The Role of Direct Insolation and Near-Surface Moisture Advection in the Recovery of CAPE on 31 March 2016 During VORTEX-Southeast

Allison LaFleur
Purdue University

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THE ROLE OF DIRECT INSOLATION AND NEAR-SURFACE MOISTURE ADVECTION IN THE RECOVERY OF CAPE ON 31 MARCH 2016 DURING VORTEX-SOUTHEAST

by

Allison LaFleur

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In Partial Fulfillment of the Requirements for the degree of

Master of Science

Department of Earth, Atmospheric, & Planetary Sciences
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STATEMENT OF COMMITTEE APPROVAL

Dr. Robin Tanamachi, Chair
Department of Earth, Atmospheric, & Planetary Sciences

Dr. Michael Baldwin
Department of Earth, Atmospheric, & Planetary Sciences

Dr. Richard Grant
Department of Agronomy

Approved by:
Dr. Darryl Granger
Head of the Graduate Program
I dedicate this to my friends, family, mentor, and teammates who continually support me to do what I love to do and inspire me to keep learning and to always push myself to grow in all areas of my life. In particular I dedicate this to my Mom and Dad, sister Julie, and friends Tara, Rachel, and Devon.
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TABLE OF CONTENTS

LIST OF TABLES .......................................................................................................................... vi
LIST OF FIGURES ...................................................................................................................... vii
ABSTRACT .................................................................................................................................. ix

CHAPTER 1. INTRODUCTION .................................................................................................... 1
  1.1 Tornadoes and Convective Available Potential Energy ....................................................... 1
  1.2 Time Tendency of CAPE ..................................................................................................... 5
  1.3 31 March 2016 Case ........................................................................................................... 7

CHAPTER 2. METHODOLOGY .................................................................................................... 16
  2.1 Data ................................................................................................................................... 16
    2.1.1 Observations ............................................................................................................... 16
    2.1.2 Numerical model ....................................................................................................... 18
  2.2 CAPE Development ........................................................................................................... 19
    2.2.1 ATDD Soundings and CLAMPS Data ....................................................................... 19
    2.2.2 Sounding replacement analysis ............................................................................... 21
    2.2.3 Constructed soundings ............................................................................................. 23
    2.2.4 ARPS Model ............................................................................................................. 24

CHAPTER 3. DISCUSSION AND CONCLUSIONS .................................................................. 31

APPENDIX .................................................................................................................................. 33

REFERENCES ............................................................................................................................. 38
LIST OF TABLES

Table 1: Surface-based, mixed layer, and most unstable CAPE and CIN (in J kg⁻¹) from three soundings taken at Belle Mina, Alabama on 31 March 2016. ........................................ 9
Table 2: UMass FMCW instrument characteristics (Ince et al. 2003). ........................................ 17
Table 3: Meteorological instruments with their corresponding instrument and sampling height from the instruments on the meteorological tower in Belle Mina, AL (Lee et al 2016). 18
Table 4: Calculated CAPE rates, in J kg⁻¹ 5 min⁻¹ from 2200 UTC on 31 March 2016 to 0200 UTC on 1 April 2016 from the ARPS simulations using equation 2. X represents when the calculated value was not available. ................................................................. 33
LIST OF FIGURES

Figure 1: 0.5° Reflectivity (in dBZ) from Huntsville-Hytop radar (KHTX) at 1329 UTC on 31 March 2016 depicting the morning storms................................................................. 8

Figure 2: Same as Fig. 1, but at 0102 UTC on 1 April 2016, depicting the evening tornadic storms.................................................................................................................. 8

Figure 3: Reflectivity factor (in dBZ) observed by the FMCW radar during the morning rain from (a) 1600-1659 UTC and (b) 1700-1759 UTC. Reflectivity maxima appearing at constant altitude in time (e.g., at 3.9 km AGL) are artifacts of power spurs in the UMass FMCW.................................................................................................................. 10

Figure 4: 0.5° Reflectivity (in dBZ) from KHTX at (a) 1900 UTC, (b) 2101 UTC, (c) 2200 UTC, (d) 2308 UTC, and (e) 2354 UTC on 31 March 2016, and (f) 0041 UTC, (g) 0204 UTC, and (h) 0300 UTC on 1 April 2016.54 UTC on 31 March 2016, and (f) 0041 UTC, (g) 0204 UTC, and (h) 0300 UTC on 1 April 2016.............................................. 11

Figure 5: 0.5° Velocity from (a) KGWX in northwest Alabama with low level rotation circled in gold and (b) KTHX in northern Alabama with rotation circled in gold. ...................... 13

Figure 6: Surface analysis from 2207 UTC in the Southeast United States (courtesy of Plymouth State University). Temperatures and dew points are in °C; pressure is shown in hPa, and wind barbs are shown in knots. The VORTEX-Southeast domain is outlined in purple. .............................................................................................................................. 15

Figure 7: Shortwave radiation data from a 2.5 m AGL meteorological tower at Belle Mina, Alabama. Data from Lee et al. (2016a)............................................................................ 20

Figure 8: Temperature data from 3 m AGL meteorological tower at Belle Mina, Alabama. Data from Lee et al. (2016a). ........................................................................................................... 20

Figure 9: FMCW reflectivity from 31 March 2016 2000 UTC to 2200 UTC with the extended Kalman filter applied to locate boundary layer heights (magenta crosses). ............... 21

Figure 10  23

Figure 11: CAPE (in J kg-1) at the Belle Mina, Alabama from 2000 to 2230 UTC from the soundings reconstructed from CLAMPS and the ATDD sounding data. ................. 24

Figure 12: INSERT CAPTION OF MODEL PICS ................................................................................................................................. 25
Figure 13: Plot of the moisture advection at Belle Mina, Alabama calculated from ARPS
simulations at (a) a single time, 2000 UTC on 31 March 2016, and (b) from 2000
UTC to 0200 UTC 1 April 2016. The cause of the enhanced moisture advection at all
levels at 2130 UTC is unknown................................................................. 26

