cumulus and stratocumulus clouds.

Furthermore, the Grenier-Bretherton McCaa (GBM) local scheme (Grenier and Bretherton 2001) accurately depicts reductions in stratocumulus clouds due to vertical mixing. It was noted that certain boundary layer parameterization schemes maximize model accuracy within the southeast U.S. cool season. Cohen et al. (2015) concluded that all schemes inflate CAPE predictions, however, non-local schemes were found to contain the most error. Due to the importance of low level vertical wind shear, the effects of cloud-topped mixed layers and turbulent mixing within the boundary layer are crucial because of the effect they have on both boundary layer shear and stability.

King et al. (2017) mentions that in a study of simulated preconvective environments within the warm sector of midlatitude cyclones, that cloud fractions greater
than 80% existed in all cases, which acted to reduce the amount of daytime radiational warming. As mentioned previously, certain model parameter schemes, namely the TEMF and GBM handle the effects of stratocumulus clouds best. Also of interest is the character of the sub-cloud lapse rate which would be indicative of subcloud stability and could affect the distribution of vertical wind shear. Cohen et al. states that turbulence kinetic energy (TKE) quantifies motions resulting from perturbations due to vertical shear, buoyancy, turbulent transport, and dampening from molecular viscosity. Resultantly, turbulent mixing within the PBL is critical to understanding the cool season HSJC tornadic environment.
CHAPTER 3

DATA AND METHODOLOGY

Severe thunderstorm and tornado events used to compile tornado climatology, case study, and parameter calculations were obtained from the National Center for Environmental Information’s (NCEI) event database and SPC event archives during the 17 week period from the last two weeks in November through the first two weeks in March for events dating to 2005. Once events were identified, details pertaining to tornado time, intensity, and location were determined using the NCEI dataset. Cloud base height observations for this study originated from the NOAA NCEI five minute ASOS database. NCEI’s one minute dataset was not employed due to lack of ceilometer data. Only ASOS locations within 80 km of the location of tornado occurrence were used in attempt to preserve spatiotemporal continuity. In recording the observed cloud base height, data were gathered from the lowest one minute ceilometer observations throughout five minute periods. This data was recorded for three hours prior, until forty-five minutes following tornadogenesis. Cloud fractions were computed using bins of five minute ASOS ceilometer data for three hours prior to tornadogenesis.

Data for LCL height were obtained from NOAA SPC’s Mesoanalysis Archive which utilizes RAP model data output. Specifically, the SPC uses lowest 100 hPa mean layer mixing ratio LCL height data in all mesoanalysis products. This has been shown by
Craven et al. (2002) to be the most precise method of calculation over a wide range of atmospheric conditions. When surface observations displayed rapid changes (e.g. during the late morning hours of March 2, 2012), conditions were captured before the change took place to ensure continuity of data. Interpolation was often used when attempting to record the most accurate LCL height value from mesoanalysis charts. Equation 1 was used to estimate height of the LCL via surface T and Td obtained from the NCEI, using the same ASOS locations chosen for all surface based observations.

Wind profiles for calculations of kinematic parameters such as bulk shear, storm relative helicity, and critical angle were gathered from WSR-88D VAD wind profile data obtained from NCEI’s NEXRAD Archive and from the University of Wyoming Sounding Database. Storm relative helicity calculations were performed using equation 2 from Davies-Jones et al. (1990) which was incorporated into an EXCEL spreadsheet.

\[ \text{SRH} = - \int_{0}^{h} k \cdot (V - c) \times \frac{\partial V}{\partial z} dz \]  

(2)

For all cases, radar observed storm motion was found using GR Level 2 software. This was done to avoid any potential error from mean wind calculations. Critical angle calculations were determined using conventional hodograph analysis performed by hand and the angle between the 10 m storm relative inflow and 500 m bulk shear vectors measured via protractor. Two minute averaged ASOS observed 10 m wind was used for surface wind statistics as well as hodograph construction. Furthermore, surface wind often exhibited some variability and as a result was time averaged to account for this. Specifically, 10 m surface wind speed and direction were averaged over the 30 minute period before tornado occurrence for best results in construction of hodographs and
surface wind statistics.

