

THESIS

AN ANALYSIS OF TOTAL LIGHTNING CHARACTERISTICS IN TORNADIC STORMS:
PREPARING FOR THE CAPABILITIES OF THE GLM

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ABSTRACT

AN ANALYSIS OF TOTAL LIGHTNING CHARACTERISTICS IN TORNADIC STORMS: PREPARING FOR THE CAPABILITIES OF THE GLM

Numerous studies have found that severe weather is often preceded by a rapid increase in the total lightning flash rate. This rapid increase results from numerous intra-cloud flashes forming around the periphery of an intensifying updraft. The relationship between flash rates and updraft intensity is extremely useful to forecasters in severe weather warning decision making processes, but total lightning data has not always been widely available. The Geostationary Lightning Mapper (GLM) will be the first instrument to detect lightning from geostationary orbit, where it will provide a continuous view of lightning over the entire western hemisphere. To prepare for the capabilities of this new instrument, this thesis analyzes the relationship between total lightning trends and tornadogenesis.

Four supercellular and two non-supercellular tornadic storms are analyzed and compared to determine how total lightning characteristics differ between dynamically different tornadic storms. Supercellular tornadoes require a downdraft to form while landspout tornadoes form within an intensifying updraft acting on pre-existing vertical vorticity. Results of this analysis suggest that the supercellular tornadoes we studied show a decrease in flash rate and a decrease in lightning mapping array (LMA) source density heights prior to the tornado. This decrease may indicate the formation of a downdraft. In contrast, lightning flash rates increase during landspout formation in conjunction with an intensifying updraft. The total lightning trends appear to follow the evolution of an updraft rather than directly responding to tornadogenesis.

To further understand how storm microphysics and dynamics impact the relationship between lightning behavior and tornadogenesis, two of the tornadic supercells were analyzed over Colorado and two were analyzed over Alabama. Colorado storms typically exhibit higher flash rates and anomalous charge structures in comparison to the environmentally different Alabama storms that are typically normal polarity and produce fewer flashes. The difference in microphysical characteristics does not appear to affect the relationship between total lightning trends and tornadogenesis.

The capabilities of GLM are yet to be determined because the instrument is still in its calibration/validation stages. However, as part of the GLM cal/val team, we were in a unique position to examine the first-light GLM data and contribute to the assessment of its performance for noteworthy thunderstorm events during the Spring/Summer seasons of 2017. The final chapter of this thesis displays a preliminary analysis of GLM data. A first look into GLM performance is established by comparing GLM data with data from other lightning detecting instruments. Overall, GLM appears to detect fewer flashes than other lightning detecting networks and instruments in Colorado storms, more so for intense storms compared to weaker storms.

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CHAPTER 1: INTRODUCTION

Lightning is both a dangerous and useful phenomenon that occurs in almost all severe storms. On average, lightning causes over 60 deaths in the United States per year and produces tens of millions of dollars in damage (Curran et al. 2000). While dangerous by itself, lightning is also associated with convective storms that bring about other hazards such as hail, flooding, damaging straight line winds, and tornadoes. Due to this association, many studies have attempted to link lightning behavior to the evolution of severe storms and the timing of severe weather at the surface (e.g, Williams et al. 1999; Buechler et al. 2000; Goodman et al. 2005; Steiger et al. 2005, 2007; Darden et al. 2010; White et al. 2012; Stano et al. 2014). These studies have found that intra-cloud (IC) lightning rapidly intensifies before the onset of severe weather due to an intensification of the updraft.

The Geostationary Lightning Mapper (GLM) orbiting on the recently launched Geostationary Operational Environmental Satellite (GOES-16) was created to make lightning information more widely available to forecasters to assist in the severe weather warning decision making process (Goodman et al. 2013). Though lightning serves as a useful indicator for identifying updraft intensification, the relationship of lightning to individual types of severe events is still poorly understood. It is important to continue to study the relationship between lightning behavior and individual types of severe weather so that forecasters can develop a better understanding of how GLM data can be best utilized. To assist in this preparation, this thesis focuses on the relationship between lightning and tornadoes. By analyzing various unique case studies of dynamically different tornadic storms, for example, supercellular vs.

non-supercellular, a better understanding of how lightning relates to the formation of tornadoes is established. This thesis also includes an initial analysis of GLM data to help assess how well GLM detects lightning in comparison to other lightning detecting instruments.

1.1 TORNADOGENESIS

To fully understand how lightning characteristics in a storm relate to the observation of a tornado at the surface, it is important to understand the currently accepted theory for tornadogenesis. There are two types of tornadoes as defined by Davies-Jones et al. (2001). Type I tornadoes form within supercell storms in association with a large, typically cyclonic, rotating updraft known as a mesocyclone. Type II tornadoes form without a parent mesocyclone. This study will explore the relationship of lightning to both tornado types.

1.1.1 SUPERCELL TORNADOES

A supercell is a long-lived storm that is characterized by a single, typically cyclonically rotating updraft called a mesocyclone (American Meteorological Society Glossary).

Mesocyclones form through the tilting of environmental horizontal vorticity into the vertical by the storm's updraft (Rotunno 1981; Davies-Jones 1984). Wind shear within the environmental flow creates horizontally oriented vortex tubes along the surface, as can be visualized in Figure 1.1 (Markowski and Richardson 2009). If surface heating and atmospheric instability are sufficient enough to allow air to begin rising, these vortex tubes are tilted upward by the rising parcel (Davies-Jones 1984). This upward tilting results in two regions of vertical vorticity situated on either side of the updraft. One region is characterized by cyclonic vertical vorticity while the other is characterized by anticyclonic vertical vorticity.

The relationship of horizontal vorticity to the environmental flow determines if the cyclonic vertical vorticity will become associated with the updraft to create a mesocyclone. If

the horizontal vorticity runs perpendicular to the environmental flow, the horizontal vorticity is said to be purely crosswise. Crosswise vorticity does not support the formation of a mesocyclone because it displaces both the cyclonic and anticyclonic vortices away from the updraft (Davies-Jones 1984). If the environmental horizontal vorticity has a strong streamwise component, meaning that it is parallel to the storm relative flow, the cyclonically rotating region of vertical vorticity is shifted toward the strongest portion of the updraft. The anticyclonically rotating vortex is shifted away from the updraft into an area of descending air (Davies-Jones 1984). This shifting results in a strong, cyclonically rotating updraft that dominates the inflow for the storm and an anticyclonically spinning downdraft.

The tilting of horizontal vorticity into the vertical via an updraft leads to strong vertical vorticity at mid-levels, but this mechanism is not sufficient for producing the strong vertical vorticity at the surface that is necessary for tornadogenesis. A downdraft is required to create sufficient vertical vorticity at the surface (Davies-Jones 1982; Klemp and Rotunno 1983; Rotunno and Klemp 1985, Markowski 2002, Davies-Jones 2015). A review paper by Markowski (2002) describes the rear-flank downdraft (RFD) as a region of descending air along the rear side of a supercell. The RFD has a well-established connection to hook-echoes and has been linked to tornadogenesis. The physical mechanism in which this downdraft forms is still poorly understood, but cooling via melting and evaporation of precipitation are probable factors. Markowski (2002) shows that the downdraft could be due to negative buoyancy caused by evaporative cooling, melting hail, or precipitation loading. The downdraft could also be caused by vertical perturbation pressure gradients caused by gradients in vertical vorticity, pressure perturbation caused by varying vertical buoyancy, or even from an immobility of air near the

updraft (Markowski 2002). Past research studies touched upon in this review paper show that each of these mechanisms can play a role in the formation of an RFD.

The RFD advects vertical vorticity to the surface where it is subsequently stretched by the updraft, allowing for the vertical vorticity to intensify near the surface, making tornadogenesis possible (Markowski 2002; Davies-Jones et al. 2001; Davies-Jones 2015). This process of tornadogenesis can occur more than once in a supercell as some supercells produce numerous tornadoes in their life time. Subsequent tornadogenesis within a supercell forms in the same way described above, but typically forms more quickly than the first tornado. The region of cold air left behind by the original RFD outflow leads to a buoyancy gradient that is conducive to baroclinic horizontal vorticity generation (Alderman et al. 1999; Davies-Jones et al. 2001). This horizontal vorticity can be tilted vertically and advected to the surface where it may stretch and form subsequent tornadoes.

1.1.2 NON-SUPERCCELL TORNADOES

Tornadoes that are not associated with a mesocyclone can form in multiple ways. For the sake of this study, non-supercellular tornadoes termed landspouts will be described and analyzed. These tornadoes were termed landspouts because they form in a similar way to the waterspouts that commonly occur over the ocean within non-severe thunderstorms. While supercells require a downdraft to advect vorticity to the surface so that it can be tilted into the vertical to form vertical vorticity, landspout tornadoes do not require a downdraft (Wakimoto and Wilson 1989).

Figure 1.2 shows a conceptual drawing of how landspout tornadoes form (Markowski and Richardson 2002). In this case, vertical vorticity must already be present at the surface. Small regions of vertical vorticity, often called misocyclones, typically occur along surface boundaries (Wakimota and Wilson 1989). For example, during the summer in Colorado, a

convergence and vorticity boundary coined the “Denver Convergence and Vorticity Zone” or “DCVZ” often sets up near the Denver International airport (Szoke et al. 1984; Brady and Szoke 1989). This boundary forms when southeasterly winds interact with the topography to create a cyclonic feature that produces a region of convergence. Within this line of convergence, small misocyclones can be embedded. This mesoscale set up makes Colorado a region where landspouts are relatively common.

In order to create a tornado out of a misocyclone, vertical vorticity at the surface must be drawn upward. This occurs when the updraft of a developing storm moves over a misocyclone. The updraft stretches the vertical vorticity, causing it to be drawn upward and to increase in intensity as its radius decreases (Wakimoto and Wilson 1989). Because the landspout forms directly from an updraft stretching vertical vorticity, the tornado is completely dependent on the strength of the updraft. Typically, these tornadoes form in ordinary thunderstorms, far from a severe classification. Once a downdraft begins to develop and precipitation begins to fall, choking off the updraft, the landspout typically diminishes. Though these landspouts are often not as long lived or as strong as their supercellular counterparts, they are still able to cause costly damage to structures and endanger people’s lives. The strong winds found within a landspout make for very dangerous aviation conditions, causing these landspouts to be of great interest to forecasters.

1.2 LIGHTNING CHARGE SEPARATION

Lightning is considered to result from charge separation within the mixed phase region of a convective cloud. Typically, this charge separation manifests itself as a normal tripole charge structure characterized by mid-level negative charge that is situated between an upper and lower region of positive charge (Williams 1989). The physical mechanisms that create this charge

structure have been the focus of numerous laboratory studies (Reynolds et al. 1957; Takahashi, 1978; Jayaratne et al. 1983; Baker et al. 1987; Saunders et al. 1991; Saunders and Brooks 1992; Saunders and Peck 1998; Mason and Dash 1999, 2000). Reynolds et al. (1957) was the first to find that significant charge separation occurs when a simulated graupel particle undergoing riming collides with ice crystals in the presence of supercooled liquid water.

In a similar laboratory study, Takahashi (1978) discovered that the sign of charging on graupel was highly dependent on the ambient temperature and the cloud liquid water content. At temperatures colder than -10°C , graupel charged negatively while ice crystals charged positively. Takahashi also established that this charge reversal temperature was dependent on the supercooled liquid water content. Positively charged small ice crystals are lofted to high altitudes, creating an upper positive charge region. Large graupel particles possess significant terminal velocities which allows these particles to fall to lower altitudes within a cloud. As graupel particles fall, many particles reach a balance level in which their terminal velocities are matched by the updraft velocities, leading to a stratified region of negative charge at mid-levels of the storm formed by the levitated graupel (Lhermitte and Williams 1985). Observations of thunderstorms have revealed that this mid-level negative charge layer tends to center somewhere between -10°C and -20°C (Dye et al. 1986).

Takahashi (1978) found that the sign of graupel charging reversed to positive below the so-called charge reversal temperature of -10°C . This positive charging of graupel would explain the lower positive charge region found in a thunderstorm tripole charge structure. Jayaratne et al. (1983) found that the charge reversal occurred at -20°C . Continued laboratory experiments have shown that graupel can also charge negatively at temperatures warmer than -10°C when cloud liquid water contents are extremely small. These studies have also discovered that graupel

charges positively at extremely large liquid water contents (Saunders et al. 1991; Saunders and Peck 1998).

While laboratory studies agree that graupel can charge both positively and negatively within a cloud, the mechanism that allows for this charge reversal is still a topic of debate. Saunders et al. 2008 summarizes the various proposed mechanisms that have been disproven over the years. The non-inductive charging mechanism is the currently accepted theory for charge separation within thunderstorms. Baker and Dash (1994) suggest that a quasi-liquid layer (QLL) surrounding an ice particle is responsible for the transfer of charge via the non-inductive charging mechanism. This layer exhibits outward negative charge because OH^- ions are less mobile than H^+ ions. H^+ ions bury themselves in the ice lattice, while OH^- ions remain within the thin quasi-liquid layer.

When an ice crystal and graupel particle collide, the QLL's of each particle intersect and interact (Dash et al. 2001). Particles growing fastest via vapor deposition maintain a thicker QLL. During collision, the QLL's quickly attempt to reach an equilibrium, resulting in the transfer of mass from the thicker QLL to the thinner QLL, or a QLL of higher chemical potential to smaller chemical potential. The contact time between the particles is just long enough for the negative charge within the thicker QLL to transfer to the particle with a thinner QLL, but not long enough for the positive charge to transfer as well (Dash et al. 2001). Baker and Dash (1994) state that because of the difference in chemical potential, negative charge is typically transferred from warm to cold, from higher surface curvature to lower surface curvature, and from high vapor growth to low vapor growth.

At higher altitudes where the liquid water content is limited, graupel is typically in a state of sublimation. Because graupel does not grow quickly via vapor deposition compared to

ice crystals, ice crystals will display a thicker QLL. When the particles collide, matter from the thicker QLL is transferred to the thinner QLL found on graupel. Thus the graupel gains negative charge at upper levels (Dash et al. 2001). At altitudes where the temperature is warmer than the charge reversal temperature, graupel must charge positively. Multiple studies attribute this positive charging to graupel growing via wet growth at warmer temperatures (e.g. Takahashi 1978). This would allow the graupel particles to maintain larger QLL's and to transfer mass to the ice crystals, reversing the charging process. Saunders and Brooks (1992) found that charging due to collisions with particles undergoing wet growth is not substantial enough to generate charge strong enough to induce lightning. Saunders and Peck (1998) suggest that positive charging on graupel may result from ice filament breaking off of growing graupel. In this case, graupel grows via vapor deposition in a high liquid water content environment. The filaments that extend out from the graupel due to accretion to the graupel's surface exhibit warmer temperatures at the ends of the filaments due to latent heat released from vapor deposition. OH^- ions concentrate in the warmer outer section while more mobile H^+ move closer to the parent graupel particle. Collisions with ice crystals cause the ends of the filament to break off, leaving the graupel charged positively. Scientists continue to debate the charging mechanism in thunderstorms.

The tripole charge structure promotes intra-cloud (IC) flashes between the mid-level negative charge layer and the upper and lower positive charge layers. It also creates an environment conducive for negative polarity ground strikes (-CGs). Though -CG's make up a majority of the ground strikes found in storms across the United State, certain regions and types of storms promote the occurrence of numerous positive polarity ground strikes (+CG's; Boccippio et al. 2001; Carey et al. 2003; Williams et al. 2005; Fuchs et al. 2015). Boccippio et

al. (2001) found that climatologically, the High Plains of the United States produce more +CG's than any other region in the country. Multiple studies have also shown that +CG's are commonly found to dominate the ground strike polarity in severe storms (e.g. Branick and Doswell 1992; Carey and Rutledge 1998; Lang et al. 2002, 2004; Carey et al. 2003; Wiens et al. 2005; Tessendorf et al. 2007). The formation of +CG's has been linked to an "inverted" or "anomalous" charge structure that is characterized by a mid-level positive charge layer typically found around -15°C to -20°C with a negative charge region positioned above it (Carey and Rutledge 1998; Wiens et al. 2005; Tessendorf et al. 2007; Fuchs et al. 2015).

The physical mechanism as to how this charge structure sets up is still largely unknown. Since +CG's typically occur in the High Plains within severe storms, there must be a specific type of storm morphology or environmental characteristic that is unique to this region. Williams et al. (2005) hypothesizes that high cloud base heights and corresponding small warm cloud depths could explain why this region commonly observes anomalous storms. The authors explain that small warm cloud depths promote more efficient ice production above the freezing level because more liquid water is lofted into the mixed phase. This occurs because a shorter warm cloud depth results in less efficient warm rain processes. Higher cloud base heights also promote broader, more intense updrafts that are subject to less entrainment of dry air into the cloud (McCarthy 1974). Broader updrafts in combination with less entrainment and enhanced lofting of liquid water above the freezing level lead to enhanced supercooled liquid water contents in the mixed phase region of the cloud, which is known to promote positive charging on graupel (Saunders et al. 1991; Saunders and Peck 1998). Fuchs et al. (2015) confirms this hypothesis and shows that Colorado is unique in that it produces predominantly anomalous

polarity storms, while regions such as Alabama typically only produce anomalous polarity storms in severe weather.

A thunderstorm does not need to be anomalous to produce +CG flashes. An “end-of-storm-oscillation” or “EOSO” sometimes occurs during the decaying portion of a storm (Marshal and Lin 1992; Williams et al. 1994; Pawar and Kamra 2007; Marshal et al. 2009). An EOSO is characterized by a change in the electric field as a storm dissipates. A typical normal polarity tripole thunderstorm is dominated by negative charge overhead, with a positive charge layer situated above and below the main negative charge region. Sometimes a small layer of negative charge can be found above the tripole. During an EOSO, the main negative charge becomes less prominent and positive charge dominates the cloud for a period. The electric field within the thundercloud then switches back to being dominated by negative charge before the storm completely dissipates. The dominance of positive charge within a cloud is conducive for +CG production.

The way in which the positive charge comes to dominate the cloud is still a topic of debate in the literature. Williams et al. (1994) suggests that an inverted dipole charge structure sets up because of a change in the charging on hydrometeors through the non-inductive charging mechanism as storm dynamics change in the dissipation stage. Marshal et al. (2009) suggests that the polarity changes primarily result from the fallout of charge as hydrometeors fall through the cloud. The study then goes on to explain that the switch back to mainly negative charge overhead results from the upper negative charge layer that formed above the tripole structure falling to lower levels. This occurs as the positive charge regions fall to the ground behind the initial fallout of the main negative charge region. Further study is required to fully understand

how EOSO's occur, but overall it is well accepted that they only occur during the dissipation phase of a storm.

1.3 LIGHTNING OBSERVATIONS IN SEVERE STORMS

Total lightning flash rates (intra-cloud and cloud-to-ground flashes combined) are closely related to the characteristics of a storm's updraft (Goodman et al. 1988; Dye et al. 1989; Carey and Rutledge 1994; Williams et al. 1999; Lang and Rutledge 2002; Wiens et al. 2005; Tessendorf et al. 2007; Deierling et al. 2008; Schultz et al. 2015, 2017). Deierling and Petersen (2008) showed that larger updraft volumes with stronger updrafts lead to a larger production of graupel and ice crystals in the mixed phase region of the cloud. With more graupel and ice crystals available, there is greater opportunity for collisions and thus charge separation to occur via the non-inductive charging mechanism. There have been multiple reports of "lightning holes" in the literature, in which lightning is virtually non-existent within the strongest portions of the updraft (e.g. Krehbiel et al. 2000; Wiens et al. 2005; Steiger et al. 2007). Studies show this occurs because intense updrafts limit the time for small particles to grow into modest size precipitation particles. The majority of the collisions between particles occurs along the periphery of the main updraft, where graupel is able to descend and collide with ice crystals as they ascend to higher altitudes. This mechanism is supported by numerous case study analyses that have found increases in total lightning flash rate are more highly correlated with updraft volume than the maximum updraft speed (Deierling and Petersen 2008; Schultz et al. 2015, 2017). The size of lightning discharges is typically small near the updraft, as turbulence leads to small pockets of intense charge separation (Bruning and MacGorman 2013). Thus, the mean flash size has been found to decrease as updrafts intensify and flash rates increase (Schultz et al. 2015).

Due to the strong observed correlation between the total lightning flash rate and the updraft strength in a storm, lightning can be useful for nowcasting severe weather. Rapid increases in lightning flash rates are termed “lightning jumps” (Williams et al. 1999). Total lightning jumps typically form before the onset of severe weather because they accompany a rapid intensification of the updraft (Williams et al. 1999; Gatlin and Goodman 2010; Schultz et al. 2011, 2015, 2017). Total lightning jump algorithms attempt to automate the detection of these flash rate increases in an effort to earlier predict the occurrence of severe weather (Gatlin and Goodman 2010; Schultz et al. 2011, 2015, 2017). These algorithms have proven successful in identifying potentially severe storms.