Figure 14: SB CAPE values calculated from the ARPS simulations (green curve), ATDD
soundings (yellow curve), and constructed ATDD/CLAMPS soundings from 2000
UTC on 31 March 2016 to 0200 UTC on 1 April 2016................................. 27

Figure 15: Each of the terms in the time tendency of CAPE equation (2), calculated from the
ARPS simulations at the grid point closest to Belle Mina.............................. 28

Figure 16: Plot of the each of the calculated terms in term A, the boundary layer entropy term, of
time tendency of CAPE equation from the ARPS simulations.......................... 30
ABSTRACT

Author: LaFleur, Allison, T. MS
Institution: Purdue University
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Title: The Role of Direct Insolation and Near-Surface Moisture Advection in the Recovery of CAPE on 31 March 2016 During VORTEX-Southeast.
Committee Chair: Dr. Robin Tanamachi

The relative roles of direct insolation and near-surface moisture advection in the recovery of convective available potential energy (CAPE) on 31 March 2016 in northern Alabama are examined using VORTEX-Southeast observations and numerical simulations. In between rounds of nontornadic morning storms and tornadic evening storms, CAPE over the VORTEX-Southeast domain increased from near zero to at least 500 J kg\(^{-1}\). A timeline of the day’s events is provided with a focus on the evolution of the lower levels of the atmosphere. In particular, we focus on its responses to solar insolation and moisture advection, which are hypothesized as the main mechanisms behind the recovery of CAPE. Data from the University of Massachusetts S-Band frequency-modulated, continuous-wave (FMCW) radar and NOAA National Severe Storms Laboratory (NSSL) Collaborative Lower Atmospheric Mobile Profiling System (CLAMPS) are used to characterize the boundary layer evolution in the pretornadic storm environment. It is found that insolation-driven diabatic heating was the primary driver of rapid CAPE recovery on this day.
CHAPTER 1. INTRODUCTION

1.1 Tornadoes and Convective Available Potential Energy

Convective available potential energy (CAPE), a commonly used parameter in tornado forecasting, is calculated using an atmospheric sounding. It is defined as maximum buoyancy of an undiluted air parcel, related to the potential updraft strength of thunderstorms (American Meteorological Society 2017). On a sounding diagram, CAPE is the sum of the area between the temperature of the environment and the temperature a parcel would take when lifted from the surface, or some predetermined layer in the atmosphere, to its equilibrium level (Dowsell and Rasmussen 1994). The equilibrium level is the level of the atmosphere where a parcel reaches neutral buoyancy and is the same density as the surrounding air. Mathematically CAPE is represented as

\[ CAPE = \int_{p_f}^{p_n} R_d (T_{vp} - T_{ve}) d\ln p \]  

(1)

where \( p_f \) is the pressure at the level of free convection, and \( p_n \) is the pressure at the equilibrium level, \( R_d \) is the specific gas constant for dry air, \( T_{vp} \) is the virtual temperature of a lifted parcel moving upward moist adiabatically from the level of free convection to the equilibrium level, and \( T_{ve} \) is the virtual temperature of the environment (American Meteorological Society 2017). Higher CAPE indicates a more unstable atmosphere that will support stronger updrafts, and therefore stronger supercells. There are different CAPE calculations depending on which starting parcel is used to calculate CAPE. The most commonly used varieties of CAPE are those from surface based (SB), the mean layer (ML), and most unstable (MU) parcels.
Tornadoes in the Southeast United States can occur under different conditions than are typically seen in the Great Plains. For instance, severe storms have been observed to form when analyzed CAPE is less than 500 J kg\(^{-1}\), as opposed to the 1000 J kg\(^{-1}\) that is generally accepted as needed for tornadic storms to develop (Sherburn and Parker 2014). In other words, these storms are able to form and persist in environments with low CAPE. It has also been hypothesized that the low CAPE values arise from inaccurate analysis, or that the environment generates regions of high CAPE which are smaller than current mesoscale analysis can resolve (Rasmussen 2015) which is about 40 km (Hart et al. 2016). These analysis errors are can occur anywhere, but the southeast sees more of these low CAPE storms, particularly in the form of quasi-linear convective systems (QCLS) (Thompson et al 2013). In the southeast United States, low CAPE storms are most common during the cold season, or at night, which are the same times false alarm rates are the highest for tornado forecasts in this area (Sherburn and Parker 2014).

Previous case studies (Johns 1993; Lane and Moore 2006; Clark 2009; Evans 2010) found that there is likely a dry intrusion aloft in low CAPE storm environments. This layer is hypothesized to coincide with the start and/or strengthening of convection and the release of instability (Johns 1993; Lane and Moore 2006; Clark 2009; Evans 2010).

Modeling studies show that robust convection is possible when the instability is concentrated in the lowest levels in these weak CAPE environments (Kirpatrick et al 2011; McCaul and Weisman 2001, Sherburn et al. 2016). In other words, when the parcel reaches its maximum buoyancy at lower levels stronger updraft accelerations are promoted, because more CAPE is available to the low-level updraft than when the same amount of buoyancy is spread throughout the depth of the atmosphere (Kirpatrick et al 2011). Additionally, because stronger updraft accelerations are promoted in lower levels, there is also enhanced stretching of
preexisting vorticity at these levels (Kirpatrick et al. 2011). A study by Kirkpatrick et al. (2011) examined how storms behaved in environments with different amounts of CAPE and precipitable water. More recently, studies have focused on identifying radar signatures within these low CAPE storms (McAvoy et al. 2000; Grumm and Glazewski 2004; Barker 2006; Clark 2011). Grumm and Glazewski (2004) looked to identify “Broken S” signatures in radar data to help identify storms to watch. These signatures occur in low-topped mesoscale convective systems and have a similar appearance to a line echo wave pattern (Grumm and Glazewski 2004). Barker (2006) examined reflectivity tags and bulges, and Clark (2011) examined the relationship between the path length of rotating storms (identified by updraft helicity swaths) in model data to the observed path lengths of tornadoes. Clark (2011) identified and discussed the benefit of using convection allowing models (CAMs) to determine the convective mode and hazards of a storm, rather than just the storm environment. However, that study did point out challenges that require further study and research. CAMs do not fully resolve some meteorological hazards associated with storms. For example, models can reproduce storms and mesocyclones that lead to tornadoes, but individual tornadoes are too small to resolve on most operational model grids (Clark 2011). Kirkpatrick et al. (2011) also asserted that using environmental parameters from models to predict tornadoes is difficult, and often these modeled environmental parameters have large uncertainty as well.