Nulls for all datasets contained within this study were defined as days where severe weather reports occurred in the absence of reported tornadoes. Storm mode classifications were assigned using data from WSR-88D data downloaded from the NCEI’s Nexrad Data Archive. Storms were classed as supercells when rotation within the cell attained a rotational velocity $\geq 18 \text{ m s}^{-1}$ and possessed a discrete outline in reflectivity pattern. Discrete supercells are defined in this study as a supercell reflectivity structure having been identified but remained associated with nearby precipitation, while isolated supercells are storms which were completely detached from precipitation external to the cell. For supercells embedded within MCS’s, care was exercised to determine that the cell displayed significant autonomy in order to be identified as a separate updraft core. The requirements for the QLCS classification was conjoined reflectivity echoes with meso-beta scale convective elements present along the forward propagating QLCS.

Doppler radar PPI images for individual case studies were obtained from the Advanced Radar for Meteorological and Operational Research - ARMOR (Peterson et al. 2005), a 350 kW C band radar located at the Huntsville International Airport, and from the WSR-88D S-band radar located at Hytop, AL via the NCEI database.

Data for subcloud lapse rate calculations were obtained from several of UAH’s in house instruments including the Microwave Profiling Radiometer (MPR), lidar ceilometer, and surface station. The MPR provided infrared temperature ($T_{IR}$) at cloud base, and care was taken to use observations from cases with optically thick low level clouds, thereby avoiding significant contributions to measured radiance from heights above the ceilometer-detected cloud base that would conceivably inflate the lapse rate
calculation. Surface data from UAH’s berm surface station was used for measurements of 0.5 and 10 m temperature to calculate surface layer stability. This surface layer stability calculation was performed by subtracting the 0.5 from the 10 m temperature. When this metric was positive, the surface layer was deemed absolutely stable.

Three cloud base height measures were used to better understand how numerical weather prediction scheme results differ from observed cloud base, and how improvements can be made accordingly. The Lawrence equation estimate provides a quick approximation of the lowest possible (surface based) LCL in a given environment, which can help indicate the degree of atmospheric mixing etc. The SPC model-derived estimate is based on the lowest 100 hPa mean values of potential temperature (θ) and mixing ratio (r_v). For an ideal mixed layer (both θ and r_v are assumed to be constant with height) the surface-based and SPC values of the LCL would be identical. These two LCL values would differ for situations in which a negative gradient in r_v, and/or a stable layer exist.
CHAPTER 4

CLIMATOLOGY

The following climatology is important due to the large number of QLCS cases observed during the southeast U.S. cool season. Specifically, this section defines the months and times of day when HSLC tornadoes occur, which promotes greater understanding of the boundary layer stability and increased wind shear conditions which spawn such tornadoes.

Although the defined period in November and March features the greatest number of tornado days, climatology strongly suggests that the most likely time periods within the southeast cool season for violent tornadoes is the last two weeks of November. However, significant (EF2 or greater) cool season tornadoes have occurred, even in association with major outbreaks of tornadoes, throughout the entire defined cool season period of November 15 - March 15. Fig. 4.1 shows that the height of the climatological tornado frequency was observed during the last two weeks of November and the first two weeks of March from March 2004 through February 2017. Cool season tornado climatology within the southeastern United States is dominated by weak (EF0/EF1) tornadoes during the nighttime hours of 0200-0700 UTC, with a secondary maximum observed during the period from 10-11 UTC as indicated by Fig. 4.2.
Figure 4.1. Plot showing distribution of cold season tornado days by month for the 17 week period from November 15 to March 15 for all years.

Figure 4.2. Plot showing tornado occurrence via bi-hourly period for all years.
Interestingly, tornadoes as strong as EF4 have occurred during this secondary early morning timeframe within the defined study period of 2004-2017. The morning of 6 February 2008 featured several long track supercell storms over northern Alabama, resulting in four confirmed tornadoes within 120 km of the ARMOR radar, two of which were rated EF4. Looking further back in time, an early morning EF4 tornado also occurred in Middle Tennessee on December 24, 1988 (NCEI), further indicating the southeast region’s long-standing history with early morning HSLC violent tornadoes during cool season months.