Though total lightning jumps (both IC and CG flashes included) are useful in identifying intensifying storms that may produce severe weather, they require other atmospheric data to help determine what type of severe weather may form. Previous studies have attempted to relate the occurrence of CG flashes with tornadogenesis, but have not succeeded in finding a robust trend that could be useful in nowcasting tornadic activity (Perez et al. 1997; MacGorman et al. 1989; Keighton et al. 1991; MacGorman and Nielsen 1991, Strader and Ashley 2014). Metzger and Nuss (2013) attempted to compare and relate trends in IC and CG flash rates to different types of severe weather. Some weak trends were observed, but overall there is still a large lack of understanding of how total lightning can be used to predict specific types of severe weather. In an effort to build upon this knowledge base, this study will focus specifically on the relationship between tornadic storms and lightning characteristics.

Multiple studies have analyzed how total lightning flash rates produced by tornadic supercells change with respect to the timing of tornadogenesis (Williams et al. 1999; Buechler et al. 2000; Goodman et al. 2005; Steiger et al. 2005; Steiger et al. 2007; Darden et al. 2010; White

et al. 2012; Stano et al. 2014). Each of these studies have found that total lightning flash rates “jump” well before a tornado is reported. Many of these studies also observed a decrease in the flash rate as a tornado reached the surface. Williams et al. (1999) attributes this decrease in total lightning flash rate to the likely formation of a downdraft that helps to bring vorticity to the surface for tornadogenesis. The height at which lightning occurs within a cloud has also shown promise in helping to determine information about storm dynamics. Steiger et al. (2005, 2007) found that lightning flashes occurred at the highest altitudes when the storms analyzed were most intense. The flashes tended to descend in height during the onset of a tornado. It is hypothesized that this descent results from an updraft weakening/downdraft formation.

1.4 GOALS OF THIS STUDY

There is proven utility in monitoring total lightning flash rates during severe weather nowcasting operations. The relationship of lightning to updraft intensification is apparent, but the way in which the dynamics that produce severe weather affect lightning are less understood. Lightning jumps provide evidence of an intensifying storm, but they do not provide any indication of what will happen after the storm intensifies. The storm could produce hail, strong winds, tornadoes, or no severe weather at all. This study takes a step away from the lightning jump framework to take a more in depth look at how lightning evolves with the evolution of severe storms. Specifically, this study seeks to understand how tornadogenesis relates to changes in lightning characteristics. The following questions are addressed:

- 1) How do flash rates react to the onset of a tornado at the surface? Does the type of tornadogenesis (i.e. supercellular vs. landspout tornadogenesis) change these results?
- 2) Do other lightning characteristics, such as the height of the lightning flashes or the occurrence of CG flashes, reveal information about the evolution of tornadic storms?

- 3) How does the environmental regime affect the lightning characteristics? Specifically, how does lightning differ between tornadic storms in Alabama and Colorado?

Detailed case study analysis of a number of lightning producing tornadic storms are analyzed to seek answers to these questions. The total lightning flash rate, CG flash polarity, and location of lightning flashes with regard to altitude are analyzed in each case. To address the difference between lightning in different environmental regimes, two supercells are analyzed in the dry, high cloud base height environment in the High Plains (Colorado) and two supercells are analyzed in the warm, humid, low cloud base height environment found in Alabama. To understand how differing forms of tornadogenesis affect lightning characteristics, lightning produced by supercells is compared to lightning produced by landspouts. This analysis only studies six tornadic storms, so the results are not robust by any means. By taking a close look at individual storms, this study looks to bring an awareness of what may be controlling the changing lightning characteristics within evolving severe storms.

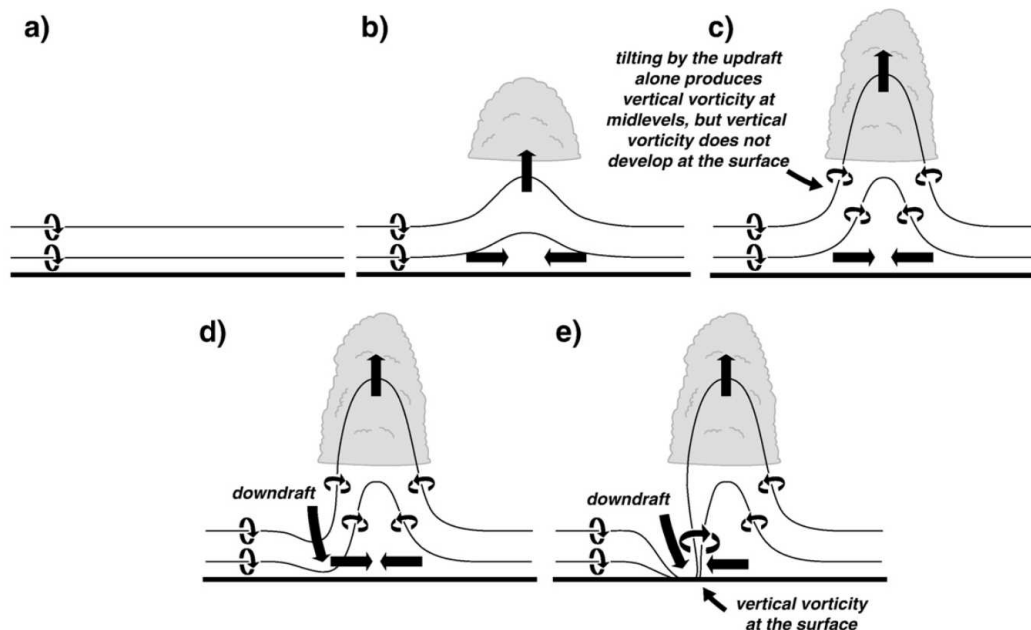


Figure 1.1 Visual depiction of supercellular tornadogenesis taken from Markowski and Richardson (2009).
a) horizontal vorticity vortex tubes set up as result of environmental shear, b) an updraft lifts these vortex tubes upwards, c) upward tilting of the vortex tubes creates vertical vorticity at upper levels, creating a rotating updraft, d) a downdraft advects vertical vorticity to the surface as horizontal vorticity continues to be turned into the vertical e) vertical vorticity advected to the surface is stretched by the updraft and enhanced, allowing for tornadogenesis to occur

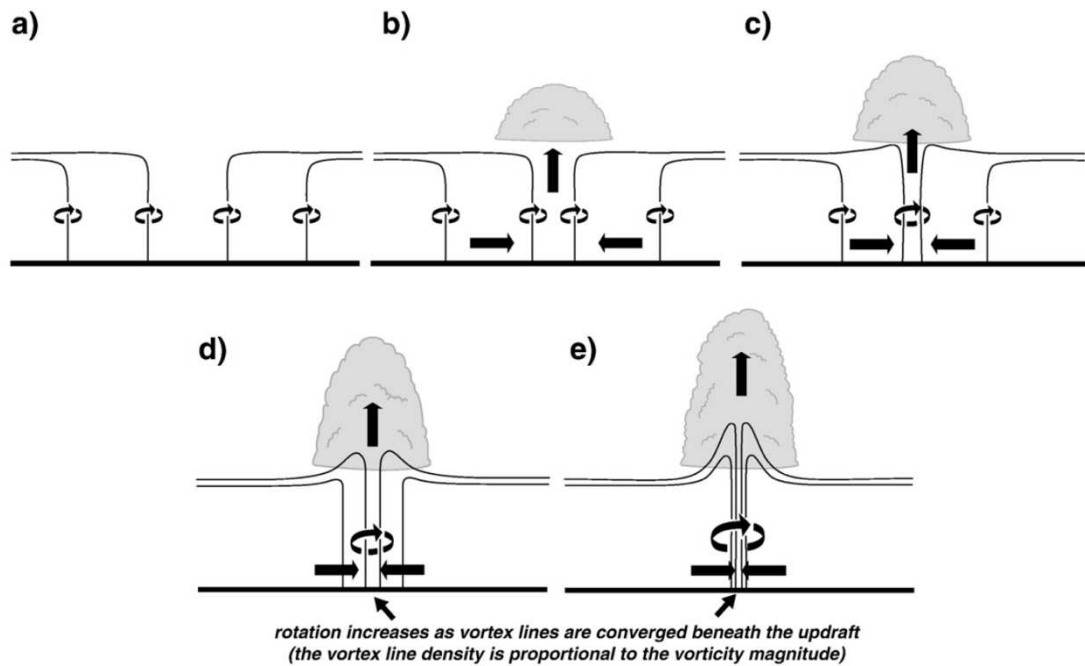


Figure 1.2 Visual depiction of landspout formation taken from Markowski and Richardson (2009). a) vertical vorticity is present at the surface, b) an updraft from a developing storm moves over a region of confined vertical vorticity at the surface, c) the updraft stretches a region of vertical vorticity upwards, d) vertical vorticity is enhanced by stretching, e) vertical vorticity converges underneath the updraft, allowing for tornadogenesis

CHAPTER 2: DATA AND METHODOLOGY

2.1 RADAR DATA

2.1.1 NEXRAD WSR-88D RADARS

The Next Generation Weather Radar (NEXRAD) program was created to develop and deploy an advanced network of WSR-88D's (Weather Surveillance Radar- 1988 Doppler) throughout the United States (Crum and Alberty 1993). These radars are run 24 hours a day and are used primarily by the National Weather Service (NWS) for forecasting and severe weather warning purposes. The WSR-88D network is comprised of S-band radars that simultaneously transmit and receive both horizontally polarized (H) and vertically polarized (V) electromagnetic radiation (Doviak et al. 2000). Radar volumes are made up of 360 ° azimuthal scans with varying elevation angles. Multiple predefined scan strategies can be utilized to create these volumes based on operational need (Crum and Alberty 1993). WSR-88D data is available for free download for any radar site in the network at <https://www.ncdc.noaa.gov/nexradinv/>.

This study utilizes WSR-88D data from the KFTG site in Denver, CO and the KHTX site in Huntsville, AL. Radar data output utilized in this study include reflectivity (Z), aliased radial velocity (V_r), and differential reflectivity (ZDR). The radial velocity field is dealiased using the Python ARM-Radar Toolkit (Pyart; Helmus and Collis 2016). Pyart has multiple automated dealiasing algorithms that can be used to correct the data. Visual inspection of the data after running the radial velocity fields through two of these automatic dealiasing algorithms show that these algorithms are able to correctly unfold the velocity fields. To remove clutter and unwanted reflectivity echoes resulting from non-meteorological objects, the data are thresholded using the standard deviation of differential phase (SDP) and the cross-correlation coefficient (ρ_{HV})

(Ryzhkov and Zrnica 1998). After dealiasing and decluttering the data, the radar data are then further quality controlled and specific differential phase (KDP) is calculated utilizing the Dual-Polarization Radar Operational Processing System (DROPS), described by Chen et al. (2017). This system calculates KDP through the Wang and Chandrasekar (2009) method. The data are then gridded to a Cartesian coordinate system.

2.1.2 CSU-CHILL RADAR

The Colorado State University - University of Chicago–Illinois State Water Survey (CSU-CHILL) National Radar Facility (Brunkow et al. 2000) is located in Greeley, Colorado. This state of the art research weather radar supports full polarimetric radar operation at both S-band (11.01 cm) and X-band (3.18 cm). S-band operations are characterized by a 1° beamwidth and X-band operations are characterized by a 0.3° beamwidth. The radar is capable of both simultaneous and alternating transmission of H and V at S-band (alternating at S-band only). S-band and X-band operations can be run simultaneously, or in a single frequency configuration.

This study utilizes S-band data for one of the Colorado cases. Because the CSU-CHILL radar is a research weather radar, it performs fully configurable surveillance scans, plan-position indicator (PPI) scans, and range height indicator (RHI) scans as commanded. RHI scans display vertical cross sections through a storm at a specified azimuth at a much higher resolution than gridded radar volume data can provide. CSU-CHILL data output includes Z, aliased Vr, KDP, and ZDR and co-polar correlation coefficient. Similar to the NEXRAD WSR-88D data, the radial velocity fields were dealiased using Pyart automatic dealiasing algorithms and quality controlled by thresholding based on SDP and ρ_{HV} .

2.1.3 ARMOR RADAR

The Advanced Radar for Meteorological and Operational Research (ARMOR) is located in Huntsville, AL. This C-band (5.33 cm) polarimetric radar transmits H and V simultaneously. The beamwidth is 1.0 °. Radar variables of interest output by ARMOR include Z, aliased V_r , and ZDR. The data were quality controlled using DROPS and thresholded based on SDP and ρ_{HV} . The radial velocity field was dealiased by first running the data through the automatic unfolding algorithms used for both the WSR-88D and CSU-CHILL data. Due to the smaller Nyquist velocity that results from using C-band, the velocity field was more severely aliased than data produced by the S-band radars. The automatic dealiasing algorithms were not able to completely unfold the radial velocity fields, so the data was subjectively analyzed and manually unfolded utilizing the ARM Radar Toolkit Viewer (ARTview; <https://github.com/nguy/artview>). The data is then gridded to a Cartesian coordinate system.

2.2 CLEAR FRAMEWORK

The Colorado State University (CSU) Lightning, Environment, Aerosols, and Radar (CLEAR) framework described in Lang and Rutledge (2011) was used to identify convective cells and track them throughout their lifetime. CLEAR takes in gridded three-dimensional radar data, determines the composite reflectivity values for each x-y grid point, and then uses reflectivity and spatial thresholds to identify convective features within the composite reflectivity field. For example, five of the six cases in this study required that each convective cell maintain a 35 dBZ composite reflectivity contour of at least 20 km² in horizontal area and a 45 dBZ contour of at least 10 km² in area. Due to the small spatial area and weak reflectivity of one of the landspout cases, these thresholds were reduced to 30 dBZ for 10 km² and 40 dBZ for 5 km².

The features identified by CLEAR are assigned storm tracks that can be referenced for further analysis. The program is also able to take in other data, such as environmental sounding data, and attribute it to each of these storm cells. CLEAR outputs the original input radar files with newly added storm cell and storm track information. If other data were included, this data is attributed as well.

CLEAR does not always perfectly attribute cells to storm tracks. Sometimes CLEAR picks up on a storm track and ends it prematurely. As a result, a subjective analysis of the storm cell and track information was performed to determine exactly which tracks made up the storms of interest. To ensure that the data analyzed for each storm of interest came exclusively from that storm, all data outside of the 35 dBZ composite reflectivity contour defined by CLEAR were masked out. The area of the 35 dBZ contour was applied to the vertical by extending the contour level upward through each vertical level.

2.3 RADAR ANALYSIS

2.3.1 *POLARIMETRIC VARIABLES*

Z , V_r , ZDR , and KDP were analyzed in this study in both horizontal and vertical cross sections for each case study storm. ZDR represents the difference between the horizontal reflectivity factor and vertical reflectivity factor (or the ratios of these two powers expressed as a log quantity). Positive ZDR values are characterized by hydrometeors that are larger in the horizontal compared to the vertical, such as oblate, falling raindrops. Spherical targets display ZDR of near zero. ZDR is biased towards larger particles since it is a reflectivity-weighted measurement. KDP is similar to ZDR in that positive values represent oblate hydrometeors. But unlike ZDR , KDP is dependent on mass content as well as oblateness. Larger values of KDP indicate a higher concentration of rain. Thus large Z values collocated with large KDP indicate

heavy rain. Small (near zero, or zero) ZDR and KDP values within a region of large Z indicate the presence of tumbling hail. When rain is mixed with hail, ZDR will be depressed towards zero but KDP will be non-zero and positive.

Analyzing vertical cross sections of polarimetric data through the updraft region of a storm can reveal valuable information about updraft strength and storm maturity. On a basic level, enhanced convergence at the surface and divergence aloft depict a strong updraft. If Z values are small along the region of convergence, this indicates the presence of a bounded-weak echo region (BWER). In this case, an intense updraft lofts small hydrometeors through the region of convergence, resulting in very little time for particles to grow to sizes detectable by radar.

Enhanced values of ZDR and KDP above the freezing level, termed ZDR and KDP columns, also indicate a strong updraft (Kumjian and Ryzhkov 2008). ZDR columns represent the presence of large water droplets produced by warm rain processes. Large drops can also be formed via the melting of hail and subsequently transported upward in the updraft. KDP columns (positive KDP) indicate the lofting of oblate drops as well.

2.3.2 HYDROMETEOR ANALYSIS

Hydrometeor identification (HID) is performed using the CSU-Radartools fuzzy logic hydrometeor identification program (https://github.com/CSU-Radarmet/CSU_RadarTools). This program takes in gridded radar data and uses Z, ZDR, KDP, and the cross-correlation ratio to determine what particular hydrometeor type is dominant at each grid point. The radar data for each case is gridded using a 1 km x 1 km x 1 km grid-spacing. The graupel volume above the freezing level of the storm of interest is derived for each radar volume. Lightning production relies on the formation of graupel and ice crystals in the mixed phase region. Analysis is limited

to regions above the freezing level so that graupel falling out of the cloud does not distort the visualization of in cloud graupel volume and its relationship to lightning characteristics over time. Freezing levels are determined through analyzing atmospheric soundings available in the University of Wyoming atmospheric sounding archive (weather.uwyo.edu/upperair/sounding.html). Graupel volume is calculated by simply summing the number of grid boxes within the storm cell of interest that contain predominantly graupel above the freezing level (derived from the HID algorithm). This value is recorded for each radar volume and then displayed as a time series so that the change in graupel volume throughout the evolution of the storm can be visualized. The same methodology is applied to calculate the hail volume.

A reflectivity – mass relationship (or Z-M) is used to determine the mass of graupel and hail within the storm at each radar volume. Graupel mass is calculated using the same Z-M relationship used in Schultz et al. (2015). This relationship shown in (1) is originally derived by Heymsfield and Miller (1988). “z” represents the linear reflectivity factor in $\text{mm}^6 \text{m}^{-3}$.

$$M_G (g \text{ m}^{-3}) = 0.0052 \times z^{0.5} \quad (1)$$

The Z-M relation is applied to each grid point within the storm cell that is identified to be predominantly graupel. The mass values calculated at each grid point are then summed and recorded to create the total graupel mass for that volume. This is done for each radar volume and displayed in a time series. The same methodology is applied for calculating the hail volume, but a different Z-M relationship (2) is used (Cheng and English 1983; Carey and Rutledge 1996; Lopez and Aubagnac 1997). Z_H is the reflectivity factor due to hail in $\text{mm}^6 \text{m}^{-3}$ and ρ_H is the density of hail, taken to be 900 kg m^{-3} .

$$M_H (g \text{ m}^{-3}) = 104.21 \times 10^{-6} \rho_H Z_H^{0.1098}$$

The CSU-Radartools blended rain algorithm is used to calculate the conditional rain rate for each storm cell over time (https://github.com/CSU-Radarmet/CSU_RadarTools). This algorithm utilizes output from HID analysis to determine where ice is present so that rain rates are not contaminated by the presence of ice. The average conditional rain rate is calculated by masking out all rain rates of 0 mm hr^{-1} and then taking the average of the resulting data field within the 35 dBZ contour defined by CLEAR.

2.3.3 DUAL-DOPPLER ANALYSIS

The ARMOR radar is located in close proximity to the KHTX WSR-88D. For a period in both the Alabama supercells, radar data from ARMOR is available to complete dual-Doppler analysis. Radial velocity only measures two components of the wind field, making it so that the vertical component of the wind field cannot be determined from one radar alone. Through using radial velocity values observed by two radars, the 3-D wind field can be determined through solving the mass-continuity equation for vertical motion with appropriate boundary conditions for the integration (Lhermitte 1968; Armijo 1969).

To accurately retrieve the vertical velocity within a storm, the fall speed of hydrometeors must be accounted for when solving the mass-continuity equation. Using a simple hydrometeor identification algorithm that determines where rain, graupel, and snow occur in a cloud, a fall speed is calculated for each grid point in the radar file being analyzed. The fall speeds are calculated based on the reflectivity utilizing relationships defined in Giangrande et al. (2013). Radar volumes used for each dual-Doppler retrieval may not occur simultaneously. Storm advection between volumes is manually determined and accounted for as well.

The Custom Editing and Display of Reduced Information in Cartesian Space (CEDRIC) software is used to synthesize the radar data and produce a 3-D wind field (Mohr et al. 1986).

This program takes in gridded radar data from two radars, as well as storm advection and hydrometeor fall speed information, and solves the mass-continuity equation. The time periods in which ARMOR data is available for dual-Doppler analysis are not long enough for useful information to be gained by calculating updraft speeds and volumes, as past studies have done (e.g., Schultz et al. 2015). In the April 27, 2011 supercell case, ARMOR was switched from completing research oriented high-resolution sector scans to low level surveillance scans to assist the NWS in their severe weather nowcasting efforts on that day. The 3-D wind field is used instead to add wind vectors to radar cross sections during the Alabama supercells so that the updraft intensity can be visualized in conjunction with other polarimetric variables.