These low-CAPE studies mention that more research is needed to fully understand the environment in which these low CAPE storms develop. In this study, we propose a method to examine various sources of CAPE in the environment, and time tendency of CAPE, and in turn the time tendency of the difference sources, to see if one source predominates.
VORTEX-Southeast was a field program to study tornadoes and tornado environments in the southeast United States (Rasmussen 2015). The field observations were collected on a domain centered on Huntsville, Alabama, that encompassed roughly the northern third of Alabama. VORTEX-Southeast looked to understand the environments of the southeast United States that influence the structure and path of the region’s tornadoes, as well as how to best communicate warnings of these storms to the public (Rasmussen 2015). The tornadoes in this area often occur at night, in forested areas, outside of the perceived tornado season. Additionally, they often occur in areas with limited visibility, inadequate shelter, and a large population. These factors are believed to contribute to the disproportionately large number of killer tornadoes in this area (Ashley 2007).

As mentioned previously, the southeast United States sees frequent storms in low-CAPE environments (Sherburn and Parker 2014). One goal of VORTEX-Southeast is to ascertain whether local values of CAPE were larger than operational analyses suggest, or whether CAPE grew large on short time scales owing to rapid destabilization of the environment (King et al. 2017). To study this phenomenon, it was hypothesized that tornadic thunderstorms are strongly forced, such as by cold fronts or low level jets, and have upper tropospheric support for quasigeostrophic (QG) ascent. This hypothesis was assessed by evaluating the rate of convective destabilization (Rasmussen 2015), using VORTEX-Southeast observations (Koch and Rasmussen 2016) to examine relative roles of direct insolation and near-surface moisture advection. Many instruments were deployed across the VORTEX-Southeast domain during its special observation campaigns in 2016 and 2017. Since the VORTEX-Southeast domain was near the Gulf of Mexico, moisture advection from the ocean was hypothesized to be a typical contributor to increased instability.
In this study, we look to develop a way to assess CAPE tendency from observations in order to compliment forecast models. To do so, we will analyze the CAPE development over the course of a particular case. We then break down the development of CAPE from individual sources to determine which source, or combination of sources, leads primarily to the destabilization of the atmosphere.

1.2 Time Tendency of CAPE

In order to examine the different factors of the buildup of CAPE over the course of the afternoon, we utilized the equation for the time tendency of CAPE (Emanuel 1994):

\[
\frac{dCAPE}{dt} = \left(T_{LCL} - T_{EL}\right) \frac{dS_b}{dt} - \int_{p_{EL}}^{p_{LCL}} \frac{R_d Q}{C_p} \left(\frac{T_v}{T}\right) - R_d T_v \frac{v_r \cdot \nabla \theta}{\theta} d\ln P \tag{2}
\]

where \(T_{LCL}\) is the temperature at the lifting condensation level (LCL), which is the level a parcel of moist air, that is lifted dry-adiabatically, would become saturated (American Meteorological Society 2017). \(T_{EL}\) is the temperature at the equilibrium level (EL), \(\frac{dS_b}{dt}\) is the change in parcel entropy in the subcloud layer, \(R_d\) is the gas constant of dry air \((287.058 \text{ J kg}^{-1} \text{ K}^{-1})\), \(T_v\) is the virtual temperature, \(Q\) is the radiative cooling rate of the free atmosphere, \(v_r\) are the winds relative to the parcel, and \(\theta\) is potential temperature.

In equation (2), term A is the boundary layer (BL) entropy term. This term can be rewritten as

\[
(1 - \frac{T_{EL}}{T_o}) C_p \frac{dT}{dt} \tag{3}
\]
as shown by Humphrey and Bosart (2011). $T_0$ is the starting temperature of the parcel, $C_p$ is the specific heat capacity at a constant pressure, and $\frac{dT}{dt}$ is the heating rate of the subcloud layer. This heating can be a result of diurnal heating, horizontal temperature advection, adiabatic warming of descending air, or the combination of the three (Humphrey and Bosart 2011). This alternative form of the equation was used to calculate the change in CAPE from the BL entropy. Increasing boundary layer entropy increases instability in this layer.

Term B is the radiative heating term. Radiative heating is only calculated from the LCL to the EL, meaning it is only calculating radiative heating or cooling in the cloud layer. Cooling this layer increases instability.

All of the variables, except $Q$, were available from the model directly. To calculate $Q$, we solved the thermodynamic equation,

$$\frac{\partial T}{\partial t} = -\bar{v} \cdot \nabla T + S_p \omega + \frac{Q}{C_p}$$

(4)

where $-\bar{v} \cdot \nabla T$ represents temperature advection, $S_p$ is a static stability parameter, $\omega$ is vertical velocity. $S_p$ was calculated as

$$S_p = \frac{\alpha}{C_p} - \frac{\partial T}{\partial p}$$

(5)

where $\alpha$ is defined as specific density, $\frac{1}{\rho}$, where $\rho$ is air density.