During the period 2004-2017, 67% of tornadoes (39 tornadoes) that occurred within 120 km of the ARMOR radar originated from QLCS’s. The remaining 19 tornadoes (33%) were found to originate from supercells, which produced all of the cool season intense (EF3+) tornadoes within this study. This finding is very similar to other cool season tornado climatologies which included the southeastern states (i.e. Thompson et al. 2012). No tornadoes were found to originate from multicell or other forms of convection during this study. This is attributed to the strength of cool season dynamics, as opposed to spring and summer months when greater buoyancy allows for a broader convective mode spectrum.

Although tornadoes occur throughout a wide range of months within the southeastern U.S., cool season events are not only the most difficult to predict, but can occur at all hours and exhibit a nocturnal maximum. The following sections will feature results and discussion pertaining to the cool season severe storm environment and the challenges that lie within. Particular attention is given to the QLCS environment, as the majority of HSLC tornadoes result from such storm systems.
CHAPTER 5

STORM ENVIRONMENT

5.1. Boundary Layer Cloud and Thermodynamic Characteristics

The following section features results from fifty cases, including QLCS’s and supercells, pertaining to cloud base height data collected from four data sources, including ASOS and UAH ceilometers, NOAA SPC Mesoanalysis, and the Lawrence Equation based on surface T and Td (Lawrence 2005).

5.1.1 Cloud Base Statistics

A comparison of KHSV ASOS and UAH MIPS ceilometer data consisting of ten cool season tornado days that featured tornadogenesis within 60 km of the ceilometer locations is seen in Fig. 5.1. This data revealed a 50 m average difference between the two ceilometers, with the ASOS having an average of 962 m and UAH 912 m. Additionally, this result indicates that the two ceilometers provide a similar cloud base height value, and can be compared or combined into a common data set.

The temporal relationship between the onset of stratocumulus clouds and tornadogenesis for 52 cool season tornado cases within 120 km of ARMOR indicates stratocumulus cloud existed for an average of 200 minutes prior to QLCS tornado events, and 90 minutes prior to supercell tornadoes. Cases exist when clouds were present for
both a significantly greater and lesser time than these averages. The length of time which
stratocumulus clouds existed prior to tornadogenesis is a factor of two greater for QLCS
cases.

Lifted condensation level height for the ASOS ceilometer and SPC mesoanalysis
LCL data observed in Fig. 5.2 are very similar, with the ceilometer observed cloud base
being 17 m higher than SPC values average for all 50 cases was considered. The
difference between the Lawrence estimate and SPC is maximized when a relatively large
negative gradient in mixing ratio exists within the boundary layer. This can occur because
the Lawrence estimate uses surface based values, while SPC uses lowest 100 hPa mean
(mixed) layer values of $\theta$ and $r_v$. The largest difference between SPC and ASOS among
the individual storm mode members was observed within the non-tornadic (null)
supercell category.
This difference could result from larger LCL heights when compared to the other convective modes, with greater values having the potential to display more variance. The QLCS null category features a boundary layer that is more mixed as indicated by the proximity of the Lawrence and SPC values. The Lawrence equation average values are 151 m less than ceilometer observed cloud base in QLCS cases and 156 m less in non-tornadic QLCS events. Interestingly, ceilometer observed cloud base heights, and both the SPC and Lawrence LCL heights, were greatest for the twelve supercell null cases. The difference among the three measurements was also the greatest, which seems to be related to boundary layer mixing. Certainly, the relatively elevated LCL height has an impact on the ability for tornadoes to occur within the supercell null environment (RB98), by causing the development of a relatively cool rear flank downdraft. Figure 5.3
indicates that ASOS ceilometer observations for QLCS’s and supercells have the greatest 10-90% percentile range.

![Ceilometer Observed and Estimated Cloud Base Height](image)

Figure 5.3. Box and whisker distribution plot indicating both ceilometer observed cloud base height and estimated LCL height.