2.4 LIGHTNING DATA

2.4.1 *LIGHTNING MAPPING ARRAY*

A lightning mapping array (LMA) is a network of sensors that use GPS time of arrival techniques to locate sources of VHF radiation emitted by lightning in space and time (Rison et al., 1999, Krehbiel et al., 2000). LMAs detect radiation at frequencies of 60-66 MHz. By knowing exactly where in space these sources originate from, LMA source data can be used to map lightning channels and infer the internal charge structure of a storm. Negative breakdown within a positive charge layer produces larger amounts of radiation compared to positive breakdown in a negative charge layer (Rison et al., 1999). Altitudes that contain higher LMA source densities are inferred to be positive charge regions due to the noisier breakdown in these regions (Wiens et al. 2005; Tessendorf et al. 2007; Lang and Rutledge 2011; Fuchs et al. 2015).

The Colorado LMA (COLMA) and North Alabama LMA (NALMA) are utilized in this analysis. Sensitivity studies in Fuchs et al. (2015) show that COLMA is a more sensitive network because it has more stations and a lower noise floor than NALMA. Both COLMA and

NALMA become less sensitive moving away from the network center (Fuchs et al. 2015), as expected due to the $1/r^2$ law. In order to affirm that accurate source counts are recorded for each LMA, storms are only analyzed when they are within 100 km of the NALMA network center and 125 km of the COLMA network center. COLMA is a larger, more sensitive network, making it so that detection is accurate at a further range.

2.4.2 NATIONAL LIGHTNING DETECTION NETWORK

The National Lightning Detection Network (NLDN) detects radiation in the VLF (3-30 kHz)/LF (30-300 kHz), throughout the entire United States at a 90-95% detection efficiency (Cummins et al. 1998; Cummins and Murphy 2009). Lower frequencies are not able to detect smaller IC flashes well, but are very good at detecting CG flashes and their corresponding return strokes. IC and CG flashes detected by NLDN are differentiated through waveform analysis. Through recommendations by Dr. Timothy Lang (personal communication), all flashes classified as CG that produced peak currents less than 15 kA were considered IC flashes. If an IC flash produced a peak current greater than 25 kA, the flash was recorded as a CG flash.

2.5 FLASH CLUSTERING AND ATTRIBUTION

An open source flash clustering algorithm described in detail by Fuchs et al. (2015) is utilized to group LMA sources into flashes. The algorithm also attributes LMA and NLDN flash information to the cells previously defined by the CLEAR framework. LMA and NLDN flashes that are within 10 km of the storm cell are attributed to each individual cell. Before sources are clustered into flashes, they are thresholded based on the number of sensors that detected each source. COLMA sources are required to be detected by 7 or more sensors to be applied to the clustering algorithm. Due to the fewer number of sensors and the decreased sensitivity, 6 or more stations are required for NALMA sources (Fuchs et al. 2015).

Numerous parameters are considered when sources are clustered together to create a flash. First, a minimum number of sources per flash is defined. For Colorado storms, 10 sources are required. For the less sensitive Alabama network, 5 sources are required. Five sources was chosen for Alabama because this value produced the most reasonable supercell flash rates when compared to typical flash rates found in the literature (e.g., Goodman et al. 2005; Gatlin and Goodman 2010; Schultz et al. 2011). A 10 source minimum was tried first, but this parameter resulted in more NLDN CG flashes being detected than total lightning LMA flashes during certain time periods, which is not reasonable. A 2 source minimum was also tried, but this solution created unreasonably high flash rates. Wiens et al. (2005) demonstrates that the choice in minimum sources does not change the overall *trend* of the flash rates produced. Results from testing various minimum source parameters for NALMA confirmed this idea.

To create a flash, the minimum number of sources must be defined within specific spatial and temporal thresholds. For both NALMA and COLMA the spatial threshold requires each source in a cluster be within 3 km of one another. Each source cannot be further than 0.15 s apart in time of occurrence as well. Once the minimum number of sources is reached such that these spatial and temporal criteria are met, the program looks for more sources to attribute to that flash. If no sources are available within the proper distance and time threshold, a new search begins for a new flash.

2.6 LIGHTNING ANALYSIS METHODOLOGY

LMA flashes are used to create total lightning flash rates, with the assumption that the LMA sees all CG and IC flashes with 100% detection efficiency. The evolution of total lightning flash rates is determined by combining the flashes attributed to every radar volume storm cell within the storm track of interest. These combined flashes are binned into 1-minute

intervals. The number of flashes within each bin is recorded and then displayed in a time series such that the total lightning flash rate at each minute of the analysis period is displayed.

To monitor the evolution of the charge structure in each case study, a source density analysis is completed for each case. To do this, the LMA sources attributed to the relevant storm cell in each radar volume are binned into 0.5 km vertical intervals. The number of sources within each vertical level is counted and divided by the total number of sources in the storm cell. The percent of sources that occur at each height interval within the storm cell is revealed. This analysis is repeated for each radar volume. The results for each radar volume are combined to create a time series of these source densities. Each row of a source density plot represents a height interval and each column represents a storm cell (or radar volume) within the storm track of the case.

Each of the supercellular cases also contain a CG flash polarity analysis. The landspout cases produced no CG flashes. Corrected NLDN CG flashes are attributed to each radar volume through the flash clustering algorithm described above. Much like the total lightning analysis, the flashes attributed to each storm cell/radar volume throughout the entire storm track of interest are grouped together and binned in 1-minute intervals. The bins are broken up into +CG and –CG flash bins. The number of +CG and –CG flashes that occur within each bin are counted and recorded. A time series of 60 s +CG flash rates and –CG flash rates throughout the duration of the storm is displayed.

2.7 IR MINIMUM BRIGHTNESS TEMPERATURES

Super Rapid Scan Operation for GOES-R (SRSOR) produced 1-minute scans of satellite imagery taken from GOES-14. This scan mode was a part of the GOES-R Proving Ground that was created to help prepare forecasters and atmospheric researchers for the capabilities that

GOES-R would provide through the creation of the Advanced Baseline Imager (ABI) (Schmit et al. 2005). This study utilizes $10.7\mu m$ infrared (IR) data from SRSOR operations on June 4, 2015 to determine how the minimum brightness temperature over the overshooting top of the storm of interest on this day changed with time.

IR data is corrected for parallax using code written by Dr. Steven Miller (personal communication) that follows the methodology outlined in Vincente et al. (2002). IR data is interpolated to a 200 by 200 grid with 0.02° grid spacing in x (longitude) and y (latitude). A box encompassing the June 4 overshooting top is then defined from $105.5^\circ W$ to $104.2^\circ W$ and $39.7^\circ N$ to $40.5^\circ N$ such that any data outside of this region is masked out. This is done to ensure that no other overshooting tops from nearby storms adversely impact this analysis. The minimum brightness temperature within this box was recorded for each 60 s file. These minutely minimum brightness temperatures were then combined to create a time series of the minimum cloud top brightness temperature over the overshooting top. It should be noted that convective cloud tops are assumed to be opaque, such that they emit as a blackbody.

CHAPTER 3: RESULTS

3.1 COLORADO SUPERCELLS

3.1.1 *BERTHOUD SUPERCELL*

On June 4, 2015, an EF3 tornado touched down near Berthoud, CO in association with a supercell that moved westward along the Larimer and Boulder county lines. The tornado was on the ground from 00:30 to 01:08 UTC and produced maximum wind speeds near 135-140 mph (Storm Data- available at <https://www.ncdc.noaa.gov/stormevents/>). This storm was especially unique to this region because the supercell and the corresponding tornado moved from east to west, unlike the typical west-to-east motion of storms in this region. The intensity of this tornado was also unique as only 10% of the tornadoes produced in either Colorado or Wyoming (from a climatological perspective) are reported as EF3 or greater (Schumacher et al. 2010).

The environmental conditions for this day were conducive for tornadogenesis. Both the 12 UTC and 00 UTC Denver soundings revealed values of Convective Available Potential Energy (CAPE) near 1500 J kg^{-1} . The 00Z sounding also showed strongly sheared winds within the lowest 3 km of the atmosphere. A video analysis completed by Dan Bikos (Cooperative Institute for Research in the Atmosphere, Colorado State University; available at <http://rammb.cira.colostate.edu/training/visit/blog/index.php/2015/07/17/4-june-2014-goes-1-minute-visible-imagery-and-time-lapse-video>) discusses the patterns revealed by one-minute visible imagery available from Super Rapid Scan operations for GOES-R (SRSOR) employed on GOES-14 during the time period of this case. The one-minute visible imagery reveals that the severe storms on this day initiated from a cumulus field that moved from eastern Colorado westward and developed into towering cumulus clouds as it approached the Front Range.

Outflow boundaries and gravity waves produced by mature storms that had developed earlier in the day aided in the development of mature convection from the towering cumulus clouds. The storm of interest, nicknamed the Berthoud supercell, was spawned by this earlier convection.

The Berthoud supercell began as a cluster of multicellular storms over Larimer County. At approximately 00 UTC, the cluster split into two separate storm cells with the southward moving cell becoming the Berthoud supercell. The total lightning flash rate sharply increases from about 20 flashes min^{-1} to a local maximum of 220 flashes min^{-1} within the first 15 minutes of the analysis period (Figure 3.1a). As past studies have found, this rapid increase in flash rate is indicative of updraft intensification (e.g. Williams et al. 1999; Gatlin and Goodman 2010; Schultz et al. 2011, 2015, 2017). Figure 3.1b displays CG flash rates distinguished by flash polarity. While the total lightning flash rates increase within the first 15 minutes of the analysis period, the CG flash rate remains very small. This suggests that the total lightning flash rate is dominated by IC flashes as the updraft intensifies. When an updraft intensifies, increased turbulence leads to enhanced collisions between graupel and ice crystals, creating small pockets of charge separation along the periphery of the updraft (Bruning and MacGorman 2013). These small regions of enhanced charge separation lead to the formation of numerous small IC flashes, resulting in an increased IC flash rate, but not an increased CG flash rate. As a storm initially intensifies these small flashes may appear to rise in altitude as the storm develops and cloud tops reach higher altitudes. This can be seen in the LMA source density plot in Figure 3.1c. A slight rise in the height of the maximum source density occurs as the total lightning flash rate increases. The beginning of the Berthoud storm is characterized by source densities maximizing around 10 km, which is typical of a normal polarity storm.

Visible imagery from SRSOR data supports the idea that the updraft is intensifying throughout the first 15 minutes of the analysis period. A cauliflower-like texture accompanied by multiple shadows over a small, discrete portion of the cloud top indicates the presence of an overshooting top during the beginning of the analysis period (Figure 3.2a). Overshooting tops occur when an updraft is strong enough to penetrate the equilibrium level (American Meteorological Society Glossary). To gauge the intensity or strength of this feature, the minimum brightness temperature above the overshooting top is plotted with respect to time in Figure 3.3. During the inferred updraft intensification, the cloud top minimum brightness temperature decreases quickly. This decrease in temperature reveals that the cloud directly above the updraft is growing in altitude as the updraft penetrates further past the equilibrium level, reaching colder temperatures. By monitoring the minimum brightness temperature as opposed to the entire convective region of the cloud top, fluctuations in the updraft intensity can be inferred. The convective region of a storm will produce rising and falling cloud tops as the storm intensifies and weakens respectively, but monitoring small scale changes in intensity is more difficult because the broader convective region is subject to being stopped at the equilibrium level. By monitoring the portion of the storm directly associated to the updraft that is able to break through the equilibrium level, it is more likely that small scale changes in updraft intensity will be revealed.

Cross sections of dual-polarized radar variables show that the Berthoud storm was well organized with a defined bounded weak echo region (BWER) in the vicinity of the updraft by 00:21 UTC, approximately 9 minutes prior to the initial tornado touch-down (Figure 3.4a). The updraft is visualized by a region of convergence in the V_r field (Fig. 3.4b). The radar is located to the southeast of the Berthoud supercell. As a result, regions where blue inbound colors are to

the left of the red outbound colors show a region of convergence. An atmospheric sounding taken in Denver, CO at 00Z shows that the freezing level was slightly lower than 3 km at this time. Regions of enhanced ZDR values are found above the freezing level in the vicinity of the updraft (Figure 3.4c). These features are termed “ZDR columns” (Kumjian and Ryzhkov 2008). ZDR columns depict the lofting of oblate drops above the freezing level associated with a convective updraft. These drops promote enhanced graupel production because the drops freeze and accrete supercooled liquid water to create graupel. The KDP field in Figure 3.4d does not show any noteworthy features at this point in time.

The ZDR column in Figure 3.4c coincides with a period in which the graupel volume and graupel mass above the freezing level are increasing while the rain rate fluctuates, as depicted in Figure 3.5. While 00:15 UTC represents a local maximum in the flash rate, the graupel mass and volume both continue to grow until 00:25 UTC where a local maximum is reached. Consequently, shortly after 00:25 UTC another local maximum in the flash rate is reached. Cross sections of ZDR continue to display oblate drops being lofted above the freezing level via the updraft through 00:26 UTC. The rain rate fluctuates at this time likely as a result of the storm continually gaining intensity (Figure 3.5a). Just prior to 00:25 UTC, the minimum brightness temperature above the overshooting top warms rapidly. Investigation of satellite visible imagery reveals that the cauliflower-like texture also diminishes at this time, suggesting that the overshooting top collapsed. An example of this change in visible imagery is displayed in Figure 3.2b. The flash rate and graupel volume also show decreases shortly after 00:25 UTC, with both parameters reaching local minimums by 00:45 UTC. These changes in storm characteristics suggest that the updraft has weakened at this time.

Overshooting tops must be maintained by a strong updraft, therefore the decrease in the altitude of the overshooting top is the first sign of updraft weakening. Updrafts tend to weaken during a storm's life time as a result of precipitation loading as ice hydrometeors become larger and heavier. These ice hydrometeors typically fall out of the cloud once their terminal velocity becomes larger than the updraft speed. This precipitation loading may result in the onset of a downdraft. The decrease in graupel volume and graupel mass between 00:25 and 00:45 suggests that graupel is falling below the freezing level. Radar cross sections within this time frame also show enhanced values of both KDP and ZDR above the freezing level that are not collocated with the region of convergence that indicates updraft location. Since these values do not occur within the updraft, they likely indicate oblate droplets shedding off of melting hail while it falls through the cloud as the updraft weakens. During this same time period, Figure 3.5a shows that there is a constant increase in the mean conditional rain rate within the Berthoud supercell. Falling rain leads to evaporative cooling, which encourages downdraft formation as the cooler air makes the parcel negatively buoyant.

The data suggest that a downdraft formed at approximately 00:25 UTC. This idea is further supported by the tornado touchdown that occurs at 00:30 UTC. As discussed in Chapter 1, tornado formation within a supercell requires a downdraft to advect vertical vorticity to low levels in the atmosphere where it can subsequently be ingested into the updraft and stretched to form enhanced vertical vorticity near the surface. The combination of an increased rain rate, decrease in graupel volume, and the collapse of an overshooting top all support the idea that a downdraft has formed, allowing tornadogenesis to take place.

Shortly after the radar-inferred downdraft begins, the lightning source densities fall in altitude to about 6 km and become denser at a localized altitude rather than being spread out

throughout the column. Some severe storms become anomalous as they intensify, but in this case, the altitude of the highest source density falls as the flash rates and all other updraft parameters are within a weakening stage. As a result, the lowering of this charge region may be related to falling of positively charged hydrometeors through the cloud. The non-inductive charging mechanism causes graupel particles to charge negatively while ice crystals charge positively, leading to an upper positive charge region and mid-level negative charge region in the case of a normal polarity storm. The source densities within a storm appear greatest in regions of positive polarity charge. This occurs because negative breakdown within a region of positive charge produces more VHF radiation compared to positive charge breakdown in a region of negative charge, and thus producing more sources for the LMA to record (Rison et al. 1999). As the graupel mass above the freezing level begins to fall and the rain rate increases, it can be assumed that the graupel is falling out of the cloud while the ice crystals fall to lower altitudes due to the now weaker updraft. This scenario would result in a fallout of the negative charge region and a lowering of the positive charge region. It is interesting to note that this lowering of charge does not lead to enhanced CG flash production. Rather, CG flash production almost completely stops during the touchdown of the Berthoud tornado, as shown in Figure 3.1b. Only IC flashes are produced during this time period.

The Berthoud tornado remains on the ground through 01:08 UTC (~38 min duration). Before the tornado lifts, there is a sharp increase in the flash rate near 00:55 UTC. The flash rate continually increases until it reaches a maximum value of nearly 290 flashes min^{-1} . This extremely rapid increase in flash rate once again suggests that an updraft is intensifying. Cross sections of the polarimetric radar variables show that throughout the tornado duration, the updraft region progressively weakens with KDP and ZDR columns disappearing (Figure 3.6).

At 00:48 UTC, the Vr field shows enhanced convergence at low levels and divergence aloft compared to the previous volumes, suggesting that the updraft is organizing once again. The Vr fields increase in intensity through 00:54 UTC, just a few minutes before the maximum in flash rate is established. The increase in convergence and velocity fields suggests that the updraft has intensified once again as the flash rates increase rapidly.

Shortly after the peak in flash rate, the flash rate drops once again. Cross sections of polarimetric radar data (not shown) display weakening reflectivity fields along with enhanced KDP and ZDR at low levels, suggesting the fallout of hydrometeors. The storm also loses its supercellular structure and begins to form into three separate cells that are in close proximity to one another. Since the initial decent in LMA source density that occurred during the initial downdraft formation, the VHF source densities have predominantly been largest at lower altitudes in the cloud, near 7km. However, beginning at 01:20 UTC, the source densities transition to predominantly upper levels. During this rise in altitude, the source densities spread out throughout the storm depth. The flash rates increase during this time period. Radar cross sections show one of the storm cells producing KDP and ZDR columns. Enhanced KDP values along the updraft region above the freezing level indicate that oblate drops are being lofted above the freezing level, just as in ZDR columns (Kumjian and Ryzhkov 2008). The rise in flash rate and appearance of KDP and ZDR columns suggest that the updraft is intensifying within one of the storm cells. The source densities increase in height and spread out once again as graupel is produced by liquid water droplets being lofted above the freezing level where they are able to freeze and form graupel through accreting supercooled liquid water.

By 01:36 UTC, nearly half an hour after the Berthoud tornado lifted, the reflectivity field has decreased in intensity substantially and enhanced KDP and ZDR are seen at the surface once

again. The flash rate slowly decreases throughout this period and the source densities gradually fall in altitude. The storm system continually weakens as it rains out through the end of the analysis period. Up to this point in the storm's lifetime, very few CG's were produced. The CG's that were produced were evenly mixed between positive and negative polarity. At 02:05 UTC, the number of CG's rises to 5 flashes min^{-1} and remains larger throughout the rest of the analysis period. The majority of the CG's produced at this time were positive polarity. The quick rise in +CG production during the end phase of the storm's life is consistent with the production of an EOSO (Marshall and Lin 1992; Williams et al. 1994; Pawar and Kamra 2007; Marshall et al. 2009).

Overall the lightning characteristics depicted in the Berthoud supercell follow the evolution of the updraft's intensity. Sharp increases in flash rate indicate updraft intensification while sharp decreases in flash rate are typically accompanied by downdraft formation or the weakening of the storm as a whole. Lightning source densities also tend to follow the evolution of the storm updraft with source densities congregating at higher altitudes as the storm intensifies and then source densities falling to lower altitudes as charge carrying hydrometeors fall out. The source densities tend to spread out through the atmospheric column and weaken as they transition from lower to higher altitudes and vice versa. The source densities then intensify and become confined to discrete regions as they reach their new altitude. Though there is not an obvious change in lightning characteristics before the onset of the Berthoud tornado, the lightning data does help to reveal the evolution of the supercell dynamics.

3.1.2 DENVER SUPERCELL

On May 21, 2014, three short-lived EF0 supercellular tornadoes spawned east of the Denver metro area. The synoptic environment on this day was characterized by veering winds,

weak upslope flow, and relatively high dew points. The combination of these environmental factors led to CAPE values between 1000-2000 J kg⁻¹. A DCVZ was present, creating a region of low level wind shear and a convergence zone for convection to spark from. The Denver supercell initiated to the southwest of Denver and moved northeastward with time. The supercell traversed directly across the KFTG WSR-88D radar. As a result of the coverage gap as the storm moved directly overhead the radar, no graupel and hail mass or volume calculations could be completed for this case. The CSU-CHILL radar, located well to the north of the storm, captured numerous RHI's through the core of the supercell that are used to supplement the lack of the KFTG WSR-88D data.