Term C, the advection term, also calculated from the LCL to the EL (i.e., in the cloud layer), represents the effects of temperature advection. Cooling this layer will increase the lapse rate, which will in turn increase CAPE.
1.3  31 March 2016 Case

We seek to develop a theory-based methodology to assess CAPE tendency using conventionally available observations. As a case study, we look to understand what allowed the northern Alabama tornadic storms of 31 March 2016 to form in what was believed to be a low CAPE environment. On this day, morning storms depleted most of the CAPE, but owing to rapid destabilization in the afternoon, tornadic storms were still able to form in the evening. This case occurred during an intensive observing period (IOP) of the Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX)-Southeast 2016 field campaign (Rasmussen and Koch 2016). In between the morning storms (Fig. 1) and the evening tornadic storms (Fig. 2), a large amount of meteorological data were collected in northern Alabama as the CAPE over the VORTEX-Southeast domain increased by at least 500 J kg\(^{-1}\) and possibly as much as 1282 J kg\(^{-1}\) (Table 1). Scale analysis of the Emanuel (1994) equation (shown in appendix), as well as recent work by Agard and Emanuel (2017), suggest that diabatic heating and moisture advection, at the low levels, are the primary drivers of CAPE buildup in severe storm environments. We hypothesize that direct insolation and near-surface moisture advection were the dominant mechanisms in the rapid recovery of CAPE over northern Alabama on 31 March 2016, which then allowed for severe storms to develop.
Figure 1: 0.5° Reflectivity (in dBZ) from Huntsville-Hytop radar (KHTX) at 1329 UTC on 31 March 2016 depicting the morning storms.

Figure 2: Same as Fig. 1, but at 0102 UTC on 1 April 2016, depicting the evening tornadic storms.
Table 1: Surface-based, mixed layer, and most unstable CAPE and CIN (in J kg⁻¹) from three soundings taken at Belle Mina, Alabama on 31 March 2016.

<table>
<thead>
<tr>
<th>Time UTC</th>
<th>20</th>
<th>21</th>
<th>22</th>
</tr>
</thead>
<tbody>
<tr>
<td>SBCAPE</td>
<td>119</td>
<td>1282</td>
<td>558</td>
</tr>
<tr>
<td>SBCIN</td>
<td>-102</td>
<td>0</td>
<td>-16</td>
</tr>
<tr>
<td>MLCAPE</td>
<td>61</td>
<td>149</td>
<td>367</td>
</tr>
<tr>
<td>MLCIN</td>
<td>-120</td>
<td>-62</td>
<td>-37</td>
</tr>
<tr>
<td>MUCAPE</td>
<td>592</td>
<td>1282</td>
<td>668</td>
</tr>
<tr>
<td>MUCIN</td>
<td>-1</td>
<td>0</td>
<td>-10</td>
</tr>
</tbody>
</table>

On 31 March 2016, a VORTEX-Southeast Intensive Operation Period (IOP3) was declared in anticipation of convective storms in the domain. Non-severe morning convection was expected in association with little to no low-level CAPE, but in the afternoon the environment was expected to support supercells due to rapid destabilization and moderate low-level wind shear (NWS Storm Prediction Center 2016). The focus of VORTEX-Southeast field observations was the rapid northward advection of warm, humid, near-surface air as a morning convective system departed, in expectation of stronger convection in the evening after the return of unstable air (Rasmussen 2016). What follows is a timeline of the day’s events.

Widespread non-severe storms moved through the VORTEX-Southeast domain early in the morning on 31 March. These storms were part of a mesoscale convective system that moved through Alabama and Tennessee that morning. Fig. 3a depicts the rainfall as seen by the
vertically-pointing UMass FMCW radar at Belle Mina, Alabama. This morning rain and its associated cool outflow stabilized the boundary layer over northern Alabama (Fig. 3b).

After the cessation of rainfall, at Belle Mina, around 1650 UTC (Fig. 3a), the convective boundary layer began to redevelop, reaching a depth of about 500 m by 1740 UTC (Fig. 3b) and 1 km by 2000 UTC (not shown). This development was evident in the refractive index turbulence detected by the FMCW (Fig 3b).

![Figure 3a](image1.png)

![Figure 3b](image2.png)