Overall, the distribution is very similar for supercells and QLCS’s when analyzing SPC mesoanalysis. Interestingly, the SPC mesoanalysis LCL values were more than 1000 m overestimated in the 25 February 2011 case. The lowest LCL height was found within the supercell storm mode as computed via the Lawrence equation. Note the significant underestimation in the Lawrence estimate, especially for supercells, which is believed to be the result of boundary layer stratification. The Lawrence equation captures the range of LCL heights very similarly to the ceilometer observations when all cases are considered. However, the inner quartile ranges of ASOS observations and SPC Mesoanalysis are most similar, and therefore the SPC Mesoanalysis is deemed to be more
representative than the Lawrence estimate. When a significant difference exists between SPC Mesoanalysis and Lawrence, the boundary layer would likely be stratified, whereas when the two values are close, a nearly constant mixing ratio profile is implied. Fig. 5.4 is the same as 5.3 but for the 23 null cases, including 11 QLCS and 12 supercell null cases. Notice the significant difference between the median for all ASOS ceilometer observed null cases when compared to all nulls for the Lawrence estimate and SPC.

Figure 5.4. Box and whisker distribution for 23 observed and estimated cloud base height null cases.

The dependence of cloud base height on time of day for the supercell storm and QLCS categories is presented in Fig. 5.5. The difference between 00-06 and 06-12 UTC for the supercell mode is the greatest by far, due to the prominence of a stratified nocturnal boundary layer which has high relative humidity during the 06-12 UTC period. Moreover, the period 06-12 UTC indicates significantly lower cloud base height, while 00-06 UTC shows less variation than 18-00 UTC. Furthermore, QLCS events displayed much greater observed cloud base height continuity among the three time periods.
QLCS’s often occur within an environment typically featuring little modification from insolation due to the high frequency of stratocumulus clouds. Moreover, if clouds are more numerous, cooler days and warmer nights would lead to lower variations in cloud base height. This illustrates well the significant differences in tornadic supercell cloud base height that exist due to daytime and nocturnal differences in boundary layer stability; however, the number of cases (3) for the 00-06 UTC QLCS period is limited.

Cloud fraction, observed via ceilometer, was determined for a three hour period prior to tornadogenesis and forty-five minutes after. This is crucial to this study as it is an indicator of moisture advection. Ceilometer derived cloud fraction shown in Fig. 5.6 indicates the median of the 50 cases to be 84%. The mean of 94% for QLCS cases exceeds that of supercell cases (80%).
5.1.2 Subcloud Lapse Rate: Case Studies

The following three QLCS case studies illustrate how subcloud stability promotes reduced boundary layer turbulence, which leads to an increase in kinematic parameters. Such an increase in bulk shear and storm relative helicity is crucial to the development of HSLC tornadoes within the southeast cool season. Observations from UAH’s suite of instrumentation were used to gather the observations for this portion of the study.

1) January 30, 2013

A QLCS oriented from 220 to 30 degrees formed over the southern Mississippi Valley and intensified as it moved east over the Tennessee Valley can be seen translating east across northern Alabama and southern Tennessee. This line consisted of transient rotational features, one of which produced an EF1 tornado in southwestern Lincoln County, Tennessee, 45 km from UAH at an azimuth of 340 degrees at 1046 UTC (Fig.
5.7). Showers are noted ahead of the QLCS, followed by a region of trailing stratiform precipitation. Mean layer parcel CAPE for this case was <250 J kg$^{-1}$. A total of six tornadoes, and seven damaging wind reports occurred with this event.

![ARMOR 0.7° PPI image of the QLCS which produce an EF1 tornado at 10:46 UTC on 30 Jan 2013.](image)

Two meter temperature and dewpoint observations remained mostly steady, at 22 and 17 °C respectively, during the several hours period prior to the occurrence of the tornado. The exception occurred at approximately 1050 UTC as precipitation evaporation produced a 1.5 °C decrease in temperature and °C increase in dewpoint. Ten m wind speed averaged near 11 m s$^{-1}$, but decreased within the post-QLCS stratiform region, with wind direction veering gradually from an average of 185 degrees during the first 90 minutes, to
190 degrees during the following 80 minutes, and 195 degrees throughout the final 40 minutes of the observation period.