Figure 3.7 displays the lightning characteristics for this case. For the first twenty minutes of the analysis period, the lightning flash rate generally increases, reaching a local maximum of about 150 flashes min⁻¹ at 20:00 UTC. Analysis of radar composite reflectivity and low level radial velocity reveal that the storm began as a cluster of non-supercellular convection. Throughout the initial twenty-minute period, the cells organized into a well-defined supercell. The mesocyclone is first visualized as a radial velocity couplet as the local peak in lightning flash rate materializes. This evolution is similar to that of the Berthoud supercell.

The initial source densities for this case began confined to low levels in the storm, at a time when the storm system was multicellular. The source densities then gradually build to an altitude of about 10 km just after the peak in flash rate, in accordance with the formation of the mesocyclone, indicating that the updraft has intensified. At this point the supercell has not passed directly over the KFTG WSR-88D, so cross sections of polarimetric radar variables are still useful. During the initial build up in flash rate, cross sections through the updraft region displayed in Figure 3.8 reveal the formation of a bounded-weak echo region (BWER). Slight

KDP and ZDR enhancements above the freezing level of 3 km indicate that a strong updraft is lofting oblate drops into the mixed phase region where graupel and ice crystals can grow and collide to provide the charge separation necessary for the increasing flash rate. Evidently graupel and small hail form upon freezing of the lofted drops. Shortly after the flash rate peaks, it quickly drops to less than 20 flashes min^{-1} . During this substantial drop in flash rate, the first, and longest-lived tornado occurs.

By 20:09 UTC, the BWER appears to collapse and the enhanced values of KDP and ZDR fall to lower altitudes. The region of convergence that indicates the location of the updraft severely weakens during this time period as well. This weakening updraft is likely due to the formation of a downdraft that is necessary to advect vertical vorticity to the surface where it can be stretched vertically to create the initial tornado. After the tornado lifts, the flash rates increase very quickly. RHI's recorded by the CSU-CHILL radar depict a recovery of the BWER, strong convergence within the updraft, and prominent KDP and ZDR columns throughout the entire period in which the flash rates are increasing. Figure 3.9 shows an example of these characteristics for time period 20:25 UTC. The flash rates reach a local maximum a few minutes after this RHI is recorded.

Two more short lived tornadoes touched down during the span of this storm. The lightning characteristics act very similarly for both of these tornadoes. There is another rapid decrease in flash rate between 20:27 UTC and 20:32 UTC. During this decrease, the second tornado is reported. RHI's produced by the CSU CHILL radar also depict ZDR and KDP columns falling in altitude. This progression can be visualized by comparing Figure 3.9 and Figure 3.10. The decrease in flash rate, decrease in ZDR and KDP column height, and the onset of the second tornado suggest another downdraft has formed. Quickly after the second tornado

lifts, the flash rates increase rapidly once again reaching another local maximum. After this maximum is reached, the flash rates quickly decrease and the third and final tornado of this storm is reported to touch ground. CSU CHILL RHI's show a lowering of the dBZ echo core during this time period, suggesting the fall out of hydrometeors as another downdraft forms.

After the third and final tornado lifts, the flash rates slowly increase until about 21:20 UTC. The flash rates decrease rapidly once again, but this time no tornado is reported. The LMA source density plot found in Figure 3.7c shows a descent in altitude that corresponds to the timing of decreasing total lightning flash rates. While the source densities fall and the total lightning flash rate decreases, the CG flash rate increases substantially. The CG flash rate reaches a maximum value of 12 flashes min^{-1} , with the majority of these flashes being positive polarity. The decrease in total lightning flash rate in accordance with descending source densities and an increase in positive polarity CG's likely result from an EOSO, similar to the decaying period of the Berthoud supercell. The total lightning flash rates do not continually decrease during this EOSO because the Denver supercell begins to converge with other convection during the end of the analysis period.

3.2 ALABAMA SUPERCELLS

3.2.1 LIMESTONE SUPERCELL

On March 2-3, 2012 a large tornado outbreak spanned a great portion of the central plains and the southeast. Over 41 tornado related deaths were associated with this outbreak. Despite the large number of deaths, this outbreak was well forecasted. The Storm Prediction Center produced a Convective Outlook that highlighted a high chance of severe storm development over Kentucky, northern Tennessee, southern Ohio, and southern Indiana. A moderate risk of severe storm development encircled this region and extended southward into Alabama, Louisiana, and

northwestern Georgia. The synoptic setup was characterized by a strong upper level trough and associated vorticity center located over the Mississippi, Tennessee and Ohio Valley regions. Strong vertical wind shear accompanied this trough. A surface low moved into this region with an associated warm front near the Ohio River and cold front aligned along the Mississippi River. These fronts brought forth shearing winds, vertical motion, enhanced vorticity, and warm air advection. All of these ingredients are important for supercell development and subsequent tornadogenesis.

Since the focus of this thesis is on how lightning relates to tornadogenesis in varying types of tornadic storms, a tornadic supercell that traversed across the North Alabama LMA will be the focus of this section. The supercell of interest, nicknamed the Limestone supercell, produced two weak, short lived tornadoes over Limestone County in northern Alabama. This long-lived supercell initially organized in Louisiana and then moved eastward towards Alabama. Once in Alabama, the supercell moved toward Alabama and Tennessee border, eventually paralleling it. The supercell eventually passed through Alabama reaching the Georgia and Tennessee border. This study will focus on a portion of the supercell's lifetime within Alabama because the study is limited to the time period in which the supercell was within 100 km of the North Alabama LMA network center.

Figure 3.11 displays the lightning characteristics of this case. The first forty-five minutes of this analysis depict relatively small flash rates that maximize near $25 \text{ flashes min}^{-1}$. The source densities are centered at 10 km altitude, characteristic of a normal polarity storm. The inferred polarity structure is confirmed by the fact that the storm produces predominantly negative CG's during this 45-minute period. At 21:05 UTC the flash rates begin to steadily increase until they maximize at 21:26 UTC. A tornado touch-down is reported at the surface at

the exact time in which the flash rate maximizes. This is different from the Colorado supercells in which the flash rate begins to decline as a result of a downdraft forming to promote tornadogenesis. In this case, the flash rate declines shortly after the onset of a tornado at the surface.

Figure 3.12 displays the microphysical parameters for this case. The hail volume oscillates between larger and smaller values for the first 45 minutes of the analysis period and then reaches a local maximum around 21:10 UTC. The graupel volume maximizes about 10 minutes after this peak in hail volume. By 21:21 UTC both the graupel and hail volume have peaked and begun a rapid decrease. The mean conditional rain rate begins increasing at 21:00 UTC and exhibited a continuous net increase in value until 21:40 UTC. The combination of hail and graupel falling from above the freezing level and the rain rate increasing suggests the onset of a downdraft, perhaps triggered by evaporative cooling of rain. This is further supported by radar cross sections through the updraft region.

At 21:13 UTC and 21:16 UTC, enhanced values of KDP and ZDR are found above the freezing level near the region of convergence that defines the updraft. The freezing level was near 4 km at this time according to a sounding taken at 18 UTC near Huntsville. Figure 3.13 displays polarimetric radar variable cross sections for 21:16 UTC. U/W wind vectors indicate upward motion along the updraft region as well as the presence of a BWER. In the next radar volume depicted in Figure 3.14, the wind vectors show weaker upward motion and KDP is no longer enhanced above 4 km. Within the subsequent radar volumes, the KDP and ZDR columns return, but a drop in radar reflectivity echo height is clearly visible, indicating the fallout of precipitation that is typically accompanied by a downdraft. The first tornado touches down during the time period in which the inferred downdraft initiates.

It is interesting that the flash rate does not decrease until the tornado is reported to be on the ground. The flash rates rise again after the tornado lifts, reaching a local maximum during the second tornado. While the flash rate increases, the hail volume also increases, but the graupel volume is still decreasing. The rain rate is increasing at this point in time. These mixed signals suggest that the mature supercell is undergoing a number of processes. The rise in flash rate along with the increase in hail volume suggest an intensification of the updraft. Analysis of the source density plot also points to the same conclusion. The source densities were gradually declining in altitude and becoming confined to one altitude during the onset of the first tornado. During the increase in flash rate and the onset of the second tornado, the sources broaden in vertical distribution and also reach greater heights than in the previous radar volumes. This suggests that an updraft is lofting graupel and supercooled liquid water to higher altitudes, allowing lightning to occur at higher altitudes. The onset of a tornado and the continual decrease in graupel volume and increase in conditional mean rain rate suggests that a downdraft is occurring simultaneously.

Radar cross sections between 21:34 and 21:38 UTC, the period in between the two tornado events, show an increase in height and intensity of the BWER and a ZDR column. This suggests the updraft was again intensifying, which corresponds well with the increase in flash rate and the spreading out of the LMA sources with height. As a convective updraft intensifies, turbulence around the periphery of the updraft intensifies allowing for small pockets of charge separation to form throughout the entire column, leading to the formation of numerous small flashes around the updraft (Bruning and MacGorman 2013). Source densities spread out throughout the column as a result of the enhanced number of small flashes occurring throughout the column, leading to a spreading out of sources in the vertical during updraft intensification.

Reflectivity at low levels shows a rain wrapped hook echo associated with the second tornado at 21:43 UTC. This wrapping of precipitation suggests that the rear flank downdraft is transporting precipitation towards the inflow region of the supercell.

This case demonstrates that mature supercells support multiple processes at once. For example, an updraft may intensify as a downdraft still materializes at the surface. Due to the nature of how lightning is formed, lightning changes in accordance with the updraft. In other words, the flash rate typically increases with an intensifying updraft and decreases with a weakening updraft, regardless of whether or not a downdraft has formed. This is true because the majority of lightning in supercell forms around the periphery of an updraft where falling graupel is able to collide with ascending ice crystals to produce charge separation that causes lightning to occur. Downdrafts tend to weaken the updraft, which results in a decrease in lightning production as less turbulence and fewer collisions between ice and graupel particles occur in a weaker vertical motion field. Though decreases in flash rate may occur with downdraft formation, they may not always occur if the updraft and downdraft occur in conjunction with one another as often occurs in an intense, mature supercell.

Shortly after the second tornado lifts, almost all storm characteristics analyzed point towards another updraft intensification. The flash rate increases in conjunction with increasing hail and graupel volume and a decreasing rain rate. Cross sections of radar data show an intensifying ZDR column between 21:47 and 22:00 UTC. There is a persistent KDP column visible during this time period as well. A unique feature is found in the source density plot (3.11c). Instead of showing an increase in altitude and a spreading out of source densities with height as would be expected with an intensifying updraft, the source densities descend to 7 km. It appears that the Limestone supercell switches polarity during this updraft intensification, with

mid-level positive charge dominating. During this descent in source densities, only two positive CG's are produced. It is unlikely that this decrease in altitude of the source densities and the production of +CG's is created by an EOSO. EOSO's are characterized by descending hydrometeors as a storm weakens. In contrast, this storm is intensifying, most likely leading to a switch in the charging on the graupel due to enhanced supercooled liquid water content being lofted above the freezing level by a more intense updraft (Fuchs 2017).

By 22:05 UTC, the Limestone supercell produces a maximum flash rate of 98 flashes min^{-1} . The graupel and hail volumes reach a localized maximum and the mean rain rate has reached a local minimum. The flash rates then begin to slowly decrease as the graupel and hail volume both drop quickly and the conditional mean rain rate increases. The combination of these observations support downdraft formation as hydrometeors fall out of the cloud. Interestingly, the source densities rise in altitude during this inferred downdraft, back to altitudes typical of normal polarity storms. The supercell appears to transition from anomalous to normal polarity during this time period. Perhaps the positive charge carrier falls out of the storm during the fallout of graupel, hail, and rain. A weaker updraft would then promote negative charging on graupel as the storm weakens and less super cooled liquid water is available at upper levels.

From about 22:25 UTC onward, the flash rates hover between 10 and 50 flashes min^{-1} , the source densities remain centered at about 10-11 km, and the supercell produces predominantly -CG's. The graupel and hail volume are increasing as well as the mean conditional rain rate. By 23:15 UTC, the supercell moves outside of the 100 km radius of the LMA network center. As a result, the analysis of the Limestone supercell ends here.

3.2.2 CULLMAN SUPERCELL

April 27, 2011 marks the most significant tornado outbreak ever recorded. An astounding 199 tornadoes were produced within 24 hours and 316 people died as a result of these tornadoes. Multiple EF-4 and EF-5 tornadoes were associated with this outbreak. Knupp et al. (2014) provides a detailed mesoscale and synoptic summary for this outbreak. The afternoon synoptic set up was close to optimal for tornadogenesis. Upper level divergence as a result of a jet streak and warm air advection over the region provided net divergence of air within the atmospheric column and the formation of a surface cyclone. A cold front was present in association with this surface cyclone. Southerly winds advected warm, moist air into the region, promoting low LCL's and CAPE values between 2500-3000 J/kg. Low level storm relative helicity values were extremely high and the significant tornado parameter suggested that a tornado outbreak was extremely likely.

April 27, 2011 was very unique in that there were three tornadic episodes with differing characteristics during the course of the 24 hours. These episodes involved a morning time mesoscale convective system that eventually transformed into a quasi-linear convective system (QLCS), a small QLCS over northern AL that occurred during the mid-day, and then isolated supercells that characterized the afternoon/evening hours (Knupp et al. 2014). This section will focus on a long-lived discrete supercell that occurred in the afternoon/evening hours. This supercell will be referred to as the Cullman supercell because it produced significant damage over Cullman, Alabama.

The Cullman supercell persisted for 7 hours, producing 8 tornadoes. Three of these tornadoes reached EF4 strength (Knupp et al. 2014). The supercell originally formed over central AL and moved northeastward with time. Due to its long-lived and long-track nature,

only a portion of this storm is analyzed. As with the Limestone supercell, only time periods in which the Cullman supercell is within the 100 km radius of the North Alabama LMA network center were considered. This storm took place before all the NWS surveillance radars had undergone the polarimetric upgrade. The ARMOR research radar produced high resolution sector scans of the Cullman supercell for a short period of time, providing limited polarimetric data for this case. ARMOR switched from sector scans to low level surveillance scans during the Cullman supercell in order to assist the NWS in their nowcasting efforts. Despite the short time period of polarimetric data available, it is useful to analyze the lightning characteristics within this very significant supercell.

The Cullman supercell crossed the North Alabama LMA 100 km radius at 19:40 UTC. An EF-4 tornado was already on the ground in association with this storm at this point in time. Polarimetric full volume scans are available from ARMOR for 19:40 to 20:37 UTC. During the first EF-4 tornado, the lightning data depicted in Figure 3.15 show a general decrease in the flash rate, though the flash rate does vary greatly on a minute by minute basis. The LMA source densities display a gradual descent during this time period as well. The storm is clearly normal polarity, with source densities remaining above or near 10 km during the entire storm. Predominantly -CG's are produced, with CG flash rates being much higher than the previous supercellular storms analyzed in this research. These flash rates exceeded $5 \text{ flashes min}^{-1}$ through the majority of the analysis period.

During the time period of the first tornado, the hydrometeor characteristics shown in Figure 3.16 suggest a strengthening updraft. The mean conditional rain rate decreases and then levels out as the graupel and hail volume both increase and then level out from 20:20 to 20:37 UTC. Cross sections of polarimetric radar data support this notion by displaying regions of

vertical motion that correspond to very large enhancements of KDP and ZDR well above the freezing level. Figure 3.17 displays an example of these extensive KDP and ZDR columns for 20:23 UTC.

The total lightning flash rates begin increasing substantially at 20:25 UTC. The flash rates continue to increase until they reach a local maximum of about $180 \text{ flashes min}^{-1}$ at 20:45 UTC. At this point in time, the source densities have increased in altitude, reaching 13 km AGL. The original tornado has also lifted. Once the peak in flash rate is reached, the flash rate begins to gradually descend throughout the rest of the analysis period. The source densities also show a descent, with a cluster of high source density occurring between 21:00 and 21:10 UTC. Interestingly, this cluster of sources is visualized within the same time period in which the second tornado touches down. This second tornado also reached EF-4 strength. While the tornado is on the ground and the flash rates decrease, the –CG flash rate reaches a maximum rate of $22 \text{ flashes min}^{-1}$. Perhaps the decrease in flash rate is associated with a downdraft that is resulting in the fallout of hydrometeors, bringing the negative charge center closer to ground. As these hydrometeors fall, they advect charge to lower regions in the cloud, leading to an isolated region of high LMA source density confined to lower altitudes than were found during the updraft intensification. Falling charge also helps to produce enhanced charge at low levels in the cloud, promoting the occurrence of cloud-to-ground lightning.

Though robust radar data is unavailable for the Cullman supercell, it is encouraging to see a trend of decreasing flash rates in association with the touch down of a tornado. It appears that the combination of different lightning data features can be used to depict what is physically happening within the supercell storm. The flash rate increases in accordance with an updraft intensification that occurs between time periods of tornadogenesis. The LMA source density

plot supports the notion of an updraft intensification as the sources spread out in height and reach extremely high altitudes. The decrease in flash rate and the descent of LMA sources in altitude as well as the confinement of many sources to a specified lower altitude supports the idea that charge is falling through the cloud as a downdraft forms, advecting vertical vorticity to the surface where it is stretched by the updraft and enhanced to form a tornado.

3.3 LANDSPOUTS

3.3.1 FORT LUPTON LANDSPOUT

On July 28, 2014 a DCVZ was visible in both the reflectivity and radial velocity field of the KFTG WSR-88D as shown in Figure 3.18. Afternoon convection over the mountains created outflow boundaries that moved eastward throughout the afternoon. An intersection of these boundaries with the DCVZ is likely what led to the development of convection along the DCVZ on this day. The synoptic environment was not conducive for tornadogenesis. With virtually no CAPE, surface boundaries were necessary for thunderstorm development.

Just before 21:00 UTC convection began firing along the southern portion of the DCVZ over the higher terrain of the Palmer Divide. Convection built northward over the next half-hour. The Fort Lupton landspout was associated with an individual storm cell that first became visible in the reflectivity field along the DCVZ fine-line at 21:33 UTC. CLEAR did not identify this cell until 21:43 UTC due to the storm's initially small area and weak reflectivity values. Thus, the lightning analysis for this case does not begin until 21:43 UTC. It is apparent in Figure 3.19 that this was a very low flash rate storm. Flash rates maximized at 16 flashes min^{-1} and no CG's were detected by the NLDN network. The source density plot does not show as large of a spread in the location of LMA sources with height compared to the source densities produced by supercell storms. This likely is a result of the small number of flashes detected.

Overall, the storm appears to be normal polarity since the majority of the LMA sources are centered at upper levels of the atmosphere, near 10 km.

As the storm cell is first identified by CLEAR, it produces less than 5 flashes min^{-1} . The storm cell grows in size over the next ten minutes in conjunction with an increase in the flash rate. Cross sections of KFTG WSR-88D data show reflectivity echo cores building in altitude as the flash rates increase. The conditional rain rate and the graupel and hail volumes depicted in Figure 3.20 are all continually increasing throughout the analysis period. A ZDR column becomes visible within the last volume of this analysis.

The lightning characteristics within this storm are very different than those produced by the supercellular cases described earlier. The flash rates are increasing in conjunction with the graupel and hail volume within the same time period of the tornado touchdown. These observations are consistent with landspout formation, as discussed in Chapter 1. Landspouts do not require a downdraft. They form when an already present region of vertical vorticity circulating near the surface, or a misocyclone, is stretched upward by an updraft. Since a DCVZ was present on this day, it is likely that there were misocyclones embedded within the convergence zone. As convection formed over the region, one of these misocyclones was likely stretched vertically to produce a short lived landspout. Since flash rates depend on the microphysics and dynamics of a thunderstorm, it is logical that the flash rate would continually increase during the onset of the tornado since the updraft was continuing to loft supercooled liquid water into the mixed phase region to produce more graupel as the tornado touched down.

The tornado only touched down briefly, dissipating well before the maximum flash rate was reached. This is also consistent with landspout dynamics since landspouts require an updraft to maintain vertical vorticity through the storm depth, or appreciable fraction of that depth. As

the updraft produces more hydrometeors, precipitation loading weakens the updraft and eventually leads to a downdraft and the rainout of the storm. A downdraft is detrimental to landspout formation because the source of vertical vorticity stretching is diminished. Thus, as flash rates peak and begin to decrease as the storm weakens, a landspout would not be expected to be maintained by the environment. The complete dissipation of the Fort Lupton landspout is not analyzed because the storm merges with other convection.

3.3.2 DIA LANDSPOUT

On June 21, 2014 three landspouts touched down near the Denver International Airport (DIA). This study focuses on one of the landspouts (EF1) that touched down north of DIA. Slightly before 3pm MDT, the Storm Prediction Center (SPC) produced a mesoscale discussion for the High Plains of Colorado that mentioned a chance of severe weather in the form of large hail and damaging winds. A strengthening surface lee trough and low were centered over the central High Plains in advance of an upper-level shortwave trough. This synoptic set up was conducive for storm development. The deep layer shear was over 30 kts. Shearing winds as a result of these synoptic features combined with microscale topography features may have allowed for the formation of misocyclones on this day.