Figure 3: Reflectivity factor (in dBZ) observed by the FMCW radar during the morning rain from (a) 1600-1659 UTC and (b) 1700-1759 UTC. Reflectivity maxima appearing at constant altitude in time (e.g., at 3.9 km AGL) are artifacts of power spurs in the UMass FMCW.
Figure 4: 0.5° Reflectivity (in dBZ) from KHTX at (a) 1900 UTC, (b) 2101 UTC, (c) 2200 UTC, (d) 2308 UTC, and (e) 2354 UTC on 31 March 2016, and (f) 0041 UTC, (g) 0204 UTC, and (h) 0300 UTC on 1 April 2016.
At 1900 UTC, thunderstorms developed over north central Mississippi (Fig. 4a). These storms were more discrete in nature when compared to the morning convection, owing to increased deep-layer wind shear. These storms moved rapidly to the northeast (Fig. 4b), with some exhibiting weak rotation (not shown). They then moved into air that was generally considered too stable to support low-level rotation. By 2100 UTC, the NWS Storm Prediction Center (SPC) released Mesoscale Discussion #311 about these cells and their surrounding environment (NWS Storm Prediction Center 2016). According to SPC forecasters, the morning convection left a large-scale outflow boundary over northern Mississippi and central Alabama, an area which now exhibited strong speed shear in the mid- and upper levels, and 0-to-1 km bulk shear around 200 m$^2$ s$^{-2}$. On the southwest side of the boundary, surface winds were from the southeast, skies were clear, temperatures were in the mid to upper 20s and dewpoints were in the upper teens to lower 20s in degrees Celsius. Northeast of the boundary, skies were cloudy, temperatures were in the low to mid 20s, and dewpoints were in the mid to upper teens in degrees Celsius.
Figure 5: 0.5° Velocity from (a) KGWX in northwest Alabama with low level rotation circled in gold and (b) KTHX in northern Alabama with rotation circled in gold.
At 2200 UTC, storms with persistent low-level rotation had reached the Alabama-Tennessee border (Figs. 4c, 5a), despite surface conditions being relatively cool and stable. Fig. 6 shows the surface analysis at this time. At 2300 UTC, the NWS SPC issued Tornado Watch #72 over northern Alabama and south-central Tennessee, citing “supercells capable of hail, locally damaging winds, and tornadoes expected to develop and move eastward to east-northeastward across the area overnight” (NWS Storm Prediction Center 2016). The SPC released two more mesoscale discussions over the next hour, at 2352 UTC on 31 March 2016 (#315) and 0041 UTC (#317) on 1 April 2016, respectively, mentioning developing discrete cells on the Alabama-Mississippi and Alabama-Georgia borders, respectively (NWS Storm Prediction Center 2016). The cells had developed into a mix of discrete supercells and clusters (Fig. 4d-f), supporting a tornado risk across northern Alabama (NWS Storm Prediction Center 2016). The environment was characterized by backing and strengthening winds above 1 km MSL, 0-1 km storm relative helicity (SRH) of 250 to 300 m$^2$ s$^{-2}$, surface dew points generally increasing from 17 to 21 °C from the Tennessee border southward across northern Alabama (Fig. 6), and mixed layer CAPE (MLCAPE) of 1000 J kg$^{-1}$. Low-level warm air advection (WAA) was also present closer to the Alabama-Georgia border.
Over the next few hours, the northern Alabama storms (Fig. 4d-f) continued to exhibit low-level rotation (not shown). These storms moved over the VORTEX-Southeast domain, generating a tornado near Priceville, Alabama at 0200 UTC (Fig. 4g). The tornado touched down just northeast of Hartselle, Alabama and moved northeast, dissipating just northeast of Priceville, Alabama. Velocity data from the KTHX radar depict these storms in figure 5b. The tornado was rated an EF-2 on the Enhanced Fujita scale with maximum winds of 69 m s⁻¹. Its track was approximately 14 km long and 180 m wide (NWS 2016). VORTEX-Southeast field observations ceased at 0300 UTC, when the storms exited the Huntsville domain (Fig. 4h).
CHAPTER 2. METHODOLOGY

2.1 Data

To diagnose the development of instability over the course of the afternoon, we utilized observations and model data. First of all, we simply examined how CAPE values changed over the course of the afternoon. This was done with soundings, radar data, interferometer data, and a meteorological tower. Then to further diagnose what resulted in the increasing instability in the area we used model data and equation (2) to determine what layer of the atmosphere and what mechanisms lead to the increase of CAPE.

To evaluate Emanuel (1994)’s equation (2), we need a number of meteorological variables. These include temperature $T$, the LCL height, EL height, potential temperature $\theta$, winds to calculate advection $\mathbf{v} \cdot \nabla \theta$, entropy values $S$, heating rates $Q$, pressure $P$, and virtual temperature $T_v$.

2.1.1 Observations

Radiosonde soundings, the University of Massachusetts (UMass) S-Band frequency-modulated, continuous-wave (FMCW) radar (Ince et al. 2003), and the Collaborative Lower Atmospheric Mobile Profiling System (CLAMPS) (Geerts et al 2016), which were collocated at Belle Mina, Alabama, were used as primary sources of thermodynamic and precipitation-related variables for this study. The UMass FMCW is a vertically-pointing, S-band mobile radar with 2.5 m vertical resolution. It has a pair of 2.4 m parabolic dish antennas with 34 dB of gain (Ince et al. 2003). Table 2 has further specifications of the instrument. The Atmospheric Emitted Radiance Interferometer (AERI; Blumberg et al. 2015), the principal instrument of the CLAMPS, is an operational ground-based spectrometer that measures the downwelling infrared
(3–19 µm) radiance emitted by the atmosphere at a high temporal and spectral resolution (Blumberg et al 2015).

Table 2: UMass FMCW instrument characteristics (Ince et al. 2003).

<table>
<thead>
<tr>
<th>Description</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Center frequency</strong></td>
<td>2.94 GHz</td>
</tr>
<tr>
<td><strong>Bandwidth</strong></td>
<td>60 MHz</td>
</tr>
<tr>
<td><strong>Transmit Power</strong></td>
<td>250 W</td>
</tr>
<tr>
<td><strong>Sweep Time</strong></td>
<td>45 ms</td>
</tr>
<tr>
<td><strong>Noise Figure</strong></td>
<td>2 dB</td>
</tr>
<tr>
<td><strong>Range Resolution</strong></td>
<td>2.5 m</td>
</tr>
<tr>
<td><strong>Height Coverage</strong></td>
<td>2.5 km</td>
</tr>
<tr>
<td><strong>Antenna Gain</strong></td>
<td>34 dB</td>
</tr>
<tr>
<td><strong>Antenna Beam Width</strong></td>
<td>3°</td>
</tr>
<tr>
<td><strong>Polarization</strong></td>
<td>Single, linear</td>
</tr>
</tbody>
</table>