MIPS ceilometer data is presented in Fig. 5.8, which once again shows mean cloud base height of 676 m. This value fits well within the distribution of cloud bases within this study. Interestingly, cloud base height increased significantly to slightly over 2000 m after precipitation arrival at the MIPS location. The cloud base initially appears to be low along the leading edge of convection and then increased in the stratiform region, which is commonly observed. ASOS ceilometer observed cloud base height for this case is 792 m, which is 116 m above the UAH MIPS ceilometer reading. This event features the greatest difference between the NOAA and MIPS platforms. Fig. 5.9 provides a measurement of integrated water vapor and integrated liquid water, the latter of which acts as a proxy to stratocumulus cloud thickness. Results indicate steady-state conditions until the

![Figure 5.8. Plot showing observed cloud base height (ceilometer), average cloud base height (dashed line), and estimated LCL height from the Lawrence equation.](image-url)
Figure 5.9. Plot showing vertically integrated vapor and vertically integrated liquid water.

The sub-cloud lapse rate plot in Fig. 5.10 for January 30, 2013 shows a uniform positive lapse rate of about 6 °C/km, indicated by the lowest derived lapse rate. Breaks in the lower level stratocumulus cloud deck and variations in liquid water path (vertically integrated liquid), which is defined as the weight of liquid water droplets in the atmosphere above a given surface area on earth in units of kg m⁻², produces variations in $T_{ir}$. Such variations result in the variable lapse rate and unrealistically high values that are particularly prevalent during the 0830-0930 UTC time frame. The retrieved lapse rate is quite consistent at ~6 C/km (a stable value) during the 0756-0835 and 0935-1050 UTC time periods. Lapse rate estimates are unreliable after precipitation began at the UAH site at 1053 UTC.
The surface layer temperature difference between the 10 and 0.5 m levels is uniformly positive at ~0.2 °C (Fig. 5.11), indicating surface layer stability that seems to be common during tornado events occurring within the nocturnal period. This behavior
is physically consistent with the derived subcloud lapse rate in Fig. 5.10. The sudden onset of rain causes cooling primarily in observations closest to the surface.

It is important to note the relationship among the stable surface layer portion of the boundary layer and very high 0-1 km bulk wind shear and storm relative helicity. The 0-1 km bulk shear value was approximately 26 m s\(^{-1}\), and the 0-1 km SRH value was 580 m\(^2\) s\(^{-2}\).

2) December 24, 2008

A QLCS oriented from 230 to 30 degrees formed over the southern Mississippi Valley around 1400 UTC and intensified as it moved east over the Tennessee Valley between 1800 and 2230 UTC. This line consisted of a Line Echo Wave Pattern (LEWP), which produced an EF1 tornado in southern Limestone County, AL at 2133 UTC as seen in Fig. 5.12. A narrow band of showers preceded the QLCS, and an extensive region of stratiform precipitation trailed it. Mean layer parcel CAPE for this case was <250 J kg\(^{-1}\). Two tornadoes and five damaging wind reports occurred within 120 km of ARMOR in association with this event. It is important to mention that a distance of 22 km existed between the tornadogenesis and UAH locations at an azimuth of 250 degrees.

The 2 m temperature rose steadily from 18 to 21 °C in the three hours leading up to tornadogenesis. Moreover, dewpoint temperature rose from 15 to 16 °C during the same three hour period. Both temperature and dewpoint fell abruptly with the passage of the QLCS at UAH. The 10 m wind speed averaged near 8 m s\(^{-1}\) during the three hour QLCS period, then peaked at 15 m s\(^{-1}\), before rapidly declining within the post-QLCS stratiform region. Wind direction remained uniform out of 200 degrees, then veered quickly to 300 degrees with the passage of the QLCS.
The 2 m temperature rose steadily from 18 to 21 °C in the three hours leading up to tornadogenesis. Moreover, dewpoint temperature rose from 15 to 16 °C during the same three hour period. Both temperature and dewpoint fell abruptly with the passage of the QLCS at UAH. The 10 m wind speed averaged near 8 m s\(^{-1}\) during the three hour QLCS period, then peaked at 15 m s\(^{-1}\), before rapidly declining within the post-QLCS stratiform region. Wind direction remained uniform out of 200 degrees, then veered quickly to 300 degrees with the passage of the QLCS.

The MIPS ceilometer cloud base evolution plot is displayed in Fig. 5.13. Cloud base slowly increased during the afternoon from 525 m at 1835 UTC to 900 m by 2115 UTC. A relatively uniform cloud base height is maintained after 2046 UTC.
except for several periods when breaks in the lowest cloud layer are observed, as well as when rainfall and frontal passage modified the subcloud layer. The MIPS and KHSV ASOS ceilometers recorded a mean cloud base height of 838 and 853 m respectively, which is above the average cloud base height for cool season cases.