The DIA landspout storm was first identified by CLEAR at 20:27 UTC. At this point in time, the storm is made up of a cluster of small, weak storms. Within the five minutes between consecutive radar volume scans, the cells quickly organize into one large, more intense cell that moves northward in time. During this five-minute period, the flash rate jumps from less than 10 fpm to 30 fpm, shown in Figure 3.21. Within the next five minutes the flash rate jumps to 65 fpm. By 20:36 UTC a landspout is reported on the ground. The tornado remains on the ground for five minutes before it lifts.

The lightning flash rate continues to increase and reaches its maximum value a few minutes after the landspout lifts. As the storm builds, the conditional rain rate and the graupel and hail volume increase in conjunction with the increasing flash rates, peaking just a few minutes before the flash rate does (Figure 3.22). The peak in graupel and hail volume occurs at the same time in which the tornado lifts. The maximum rain rate lags the peak in graupel production by about five minutes. The increasing flash rates and graupel/hail volume with time supports an updraft intensification occurring as the storm system organizes. Cross sections of polarimetric radar fields show increasing reflectivity values with echo cores reaching higher altitudes with time. This building updraft at the time of tornado formation is consistent with landspout formation. Just as in the Fort Lupton case, the landspout occurs as a result of stretching vertical vorticity found at the surface via an intensifying updraft.

The tornado lifts as the graupel and hail volume reach a peak and as the flash rate begins to level out. Shortly after this peak, the flash rate quickly decreases. This decrease is accompanied by a decrease in the graupel and hail volume while the rain rate remains relatively steady. The decrease in flash rate and graupel volume in conjunction with a relatively large rain rate supports the formation of a downdraft. The graupel produces significant water loading, and begins to fall out as its terminal fall speed becomes larger than the updraft strength. This may spark a downdraft that is accompanied and strengthened by falling rain that may evaporate to cool the parcel. Cross sections of radar reflectivity from the KFTG WSR-88D show a lowering of the echo core during this time period as well. Lightning source densities remain near 10 km during tornadogenesis, but begin to gradually decrease as the flash rates decrease and the inferred downdraft forms. This is consistent with charge carrying hydrometeors falling through the cloud as the updraft weakens and downdraft materializes.

The evolution of the DIA landspout lightning characteristics supports the dynamics of landspout formation. Stretching of vertical vorticity results from an updraft intensification that is visible in the increasing flash rate and increasing graupel/hail volume. The tornado lifts before the maximum flash rate is reached, shortly before the storm begins to weaken. A downdraft is detrimental to landspout formation since there is no source of stretching for already present vertical vorticity. Overall this storm mimics the same characteristics as those found in the Fort Lupton case, but the weakening of the storms is more prominent in this case since the storm remained isolated, allowing for a longer analysis period.

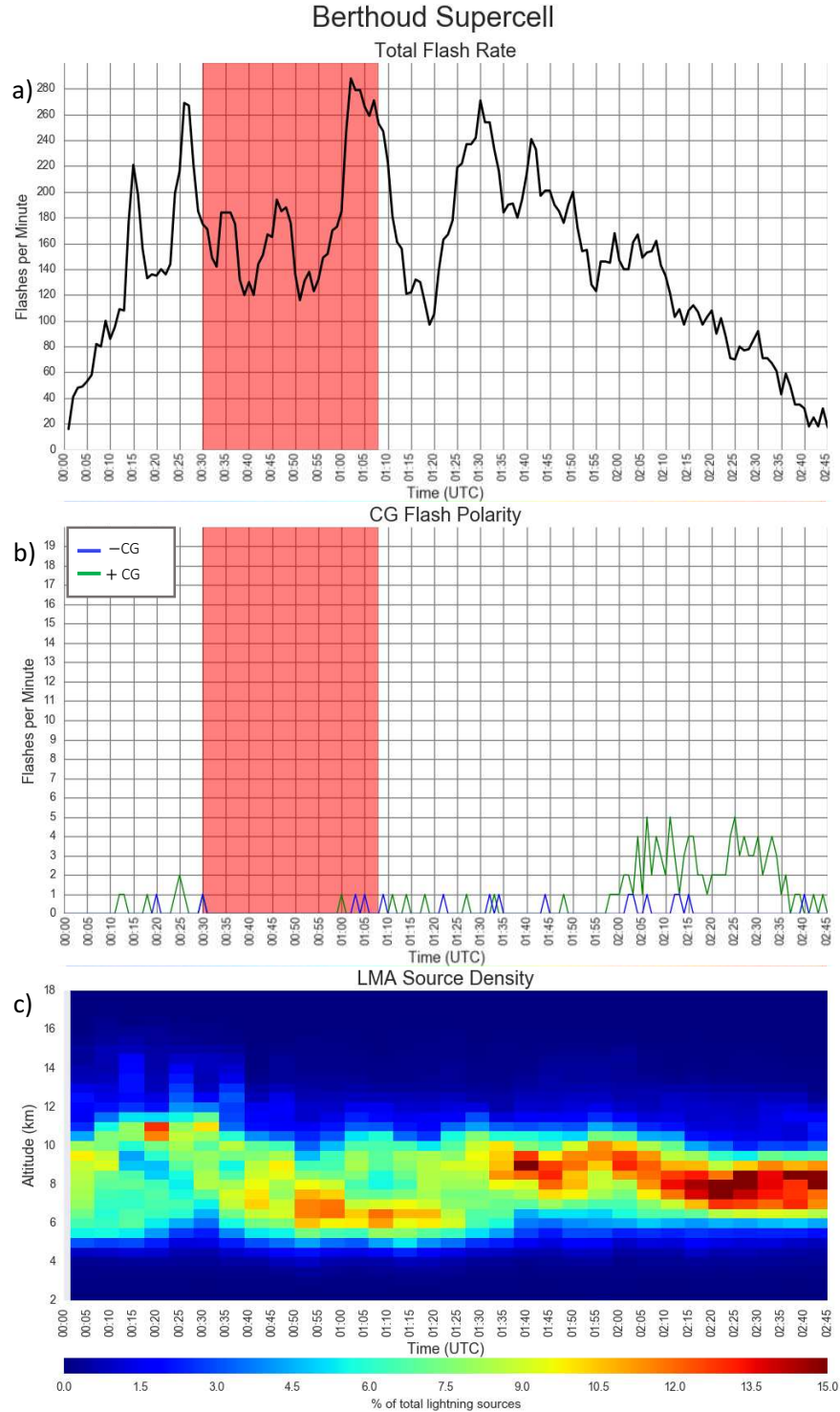


Figure 3.1 Lightning characteristics for the Berthoud supercell that took place on June 4, 2015. a) depicts the total lightning flash rate per minute, b) displays the CG lightning flash rate for positive and negative polarity flashes. Positive polarity CG flash rates are in green and negative polarity CG flash rates are in blue. c) displays the percent of LMA sources within each 0.5 km vertical level at each radar volume time throughout the analysis period. The red shading in a) and b) represents the time period in which the Berthoud tornado was on the ground.

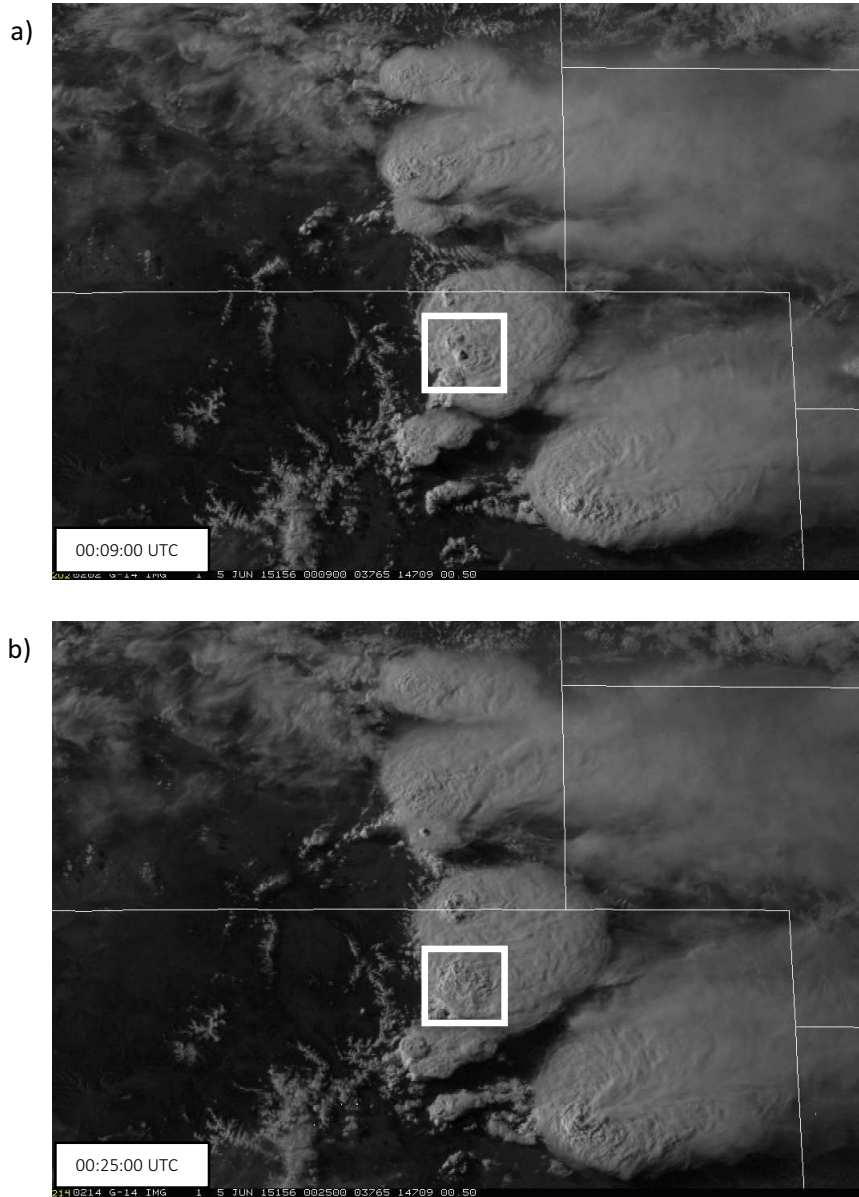


Figure 3.2 Visible imagery over the Berthoud supercell and surrounding storms. The white box outlines the region where the overshooting top is visible for the Berthoud supercell. a) shows well defined shadows and a cauliflower like texture at 00:09:00 UTC while b) shows no such structure at 00:25:00 UTC.

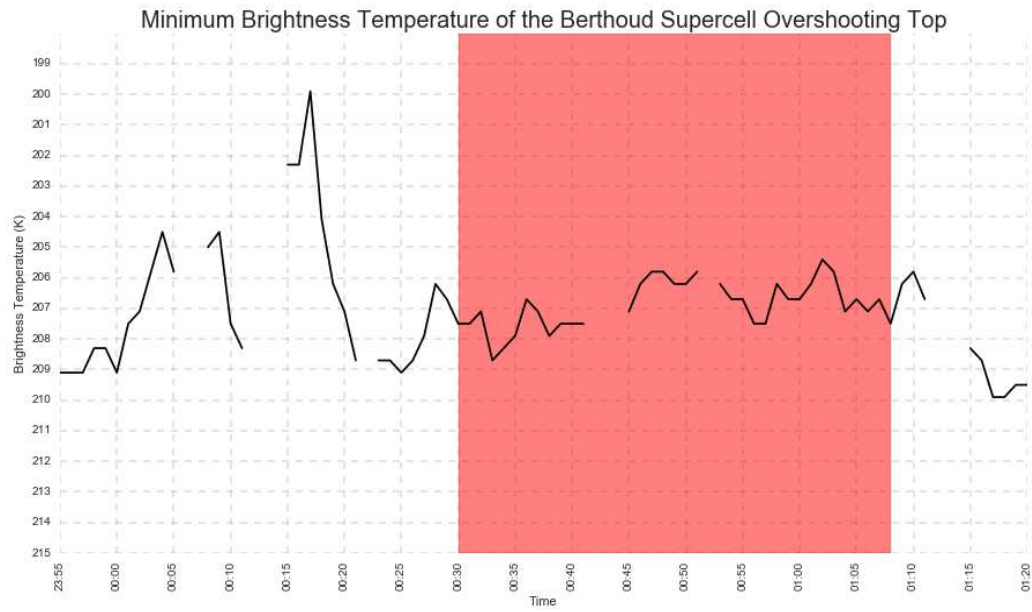


Figure 3.3 Time series of the minimum IR brightness temperature over the Berthoud supercell overshooting top. The red shading represents the time period for which the Berthoud tornado was on the ground.

Berthoud Supercell at 00:21:38Z

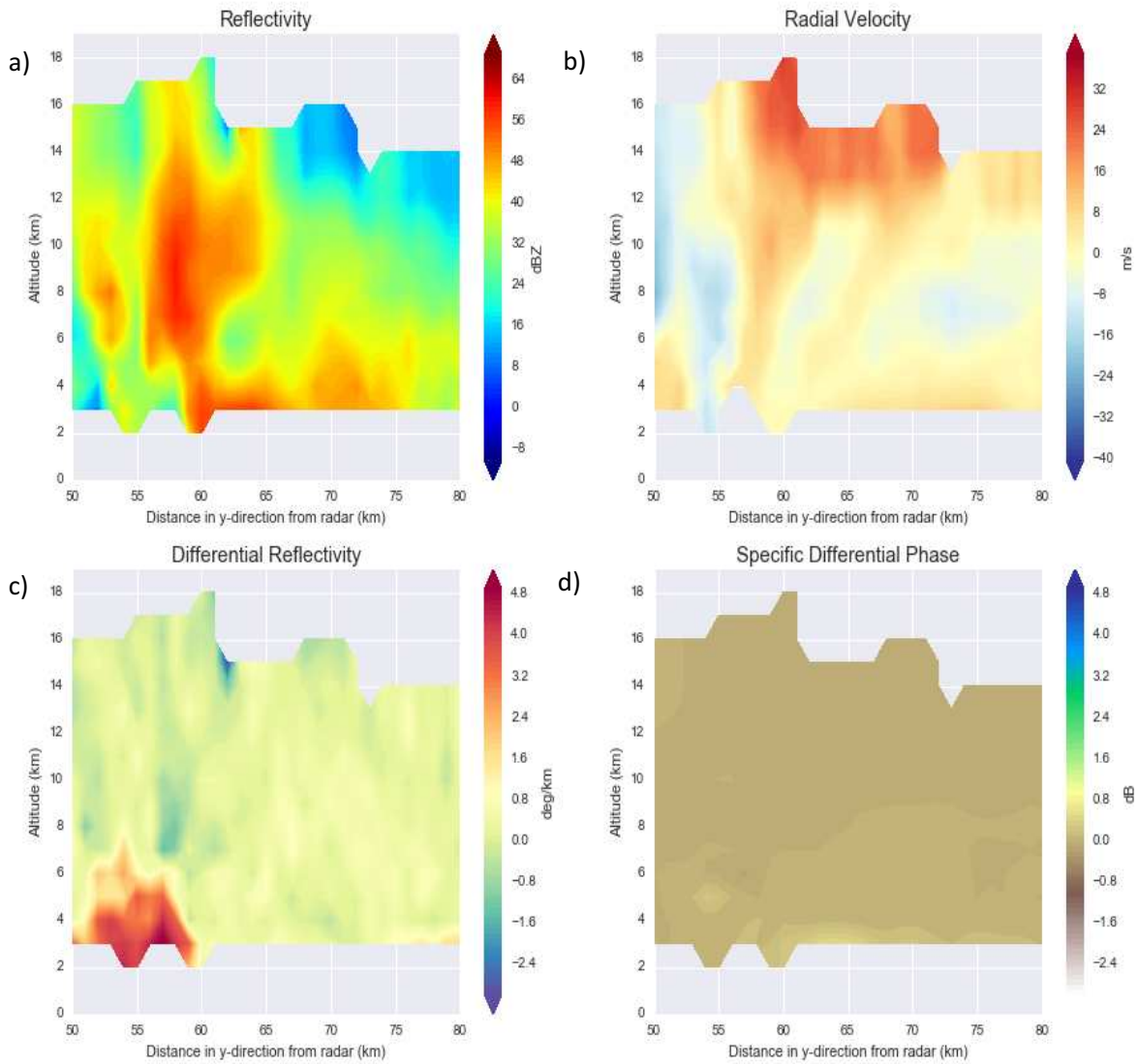


Figure 3.4 Polarimetric radar variable west to east cross sections through the updraft region of the Berthoud supercell for 00:21 UTC. Radar data is from the KFTG WSR-88D, which is located to the southeast of this storm.

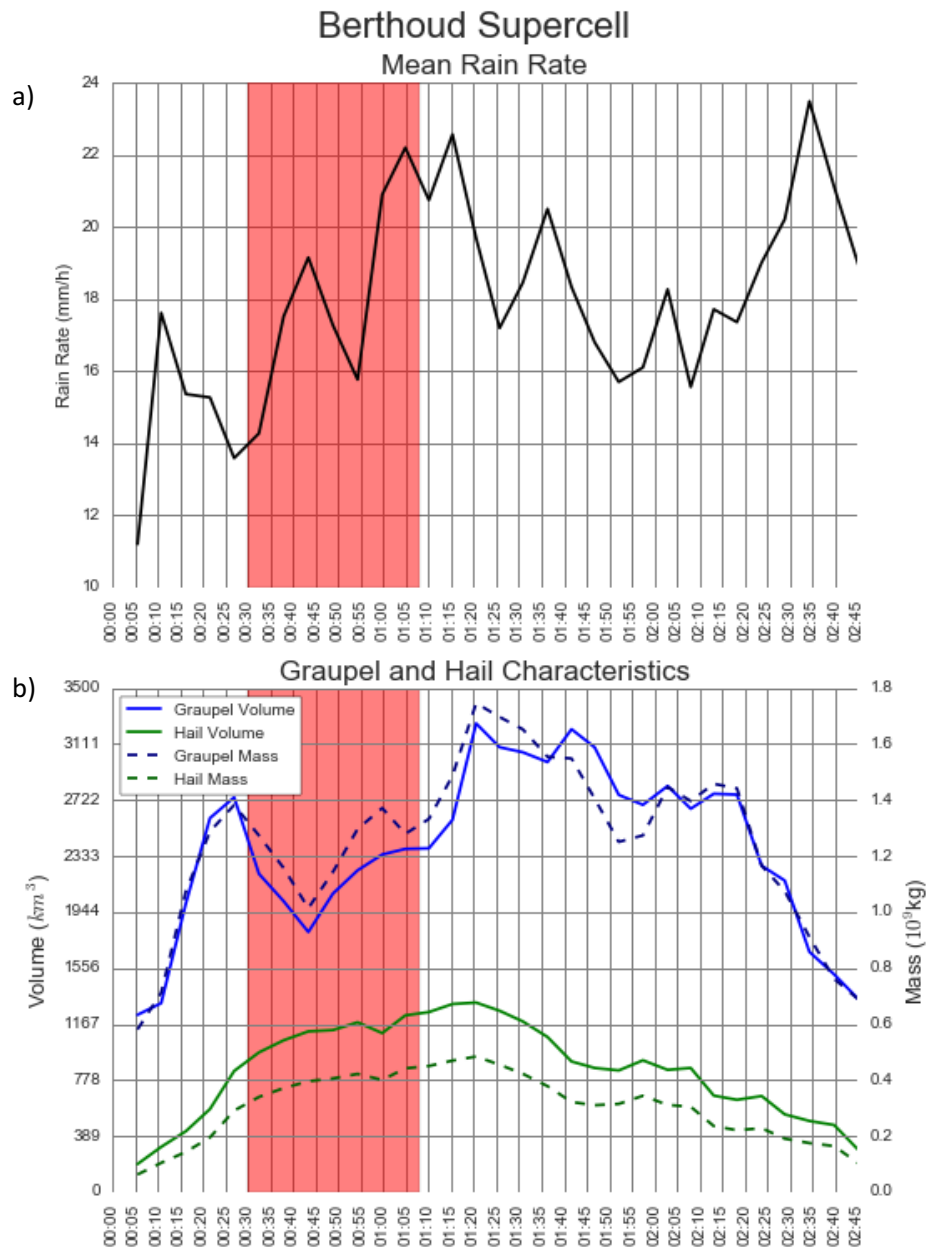


Figure 3.5 Hydrometeor characteristics for the Berthoud supercell. a) represents the mean conditional rain rate throughout the duration of the analysis period with a value calculated for each radar volume time. b) displays the graupel mass (dotted blue line), graupel volume (solid blue line), hail mass (dotted green line), and hail volume (solid green line) found above the freezing level throughout the duration of the storm over time. Red shading represents the time period in which the Berthoud tornado was on the ground.