Data from the NOAA Air Resources Laboratory’s Atmospheric Turbulence and Diffusion Division (ATDD) were also utilized in this study. During VORTEX-Southeast IOPs, ATDD personnel released GRAW DFM-09 radiosondes with 1 second temporal resolution from Belle Mina, Alabama. On 31 March 2016, soundings were launched hourly at and after 2000 UTC (Lee et al 2016). ATDD also had a meteorological tower at Belle Mina. This tower collected meteorological, soil, and flux observations. Table 3 provides the specifications for the instruments on the meteorological tower (Lee et al 2016). The tower sampled data every 5 seconds and reported the 30-minute mean, except for the sonic anemometers and gas analyzers that sampled at 10 Hz, or every 0.1 seconds (Lee et al 2016).
Table 3: Meteorological instruments with their corresponding instrument and sampling height from the instruments on the meteorological tower in Belle Mina, AL (Lee et al 2016).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Instrument</th>
<th>Sampling Height(s) (m above ground level [agl])</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature, dew point temperature</td>
<td>Vaisala HMP110 humidity and temperature probe</td>
<td>1.5</td>
</tr>
<tr>
<td>Temperature</td>
<td>Platinum resistance thermometer (PRT) in aspirated shield</td>
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<td>RM Young 61302V</td>
<td>1</td>
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<tr>
<td>Net radiation</td>
<td>Hukseflux 4-component net radiometer</td>
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<tr>
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<td>TP01 soil temperature probe</td>
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<tr>
<td>Soil moisture</td>
<td>Vegetronix</td>
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<tr>
<td>Wind speed, wind direction</td>
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<td>3, 10</td>
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<tr>
<td>Rainfall</td>
<td>TB3 tipping bucket rain gauge</td>
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<tr>
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<td>EC155 closed path infrared gas analyzer</td>
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<td>CO₂ flux</td>
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<td>Sensible heat flux</td>
<td>CSAT3 sonic anemometer</td>
<td>3, 10</td>
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<tr>
<td>Photosynthetically active radiation (PAR)</td>
<td>LI-190</td>
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</table>

Soundings were used to directly calculate CAPE using SHARPpy (Blumberg et al. 2017).

The meteorological tower provided surface temperature $T$, dewpoint $T_d$, and net radiation $R$. The CLAMPS system also provided boundary layer profiles of temperature $T$ and dewpoint temperature $T_d$ that were used to calculate CAPE at times intermediate to those of the soundings.

2.1.2 Numerical model

The Advanced Regional Prediction System (ARPS; Xue et al. 2000) was used to model the storms on 31 March 2016. The ARPS is a nonhydrostatic numerical cloud model designed to be used on the regional to storm scale (Xue et al 2000). The model was run at 6 km grid spacing, nested within the 1200 UTC The North American Mesoscale Forecast System (NAM), starting from 1200 UTC on 31 March and ending 0300 UTC on 1 April. The Yonsei University Planetary Boundary Layer Scheme (YSU PBL) parameterization and the triple-moment version of the NSSL precipitation microphysics scheme were used.
Hereafter, all of the terms in the time tendency of CAPE equation (2) were calculated from model data, and these results were compared to what was observed (Section 2.1.i).

2.2 CAPE Development

2.2.1 ATDD Soundings and CLAMPS Data

Following the morning rain on 31 March, CAPE values grew to over 1200 J kg\(^{-1}\), as evidenced by hourly radiosonde soundings (Lee et al. 2016b) taken at Belle Mina, Alabama (where UMass FMCW and CLAMPS were collocated). SHARPpy (Blumberg et al. 2017) was used to calculate the instability parameters from these soundings (Table 1). It can be seen that the greatest SBCAPE increase occurred between 2000 UTC and 2100 UTC (Table 1). During that hour, shortwave solar insolation increased over the Belle Mina site (Fig. 7), resulting in more surface heating over the area (Fig. 8). UMass FMCW reflectivity observations (Fig. 9) show that the boundary layer deepened slightly and became more convectively active during this period from 2000 UTC to 2100 UTC. An extended Kalman filter (Lange et al. 2015) was used to objectively identify and track the boundary layer heights over this afternoon, in one hour increments. An extended Kalman filter is a non “memoryless” algorithm. It uses both the current profile of the boundary layer and the past boundary layer heights to determine the current boundary layer height. After the data is pre-processed a measurement model is used that models the mixed layer- free troposphere interface. From there the boundary layer height is extracted. From the EKF algorithm, it was found that boundary layer height reached 600 m by ~2130 UTC. An example of this filter applied to data is shown in Fig. 9.
Figure 7: Shortwave radiation data from a 2.5 m AGL meteorological tower at Belle Mina, Alabama. Data from Lee et al. (2016a).

Figure 8: Temperature data from 3 m AGL meteorological tower at Belle Mina, Alabama. Data from Lee et al. (2016a).
2.2.2 Sounding replacement analysis

We performed an initial “sounding replacement” analysis to estimate the amount CAPE we would expect to see from the increased heating versus increasing moisture. For the purposes of this preliminary analysis, it was assumed that solar insolation (Fig. 7) was the only factor contributing to the change in temperature (Fig. 8), while moisture advection was the only factor contributing to the change in the dewpoint profile. These assumptions were made to get an initial idea of CAPE recovery mechanisms, and should not be considered representative of the actual CAPE recovery that occurred.
Our “sounding replacement” analysis proceeded as follows. The temperature profile from the 2000 UTC ATDD sounding at Belle Mina (Fig. 10a) was replaced with the corresponding temperature profile from the 2100 UTC sounding, with the remaining variables unchanged. CAPE was then recalculated to estimate the changes in CAPE values due to insolation. Using this technique, we found that solar insolation should have increased CAPE values by approximately 1822 J kg\(^{-1}\). To estimate the increase in CAPE due solely to moisture advection, this process was repeated on the dew point profile in the 2000 UTC ATDD sounding. It was found that replacing the moisture profile in the 2000 UTC ATDD sounding with that from the 2100 UC sounding increased CAPE by approximately 141 J kg\(^{-1}\). The observed CAPE increase was 1282 J kg\(^{-1}\) (Table 1), signifying that some other process or processes were at work to modulate the changes in CAPE, and that additional analysis was needed.
2.2.3 Constructed soundings