![Cloud Base Height Graph](image)

**Figure 5.13.** Plot showing observed and estimated cloud base height, as well as average cloud base, and time of the onset of rain.

Fig. 5.14 provides a measurement of integrated water vapor (IWV) and integrated cloud water, the latter of which acts as a proxy to stratocumulus cloud thickness. Results indicate a slow increase in stratocumulus thickness and coverage until the approach of the QLCS at approximately 2134 UTC. Increases in IWV are indicative of the horizontal advection of water vapor in advance of the QLCS. Results from analyzing the 24 December 2008 case indicated that the subcloud lapse rate was uniform at near 5 °C/km until 2135 UTC when a tornado formed in Bell Mina in southern Limestone County. These values were derived using the lower values of subcloud lapse rate in Fig. 5.15.
Times where the lapse rate is observed to experience sudden increases results from very large discontinuity in the coverage or thickness of the low level cloud field. Moreover, periods of near uniformity were noted from 1840-1910 UTC, near 2120 UTC, and after the onset of rain. The minimum values in lapse rate of 6-7 °C/km are shown in this case. It should be noted that values exceeding 10 °C/km are not realistic, and likely associated with thin stratocumulus clouds that produce a low bias in $T_{ir}$. This case is not as well-behaved as the 30 January case due to inferred variations in cloud thickness during the daytime event. Fig. 5.16 represents the temperature difference between the 10 and 0.5 m temperature observations during the period of observation for this case. The observed temperature difference trended positive as time progressed toward tornadogenesis, indicating that the temperature near the ground was no longer warmer than at 10 m as the surface layer transitioned from unstable to stable.
Figure 5.15. Plot showing derived subcloud lapse rate averaged to five minutes based on one minute ceilometer and MPR Data.

Figure 5.16. Plot showing the difference between observed temperature at 10 and 0.5 m. Positive (negative) values represent a stable (unstable) surface layer.
3) February 25, 2011

A QLCS, oriented from 200 to 20 deg, developed over the southern Mississippi Valley at 2100 UTC and intensified as it moved eastward into middle Tennessee at 0400 UTC, as well as northern and central Alabama. This pronounced QLCS exhibited moderate slabular structure, variations in width of high Z (>40 dBZ) along the line, and significant leading-line stratiform precipitation from near the tornado location northward. This pattern was observed in the previously discussed 2013 and 2008 cases to a degree also. An EF0 tornado was produced at 0510 UTC as seen on Fig. 5.17, at a distance of 55 km and an azimuth of 150 degrees from UAH. One additional tornado report, 14 severe wind reports, and 2 severe hail reports were tabulated within 120 km of ARMOR in association with the event. Mean layer parcel CAPE for this case was <100 J kg\(^{-1}\).
UAH 2 m temperature and dewpoint observations remained steady at 21 and 14°C respectively, during the several hours prior to the passage of the QLCS. Furthermore, the 10 m wind speed averaged near 12 m s\(^{-1}\) and began a slow decline within the post-QLCS stratiform region, with wind direction shifting quickly from 190 degrees before line passage, veering to 270 degrees following passage of the line feature.

Ceilometer observations shown in Fig. 5.18 indicate a low-level stratocumulus cloud layer containing frequent breaks, which resulted in sudden spikes in cloud base height. Average minimum cloud base height was just below 1000 m, which is anomalously high when compared to the average cool season average of 650 m, and by southeast tornado event standards across all seasons. Fig. 5.19 provides a measurement of integrated water vapor and vertically integrated liquid. Results indicate steady-state conditions until the approach of the QLCS at approximately 0500 UTC. Also, the
liquid water path value remained near 4 cm during the period leading up to the approach of the QLCS, which indicates a thick blanket of stratocumulus clouds. Retrieved subcloud lapse rates displayed in Fig. 5.20 were consistently near 5 °C/km until the onset of rain at 0500 UTC. The lapse rate fluctuates between 5 and 12 °C/km. The median is near 8, and the minimum near 5 °C/km. These fluctuations are likely produced by changes in liquid water path. Slightly stable conditions were indicated by the 9.5 m temperature difference, T(10 m) – T(0.5 m) presented in Fig. 5.21. This case serves as a testament to the extremely high vertical wind shear often required to produce tornadoes in the cool season. Values of kinematic variables with this event were the highest of the three cases, in excess of 30 m s\(^{-1}\) alongside approximately 650 m\(^2\) s\(^{-2}\) of storm relative
helicity within the 0-1 km layer. Both values were in the top 5% of the distribution for the 50 cases analyzed within this study. Neither the low value of CAPE nor the relatively high cloud base height were particularly conducive for tornadoes in this case.