Berthoud Supercell at 00:43:24Z

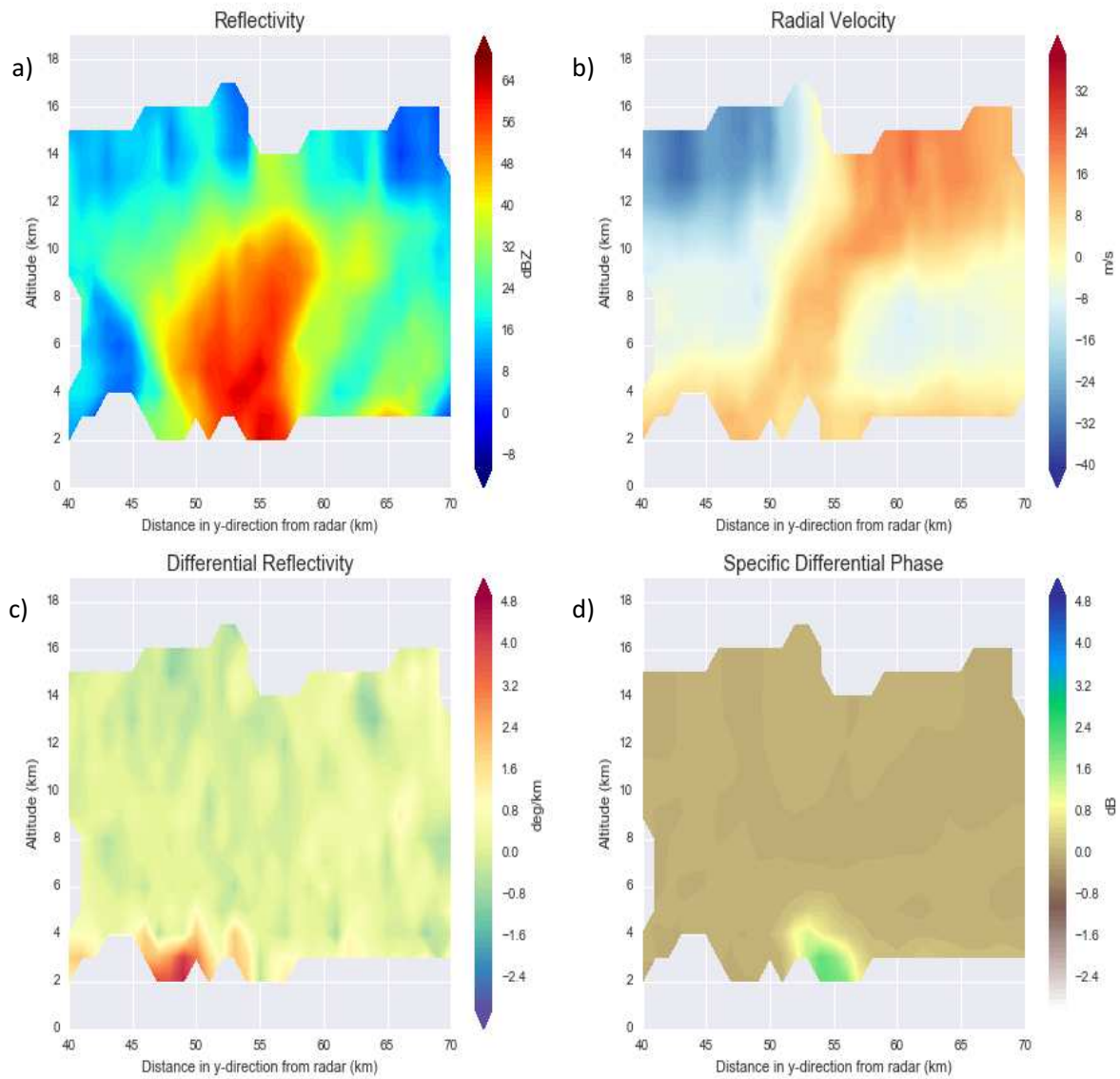


Figure 3.6 Same as 3.4, but for the updraft region of the Berthoud supercell at 00:43:24 UTC.

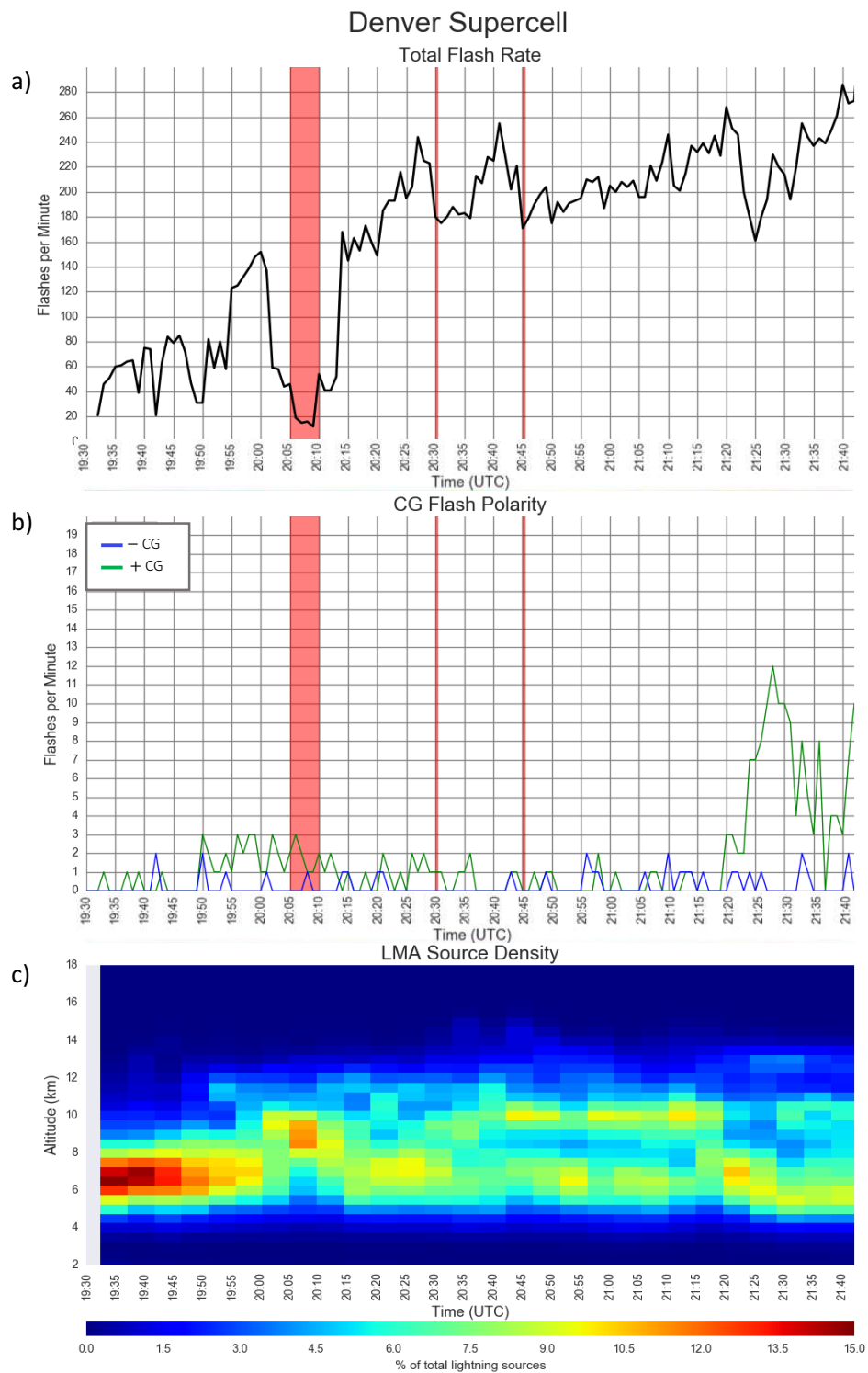


Figure 3.7 Same as 3.1 but for the Denver supercell that took place on May 21, 2014.

Denver Supercell at 19:55:17Z

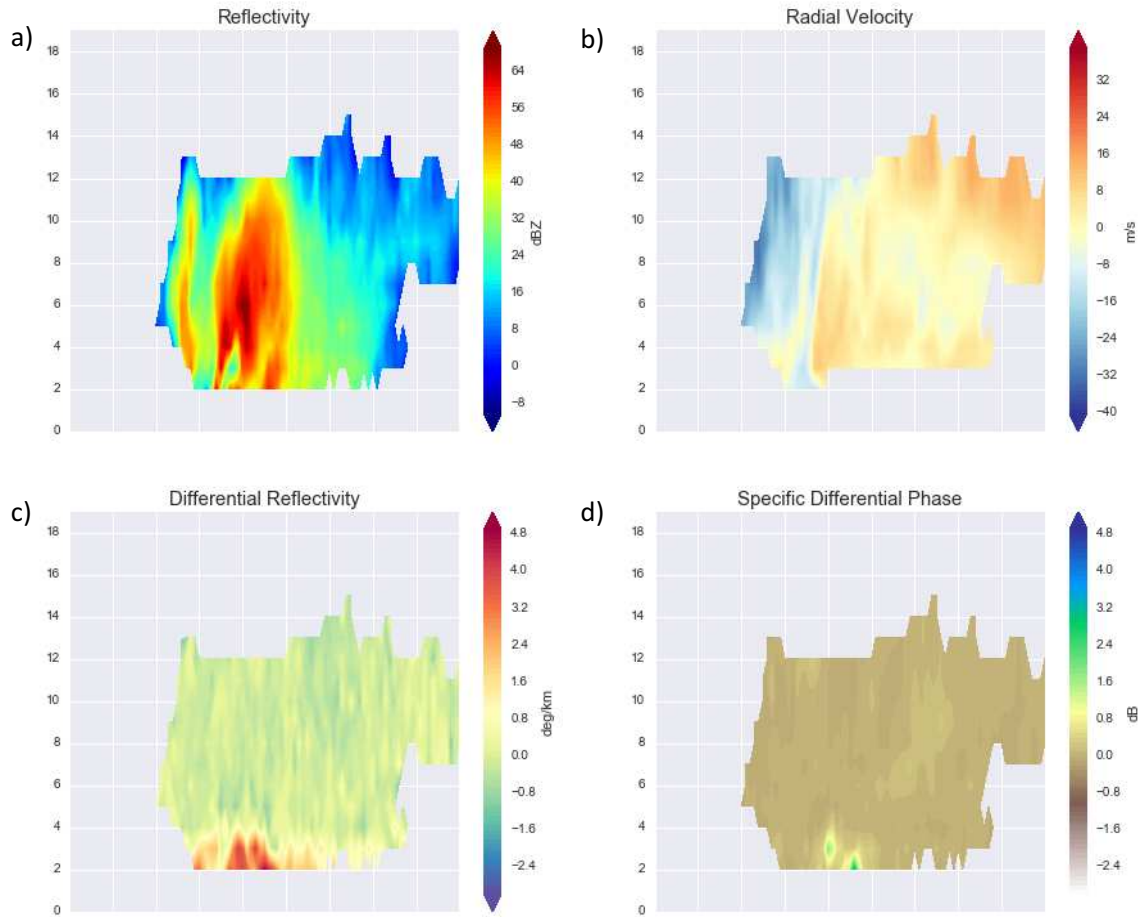


Figure 3.8 Cross sections of polarimetric radar variables through the updraft region of the Denver supercell. Radar data is acquired from the KFTG WSR-88D which is located to the east of this storm.

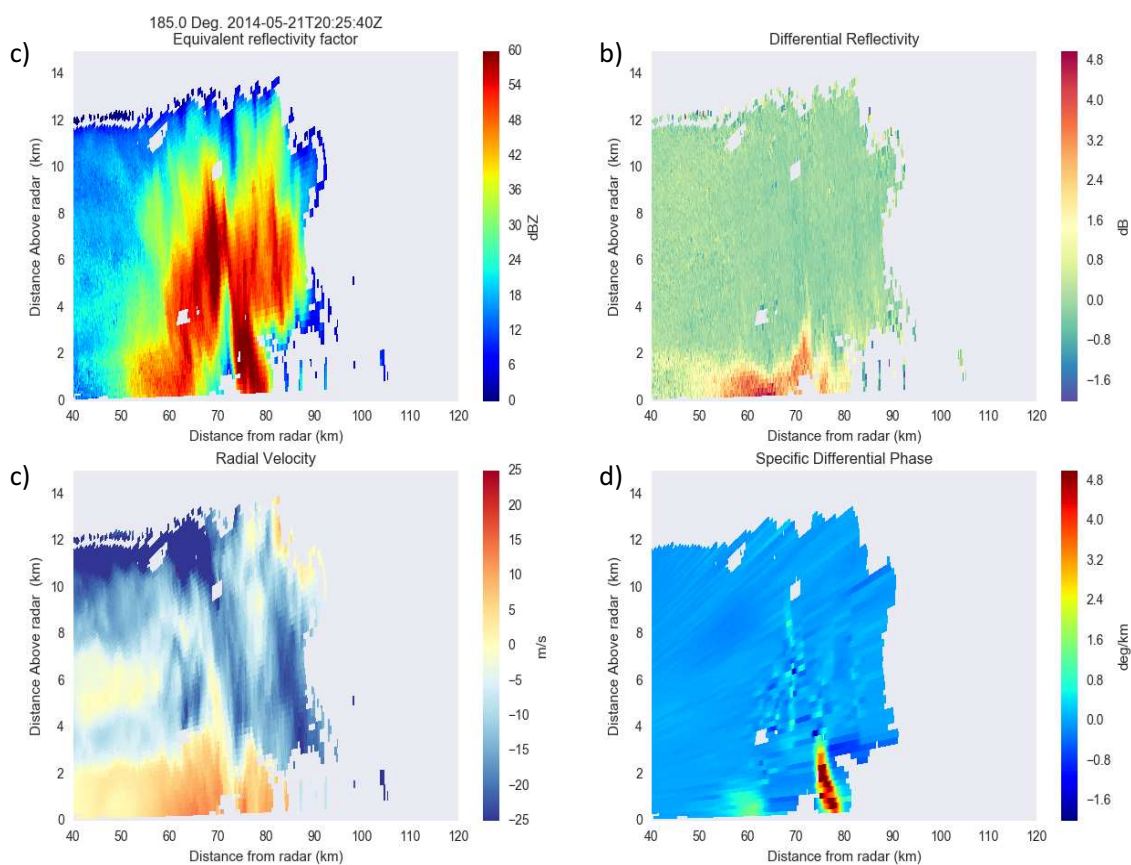


Figure 3.9 RHI of updraft region in the Denver supercell for 20:25 UTC. Polarimetric radar data acquired by the CSU CHILL radar which is located to the north of this storm.

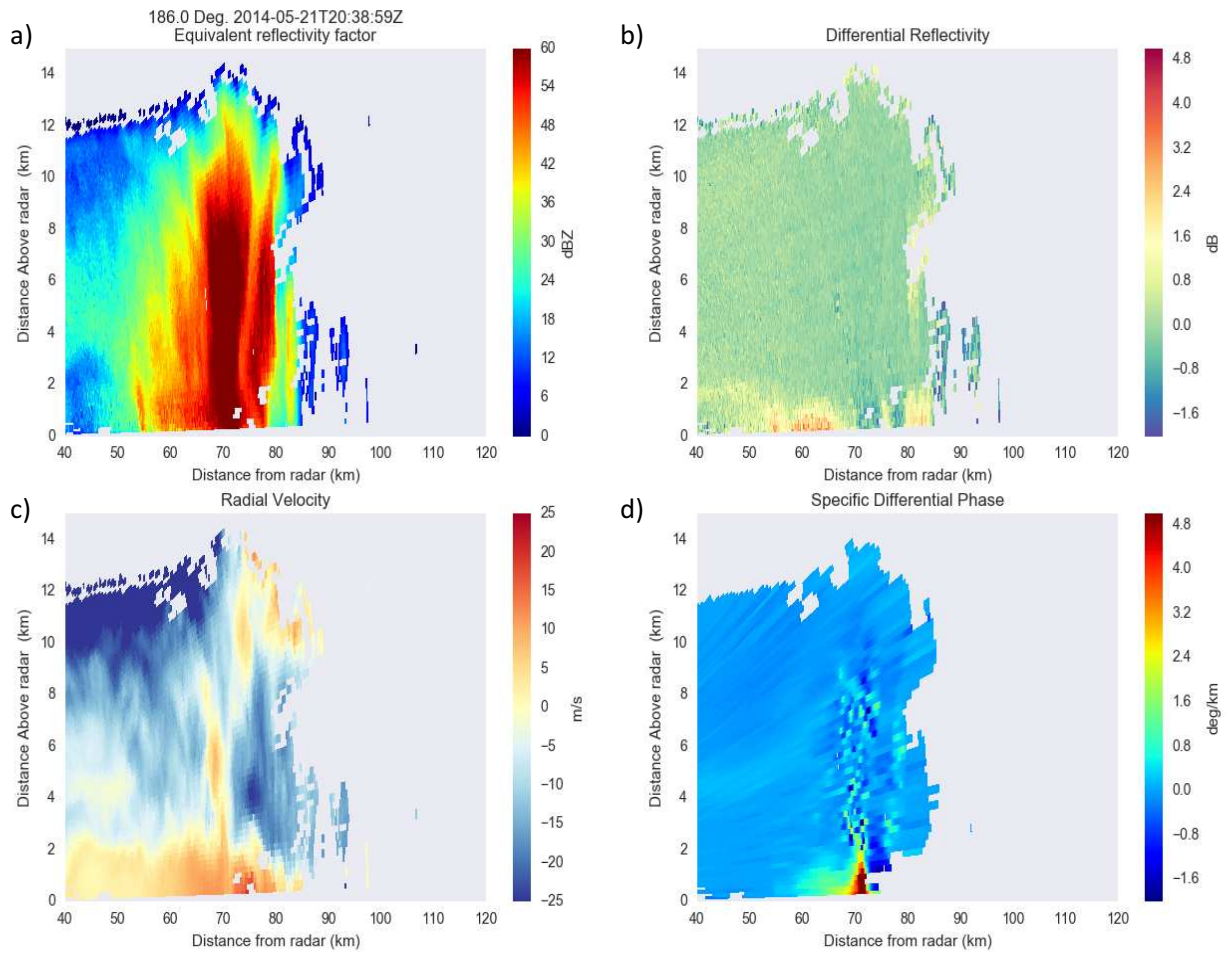


Figure 3.10 RHI of updraft region in the Denver supercell for 20:38 UTC. Polarimetric radar data acquired by the CSU CHILL radar which is located to the north of this storm.

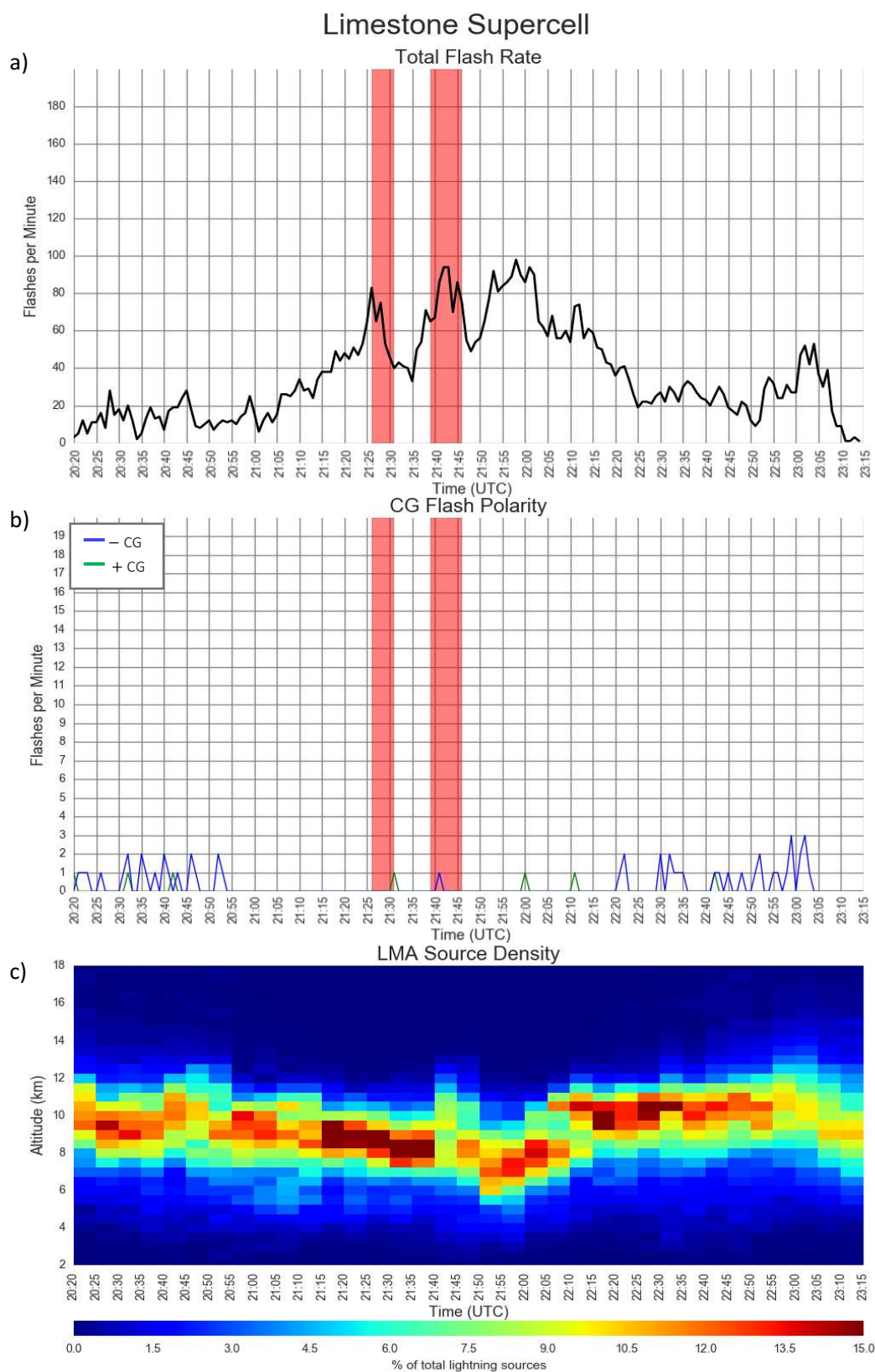


Figure 3.11 Same as 3.1 but for the Limestone, Alabama supercell that took place on March 2, 2012.