While hourly soundings are useful, they are still too coarse temporally to resolve rapid changes in CAPE. To examine how CAPE changed on a sub-hourly scale, the CLAMPS temperature and moisture profiles, which extended from the surface to 4 km AGL, were combined with the ATDD soundings above 4 km AGL to create constructed soundings every five minutes from 2000 UTC to 2230 UTC. Additionally, the ATDD soundings were temporally interpolated to five-minute intervals from 2000 to 2230 UTC. This was done to match the
temporal resolution of the CLAMPS data, and to avoid large discontinuities in CAPE resulting from sudden changes in the profile above 4 km AGL. From these constructed soundings, CAPE and other parameters were calculated at 5-minute intervals over the period. It was found that CAPE increased from about 400 J kg\(^{-1}\) to 1100 J kg\(^{-1}\) from 2000 to 2230 UTC (Fig. 11), depending on which CAPE calculation is performed. The SB parcel inherits its initial characteristics from the surface measurements, the ML parcel uses the mean conditions in the lowest 100 mb of the atmosphere, and the MU parcel is the most unstable parcel found in the lowest 300 mb of the atmosphere.

![CAPE Values](image)

Figure 11: CAPE (in J kg\(^{-1}\)) at the Belle Mina, Alabama from 2000 to 2230 UTC from the soundings reconstructed from CLAMPS and the ATDD sounding data.

2.2.4 ARPS Model

To examine moisture advection more closely, we used the Advanced Regional Prediction System (ARPS, Xue et al. 2000) model to simulate the weather conditions from 12 UTC on 31 March 2016 to 0300 UTC on 1 April 2016. These simulations were run with 6 km horizontal grid
spacing, and initial conditions nested inside the 1200 UTC NAM. These simulations showed southerly winds and increasing equivalent potential temperature ($\theta_e$) over northern Alabama over the course of the afternoon (not shown) suggesting higher temperatures and/or dewpoints over the region are being advected north, possibly as far as Belle Mina. The simulations also captured the morning rainfall event, and convective initiation later in the afternoon. From the simulations, we found low-level moisture (below 500 m) at the Belle Mina site did not increase much during this time (Fig 13a, b). However, at altitudes at and above 500 m, there were alternating layers of drying and moistening. These alternating layers persisted throughout the time period (Fig. 13b). These alternating layers of drying and moistening were also evident in each of the soundings (Fig. 10a-c) and the plot of the moisture advection (Fig. 13) calculated from simulations of the day’s weather.

Figure 12: Model output from the (a) 13:30 UTC, (b) 1600 UTC, (c) 1900 UTC, and (d) 2100 UTC. The background shading in $\theta_e$, the arrows wind in m s$^{-1}$, and the contours are reflectivity.
Figure 13: Plot of the moisture advection at Belle Mina, Alabama calculated from ARPS simulations at (a) a single time, 2000 UTC on 31 March 2016, and (b) from 2000 UTC to 0200 UTC 1 April 2016. The cause of the enhanced moisture advection at all levels at 2130 UTC is unknown.

We also calculated CAPE from the model by creating soundings from the model data (Fig. 14s). The order of magnitude of model CAPE is comparable to the observed CAPE values, but the increase in CAPE happens more abruptly and later in time (2300 UTC) than in the observations. To examine the drivers of CAPE in more detail, the time tendency of CAPE equation (2) was utilized. To calculate the terms of the equation, we used data from the model.
and plugged it into each term of the equation individually. The value of each of these terms was calculated for each five-minute interval from 2000 UTC on 31 March to 0200 UTC on 1 April (Fig. 15). At the time of the sharp increase in modeled CAPE (2300 UTC), the changes in CAPE due to radiational cooling boundary layer entropy were much larger than other times in the model, and the advection term was generally slightly negative (Fig. 15). It is inferred that, within the model, there was a great deal of heating in the lowest levels of the atmosphere, and cooling aloft. Interestingly enough, CAPE calculated from the reconstructed soundings reaches its peak at a time (2230 UTC) intermediate between the ATDD soundings (2100 UTC) and the model (2300 UTC) (Fig. 14). It is hard to conclusively say why this happens, particularly because our last ATDD sounding was at 2200 UTC, so we do not see if the CAPE increased again around 2300 UTC.

![CAPE on 31 March from Different Measurements](image)

Figure 14: SB CAPE values calculated from the ARPS simulations (green curve), ATDD soundings (yellow curve), and constructed ATDD/CLAMPS soundings from 2000 UTC on 31 March 2016 to 0200 UTC on 1 April 2016.
Figure 15: Each of the terms in the time tendency of CAPE equation (2), calculated from the ARPS simulations at the grid point closest to Belle Mina.

It appears that the greatest source of CAPE was boundary layer entropy, or the subcloud layer (Fig. 15). We sought to break down this term (A) further in order to figure out what sources contributed to the boundary layer instability that day.

The static stability portion of term A (Eqn. 2) can be split up further as follows:

\[
\frac{dS_i}{dt} = F_s + \frac{Q_b}{T_b} \frac{\Delta z_b}{\Delta z_b}
\]

where \(F_s\) is the surface flux of entropy, \(Q_b\) is the radiative heating of the subcloud layer, \(T_b\) is the temperature of the subcloud layer, and \(\Delta z_b\) is the subcloud layer depth. We can calculate \(Q\) using the thermodynamic equation (Eqn. 4). This allows us to calculate the last term in the equation.
leaving the surface flux of entropy term as a residual. It is possible for us to retrieve the subcloud
layer depth from the FMCW data but we cannot calculate F_s with available data. This is
something to be done in the future.