Results found within the three case studies presented indicate the existence of stratocumulus cloud within the three hours leading up to the tornadogenesis event. Retrieved subcloud lapse rates for each case were subadiabatic, and in the range 5-8 °C km, except for periods when T
\text{ir}
 are either sampling a higher cloud level or stratocumulus clouds with a low liquid water path. The 9.5 m temperature difference is indicative of absolutely stable conditions, which act to produce greater wind shear by reducing boundary layer turbulence.
5.2 Kinematic Results

As shown in previous sections, kinematic properties in HSLC events are strongly linked to the stability class of the boundary layer. The following kinematic results feature QLCS and supercell events, for both tornadic and null cases in order to further understanding of the kinematic conditions within the HSLC boundary layer.

5.2.1 Storm Relative Helicity

Storm relative helicity observations for this study reveal the importance of enhanced SRH produced by boundary layer stability observed in many cool season tornado events. Fig. 5.22 shows that the range of SRH values is much greater for QLCS’s than for supercells as indicated by the 25th to 75th percentile box and 10th to 90th percentile whiskers. In fact, data from this study indicates that QLCS tornadoes are the most likely when SRH (0-3 km) is greater 450 m$^2$ s$^{-2}$. 

Figure 5.21. Plot showing the difference between observed temperature at 10 and 0.5 m.
Supercell SRH depicted in Fig. 5.23 indicates that the range of values increases as the layer from which SRH is calculated increases. Results show that although the 0-500 m observations provide some ability to discriminate between tor and null events, the 0-1000 m layer offers the best discrimination. Furthermore, 0-3000 m observations provided the least separation between tornado and null cases. It was found that 0-500 m storm relative helicity provides the greatest tor/no tor discrimination among QLCS cases, with nearly no overlap among the inner quartile ranges between tornadic and null events as seen in Fig. 5.24. This result indicates the important relationship between boundary layer storm relative helicity and QLCS cases. Interestingly, 0-3000 m SRH offered the second greatest amount of discrimination. This result was not anticipated, as the 0-1000 m layer was expected to provide the second greatest discrimination.
Figure 5.23. Box and whisker plot showing VAD observed SRH for 29 supercell cases across the southeastern United States, including an additional 12 supercell null cases.

Figure 5.24. Box and whisker plot of VAD observed SRH for 21 QLCS tornado cases across the southeastern United States, including an additional 11 QLCS null cases.
5.2.2 Bulk Shear

Cold season bulk shear results shown in Fig. 5.25 indicate that supercell/QLCS values differed the most in the low levels, where tornadogenesis processes are most important. Note the strong similarity between QLCS and supercell tornado event observed bulk shear within the 0-6000 m layer. Figs. 5.26 and 5.27 include null cases for the bulk shear parameter for both QLCS and supercell events, which allows for additional discrimination between tornadic and non-tornadic environments. Once again, the greatest separation between observed tornadic and null events is the 0-500 m layer for both QLCS and supercell cases. Furthermore, bulk shear within the 0-1 km layer offers the second greatest ability to separate tornadic vs. null supercells. Discrimination decreases

Figure 5.25. Cool season bulk shear plot. Results come from the same 50 case dataset as storm relative helicity.
Figure 5.26. Box and whisker plot showing VAD observed bulk shear for 29 supercell tornado cases across the southeastern United States, including an additional 12 supercell null cases.