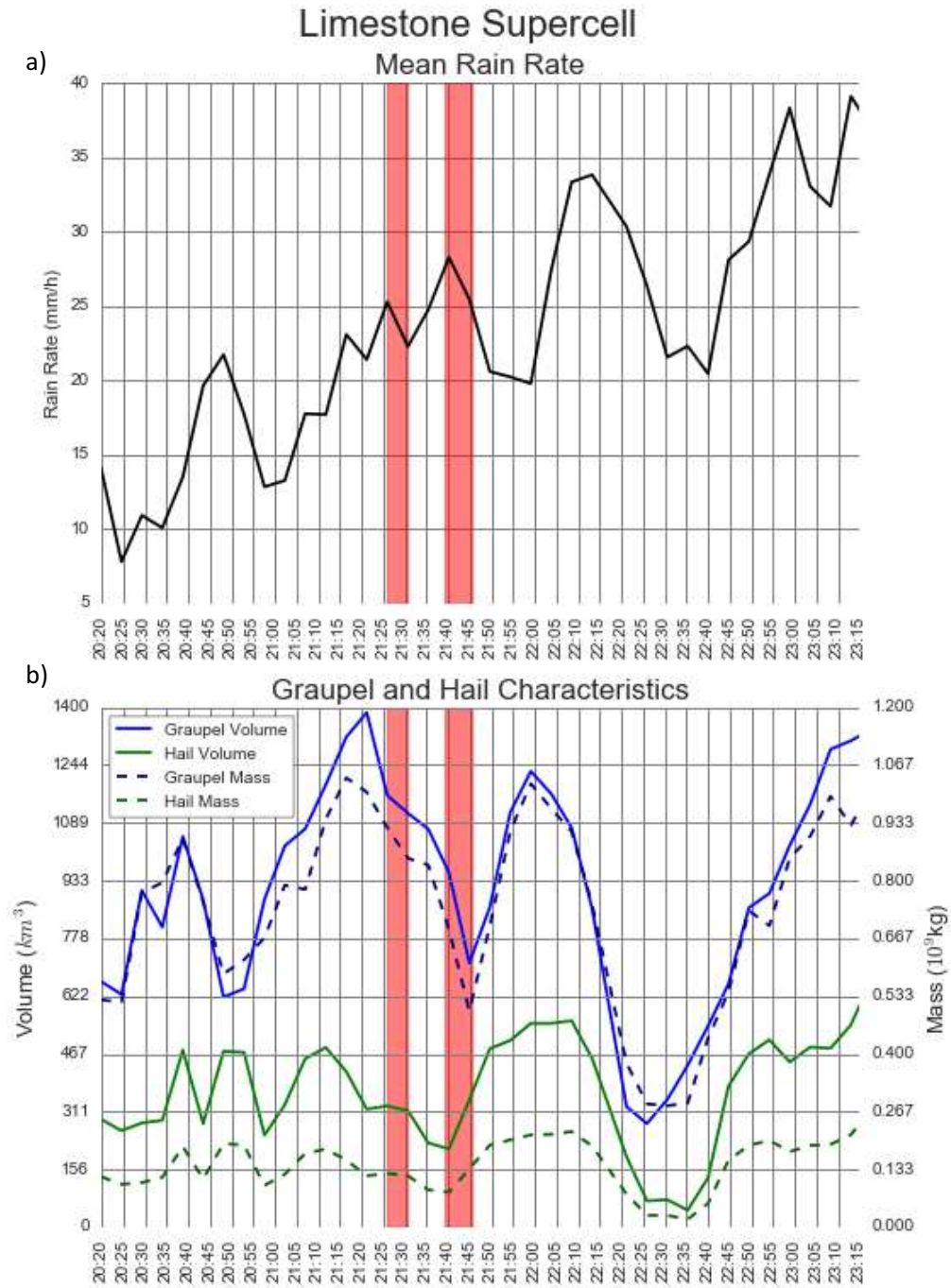


Figure 3.12 Same as figure 3.5 but for the Limestone, Alabama supercell that took place on March 2, 2012.

ARMOR Polarimetric Variables and U/W Wind Vectors near the Updraft at 2116UTC

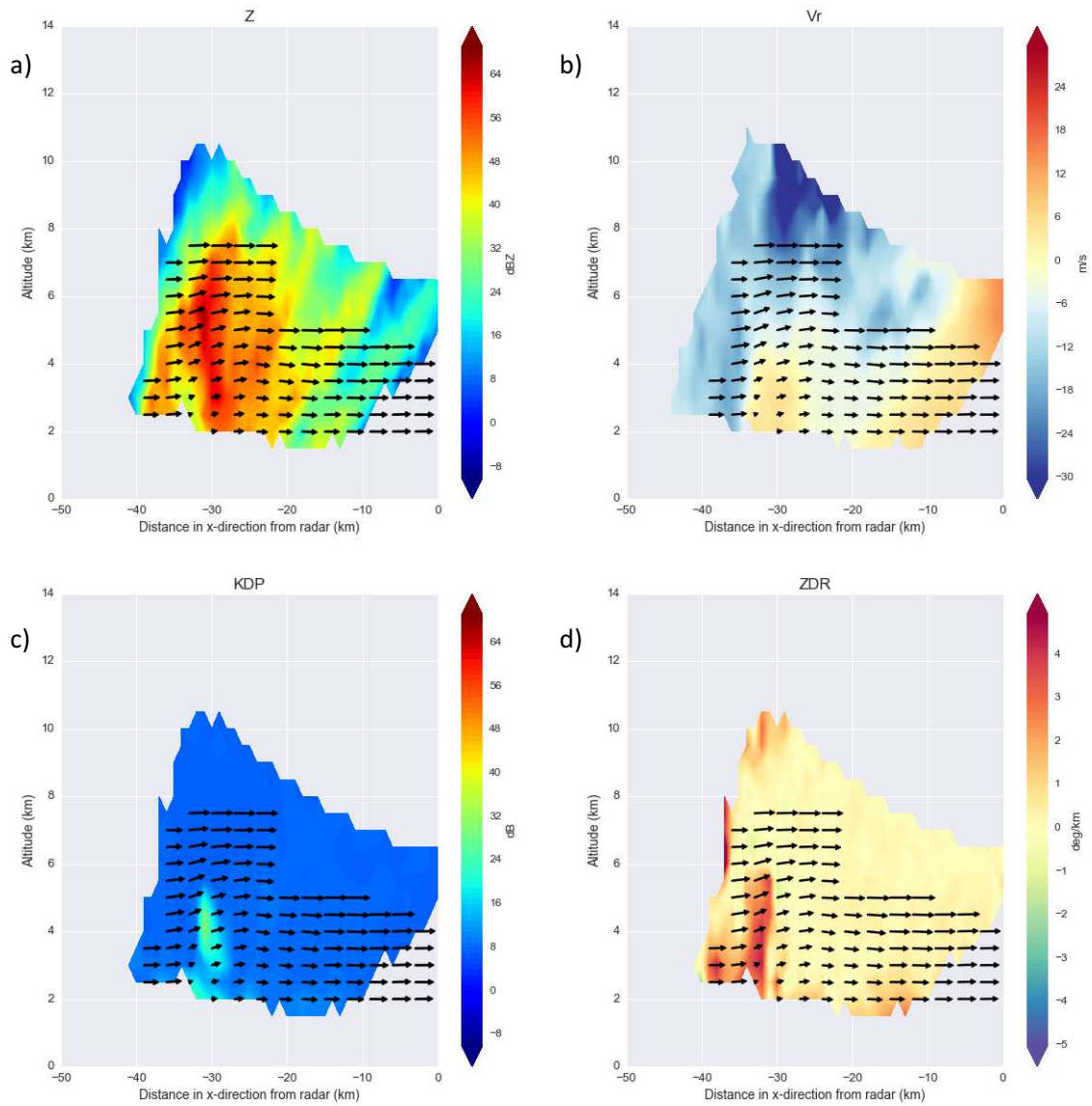


Figure 3.13 Cross section of polarimetric radar variables along the updraft region of the Limestone supercell for 21:16 UTC. Radar data shown is from the ARMOR radar. Arrows represent the U/W wind vector created through dual Doppler analysis between the ARMOR radar and KHTX WSR-88D.

ARMOR Polarmetric Variables and U/W Wind Vectors near the Updraft at 2119UTC

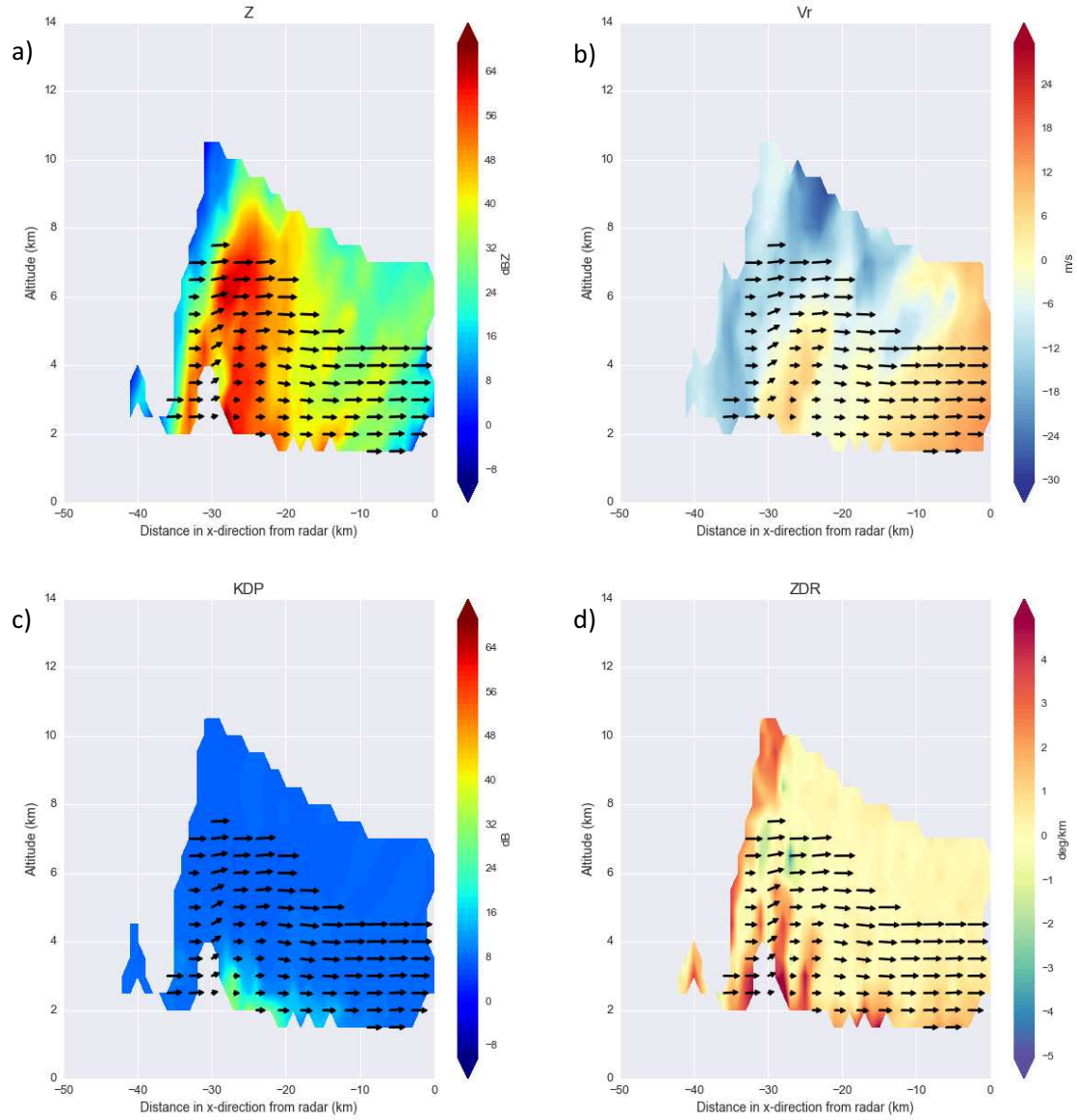


Figure 3.14 Same as 3.13 but for the Limestone supercell at 21:19 UTC.

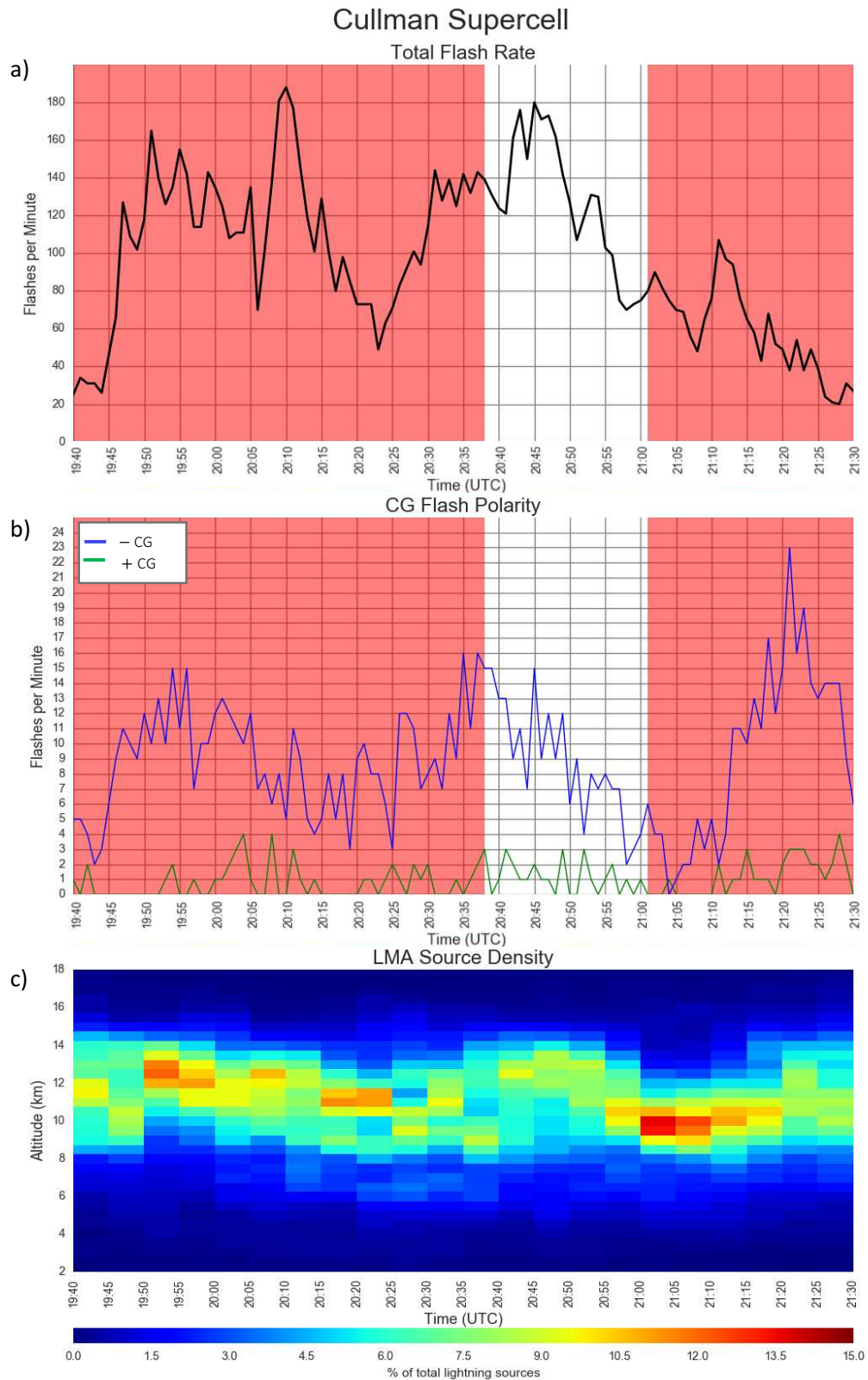


Figure 3.15 Same as 3.1 but for the Cullman, Alabama supercell that took place on April 27, 2011.

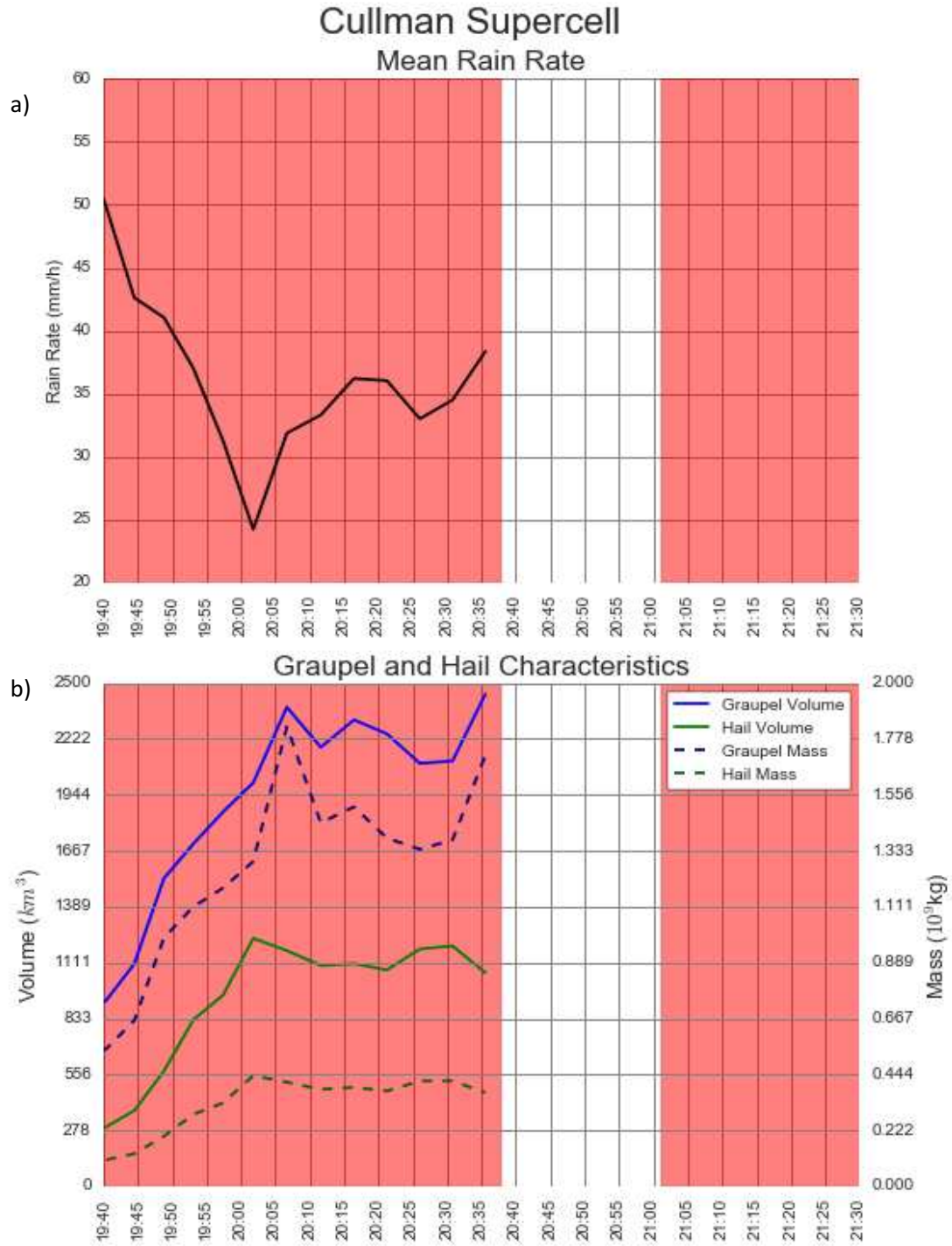


Figure 3.16 Same as figure 3.5 but for the Cullman, Alabama supercell that took place on April 27, 2011.