Using the alternate form of term A (Eqn. 3) $\frac{dT}{dt}$ can be split up the same way as done in
term B, again utilizing the thermodynamic equation (Eqn. 3). To do this we start with Eqn. 3

$$
(1 - \frac{T_{El}}{T_o}) C_p \frac{dT}{dt}
$$

and replace $\frac{dT}{dt}$ with the thermodynamic equation, Eqn. 4 to yield

$$
(1 - \frac{T_{El}}{T_o}) C_p (-\vec{v} \cdot \nabla T + S_p \omega + \frac{Q}{C_p}) \quad (7)
$$

A1        A2      A3

We can directly calculate the advection and vertical terms and leave Q as the residual.

We used this method to split up the boundary layer term, as it allowed us to examine more
factors that contributed to the increase in CAPE in this layer. Fig. 16 depicts the results of this
analysis. The residual is taken to be the heating rate, as it is the only term we could not directly
calculate. Advection appears to be the largest contributors to the CAPE in the boundary layer.
In both methods, $Q$ is technically calculated as a residual of the thermodynamic equation (Eqn. 4) but a comparison was done to see if the different heating terms as a whole gave similar results. This comparison was performed between the rate of CAPE change for the residual and the heating component of CAPE change in Eqn. (6). Again, Eqn. (6) was ultimately not used to break down the boundary layer time tendency of CAPE, because we could not directly calculate the surface flux of entropy $F_s$, but we did calculate the heating rate in this term with the thermodynamic equation (Eqn. 4). The differences between the two were large, likely because the calculation $Q$ as a residual of Eqn. (7) was largely dominated by the temperature advection calculation. Direct calculation of $Q$ is needed in future work.
CHAPTER 3. DISCUSSION AND CONCLUSIONS

We conclude subcloud layer heating due to insolation was the primary driver of CAPE increase at Belle Mina, Alabama on 31 March 2016, followed by moisture advection.

The CAPE increase estimated using the “sounding replacement” method (Section 2.2.2) (1822 J kg$^{-1}$ and 141 J kg$^{-1}$ from solar radiation and moisture advection, respectively) suggests that CAPE should have increased more than the observed 1163 J kg$^{-1}$, which leads to the conclusion that some other mechanism was present that modulated CAPE values. This modulating influence could be cooling of the subcloud layer. Vertical air motion could also be cooling the subcloud layer (Fig. 16). Further observation and simulation are needed to isolate the effects of each of these processes on CAPE.

Over the course of the afternoon, CAPE increased from about 200 J kg$^{-1}$ to the observed 1163 J kg$^{-1}$. This increase was seen in both the simulations and observations, suggesting the model could be used for some further interpretation of the observations. Utilizing the model, we were able to determine that most of the increase in instability in the atmosphere occurred within the subcloud layer (consistent with Kirkpatrick et al. 2011). Within this layer, heating played the largest role in destabilizing the atmosphere and controlling the time tendency of CAPE, as evident in the analysis from the soundings (Fig 10) and model analysis (Fig 14, 15). However, it was found that radiative cooling and advection of relatively cool air toward Belle Mina exerted a modulating influence, limiting the peak value of CAPE.

Both the model data and the observations lead us to conclude that heating of the atmosphere, particularly of the boundary layer, lead to the CAPE increase preceding the Priceville, Alabama storm and tornado on 31 March 2016. The results of numerical simulation also lead us to conclude that temperature advection played a significant role in the warming of
the boundary layer. Moisture advection, while present, played a much smaller role than initially thought. These results are consistent with previous studies that found that robust convection is possible in low CAPE environments when it is concentrated in lower levels (Kirpatrick et al. 2011; McCaul and Weisman 2001).

In the future, we will also continue to examine the evolution of CAPE values in more detail by directly calculating the change in boundary layer entropy, and directly calculating the heating rate without the thermodynamic equation. To complete these calculations, we need a way to directly measure the surface flux of entropy $F_s$, or directly measure $Q$. We also look to calculate components of equation 2 using observation data instead of using just the model and look at the role of moisture in more detail. To complete these calculations, we need higher temporal resolution sounding observations from around the state to calculate advection, as well as the same measurements required for the boundary layer entropy and $Q$. Additionally, we will examine the boundary layer growth in more detail using an extended Kalman filter technique (Lange et al. 2015) to objectively identify the top of the boundary layer and track its evolution.
APPENDIX

Scale Analysis of Time Tendency of CAPE:

\[
\frac{d\text{CAPE}}{dt} = (T_{LCL} - T_{EL}) \frac{dS_b}{dt} - \int_{P_{EL}}^{P_{LCL}} \frac{R_d Q}{C_p} \left( \frac{T_v}{T} \right) - \frac{R_d T_v}{\theta} (v_r \cdot \nabla \theta)d\ln P
\]

A. B. C.

For this scale analysis term A was rewritten as:

\[
(1 - \frac{T_{EL}}{T_o}) C_p \frac{dT}{dt}
\]

Both \(T_{EL}\) and \(T_o\) are on the order of \(10^2\) K, but \(T_o\) is larger than \(T_{EL}\). The fraction of \(\frac{T_{EL}}{T_o}\) scales to \(10^{-2}\). \(C_p\) scales to \(10^3\) J kg\(^{-1}\) K\(^{-1}\), and \(\frac{dT}{dt}\) scales to \(10\) K per 6 hours. This results in the boundary layer term (A) scaling to \(10^2\) J kg\(^{-1}\) 6 hour\(^{-1}\).

With typical mesoscale values, and taking Q scale to 2 K per day, term B and C scale to \(10\) J kg\(^{-1}\) 6 hour\(^{-1}\). This means changes in the low level, or subcloud layer will likely contribute most to CAPE. This analysis was based off of the one done in Humphry and Bosart 2011.

Table 4: Calculated CAPE rates, in J kg\(^{-1}\) 5 min\(^{-1}\) from 2200 UTC on 31 March 2016 to 0200 UTC on 1 April 2016 from the ARPS simulations using equation 2. X represents when the calculated value was not available.

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<th>Advection</th>
<th>Radiative Cooling</th>
<th>Boundary Layer Entropy</th>
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