Figure 5.27. Box and whisker plot of VAD observed sure bulk shear for 21 QLCS tornado cases across the southeastern United States including an additional 11 QLCS null cases.
5.2.3 10 Meter Wind

The motivation for examining 10 m wind is to gain an understanding of how the near surface winds may depend on boundary layer stability. Surface wind data plotted in Fig. 5.28 indicates that tornadic supercell cases possess greater 10 m wind speed when compared to QLCS’s, which lends credence to the idea that a stable boundary layer will experience reduced turbulence. Supercell events also featured a significantly greater range of values. However, values for null events were greatest for QLCS’s, which appears to stress the importance of high storm relative helicity, stemming from large values of 0-1km shear found in tornadic QLCS environments as boundary layer turbulence is often reduced. Major outbreaks of tornadoes resulting from supercell thunderstorms occurred with a surface wind flow of 180-185 degrees with five minute average speeds of 12-13 m s\(^{-1}\). All data for this section comes from a pool of 25 tornado events, and 18 null events, for a total of 43 cases.

Figure 5.28. Plot showing 10 m wind for 25 cases, and 18 separate null cases.
10 m wind gust information displayed in Fig. 5.29 shows the strongest inner quartile range separation existing between the supercell and supercell null categories. Furthermore, discrimination exists above 18 m s\(^{-1}\) for the QLCS category, showing that tornado and null case discrimination exists above 18 m s\(^{-1}\) for the QLCS category. It is important to note that QLCS null cases possess a deeper subcloud layer, which is represented by an LCL height averaging 100 m greater than tornado producing QLCS’s as seen in Figs. 5.3 and 5.4 in section 5.1.1.

![10 Meter Wind Gust](image)

Figure 5.29. Plot showing 10 m wind gusts for 25 tornado cases, and 18 separate null cases.

The distribution of surface wind direction was found to be very similar for both tornadic supercells and QLCS’s, with the mean of each category being near 190 degrees as seen in Fig. 5.30. Both supercell and QLCS null events interestingly possessed a much greater backed southeasterly surface flow component which is likely is due to cool air wedging from a source airmass east of the Appalachian Mountains.
Figure 5.30. Plot showing 10 m wind gusts for 25 cases, and 18 separate null cases.
CHAPTER 6

DISCUSSION AND CONCLUSIONS

Cool season tornado climatology presented in this investigation indicates a storm mode consisting of 39 QLCS tornado events and 19 supercell events during the 17 week period from November 15 - March 15, from 2004-2017 within 120 km of ARMOR. Occurring primarily during the late evening and early overnight hours, fast moving QLCS events typically produce weak tornadoes and numerous instances of wind damage.

Thermodynamic results from this study indicate that significant stratocumulus cloud cover often precedes cool season tornado events within the southeastern states. This stratocumulus layer typically precedes QLCS tornadogenesis cases by an average of 200 minutes and features a cloud fraction average value of 94%. Supercell tornado events were preceded by stratocumulus cloud an average of 90 minutes, with a mean cloud fraction of 80%. Cloud base heights were slightly higher for QLCS’s (659 m) as opposed to supercells (649 m), however, a large difference was observed when comparing tornadic (649 m) and non-tornadic (993 m) supercell cases. Differences in the 0.5 to 10 m temperature difference indicate that as expected, cool season cases often feature a stable surface layer, which implies a stable boundary layer. Results shown by subcloud lapse rate retrievals among the three cases suggest that the subcloud layer is often not strongly
coupled to the cloud layer, as indicated by subadiabatic lapse rates in the range of 6-8 
°C/km.

Kinematic parameters are influenced by the degree of subcloud stability which in
turn affects the amount of boundary layer turbulence as observed among the three
presented case studies. As suspected, boundary layer kinematic variables within the 0-
500 m layer offered the most significant discrimination for tornado vs. non-tornado cases
for supercells and QLCS’s. 0-500 m supercell observed storm relative helicity averaged
190 m² s⁻², while values for QLCS’s averaged 220 m² s⁻². Moreover, 0-500 m observed
bulk shear values averaged 12 and 14 m s⁻¹ for supercells and QLCS’s respectively.
Additionally, 10 m wind flow is observed to originate from a more southeasterly
direction in association with QLCS null events, which is likely associated with the
westward progression of wedge fronts originating in southern Appalachia.

Southeastern cool season severe storm events certainly pose a distinct challenge to
forecasters, however, it is believed that these results will make a difference in improving
forecasts as well as supplying information to future researchers of this topic.
REFERENCES