ARMOR Polarimetric Variables and U/W Wind Vectors near the Updraft at 2023UTC

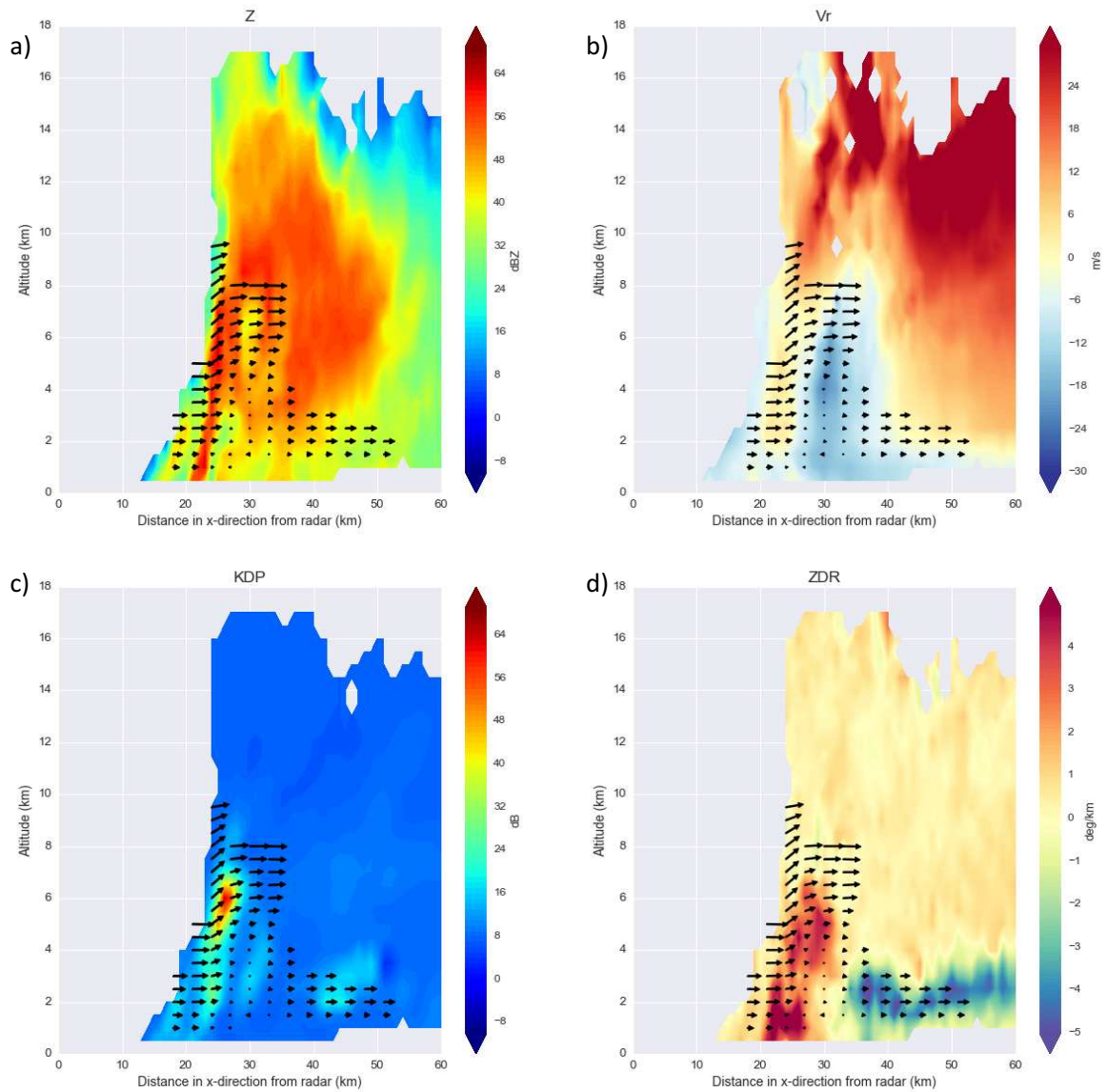


Figure 3.17 Polarimetric radar variable cross sections through the updraft region of the Cullman, Alabama supercell at 20:23 UTC. Polarimetric data displayed is from the ARMOR radar. Dual-Doppler analysis was completed by using the ARMOR radar and the KHTX WSR-88D.

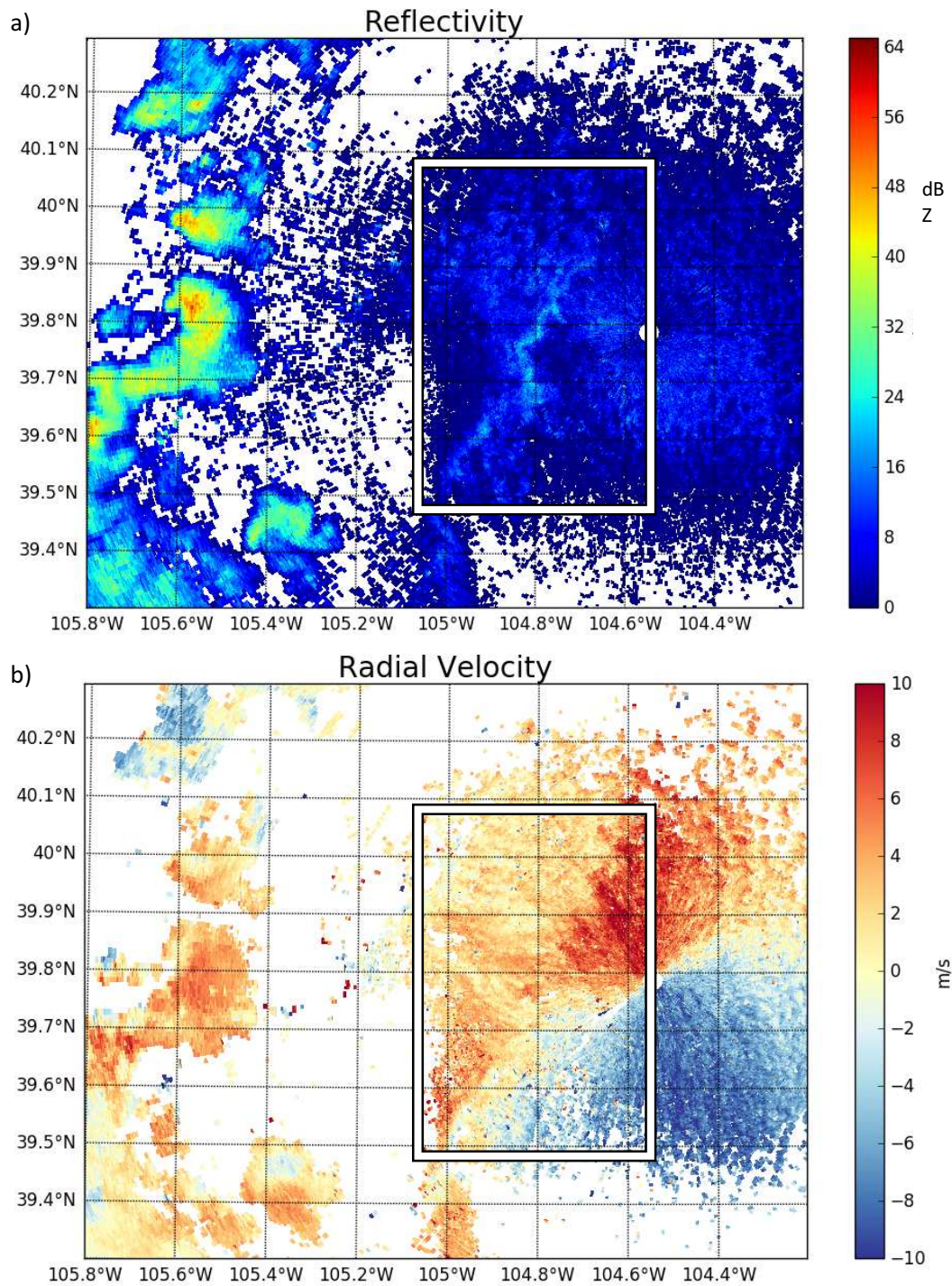


Figure 3.18 Low level PPI scan of the pre-storm environment before the Fort Lupton Landspout. A fine-line can be seen in the boxed region of the reflectivity field in a). This fine line corresponds to a region of convergence depicted in the radial velocity field in b).

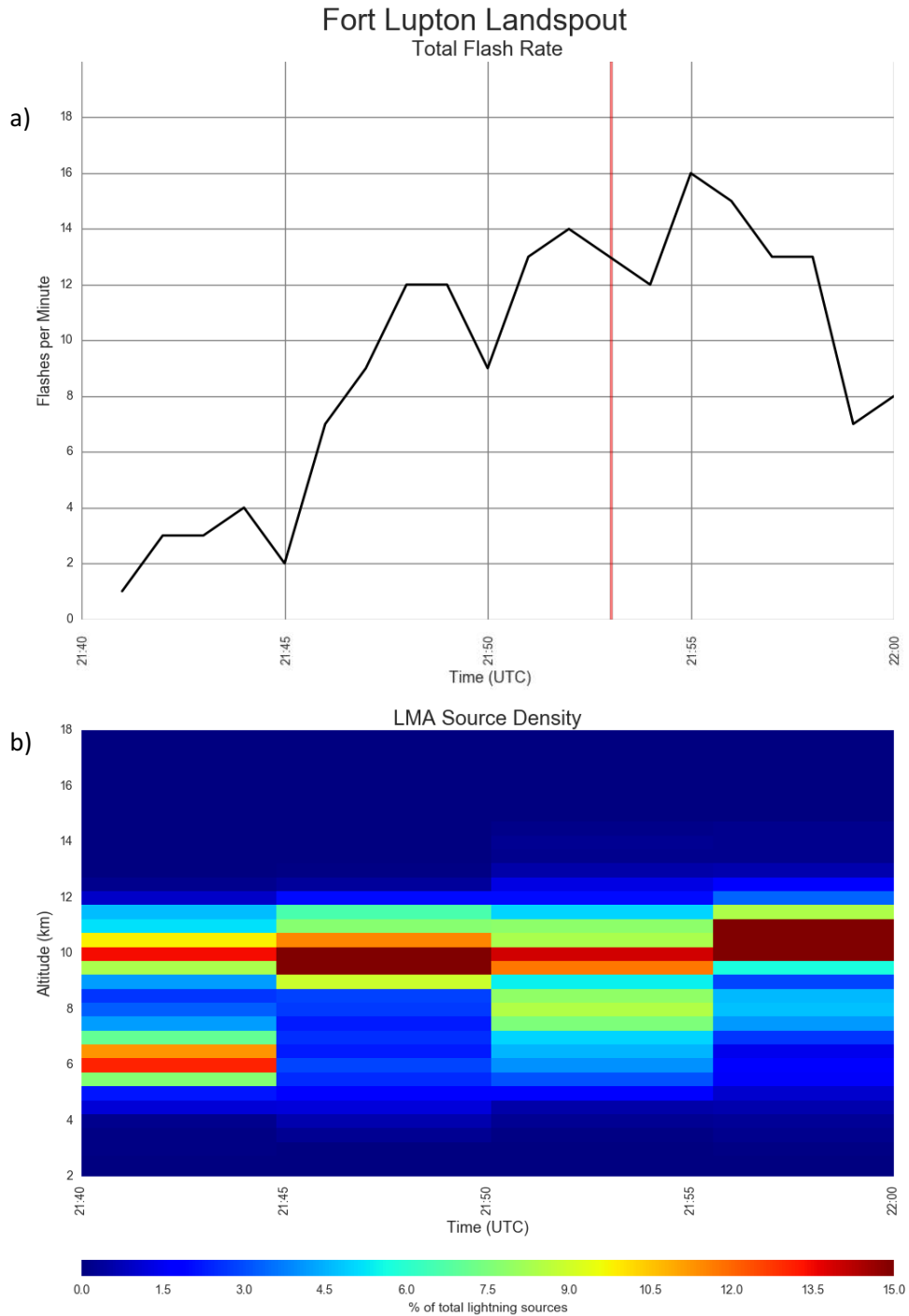


Figure 3.19 Lightning characteristics for the landspout storm that occurred near Fort Lupton, CO on July 28, 2014. a) depicts the total lightning flash rate for the time period in which this storm was isolated from other nearby convection. b) displays the percent of LMA sources found at each 0.5 km height interval in the Fort Lupton storm. Each column represents a radar volume. There were no CG's produced during the analysis period for this storm. Red shading represents the time period in which a landspout was reported.

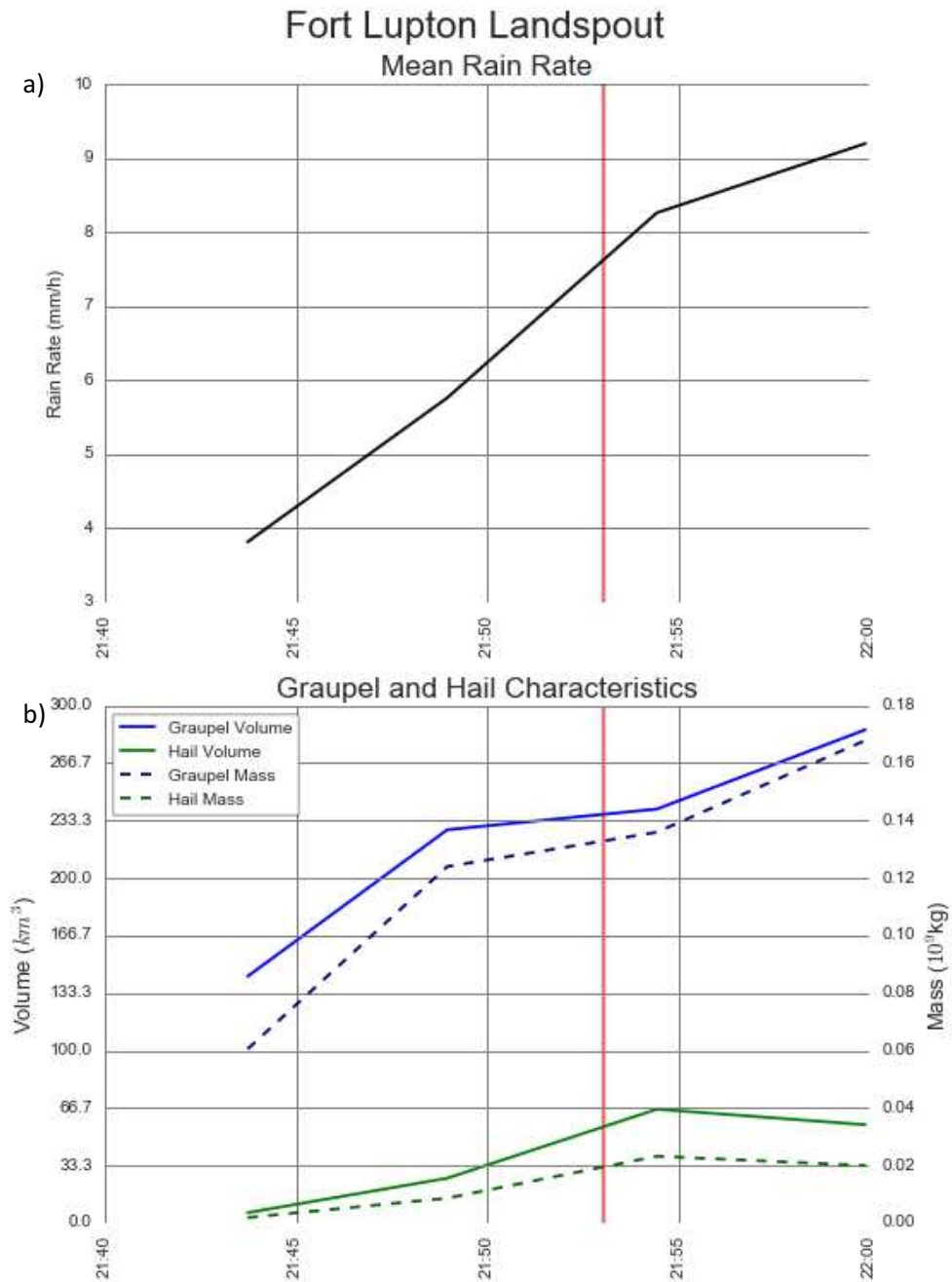


Figure 3.20 Same as figure 3.5 but for the Fort Lupton landspout case that occurred on July 28, 2014.

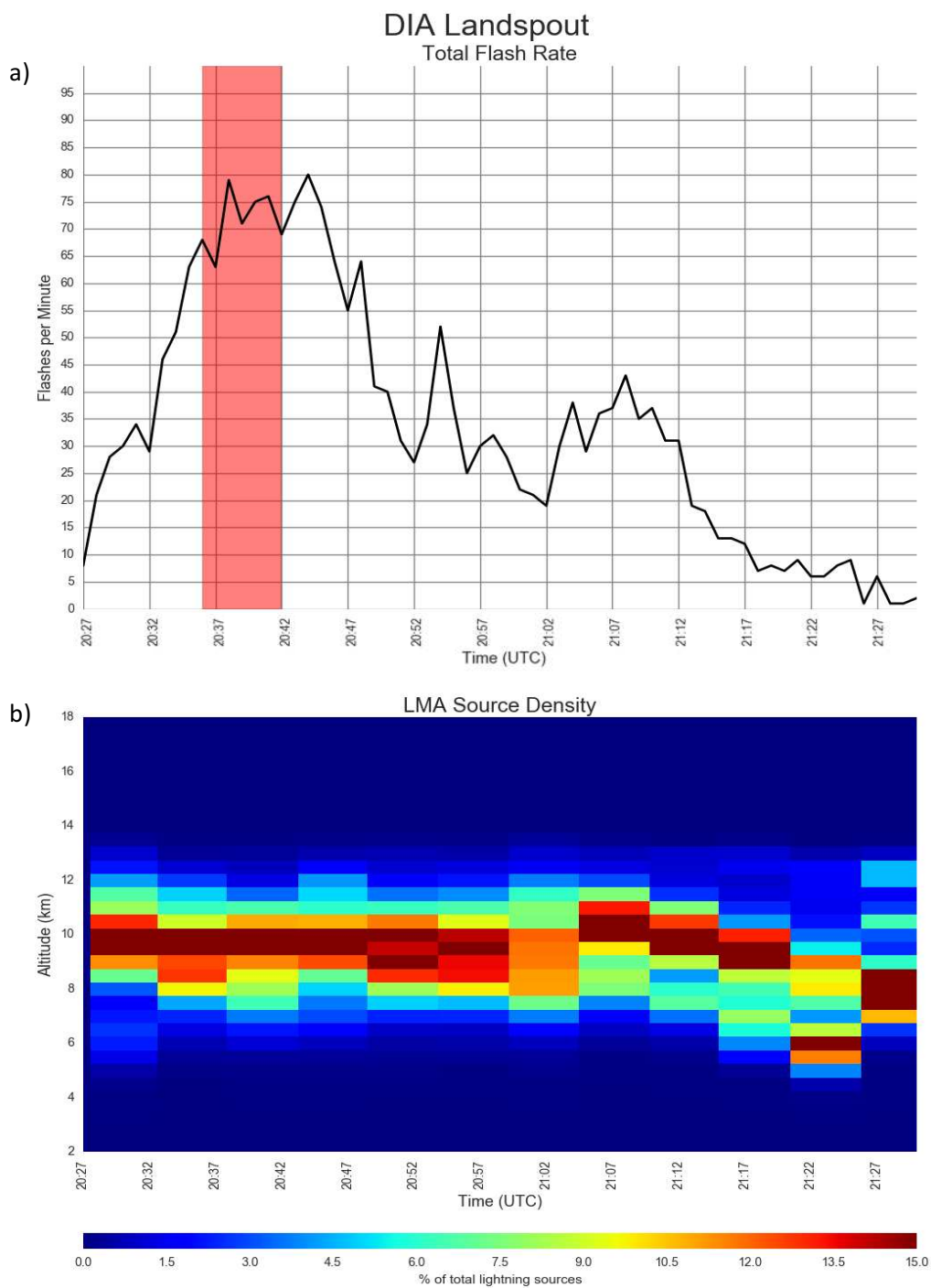


Figure 3.21 Same as 3.19 but for the DIA landspout that occurred on June 21, 2014.

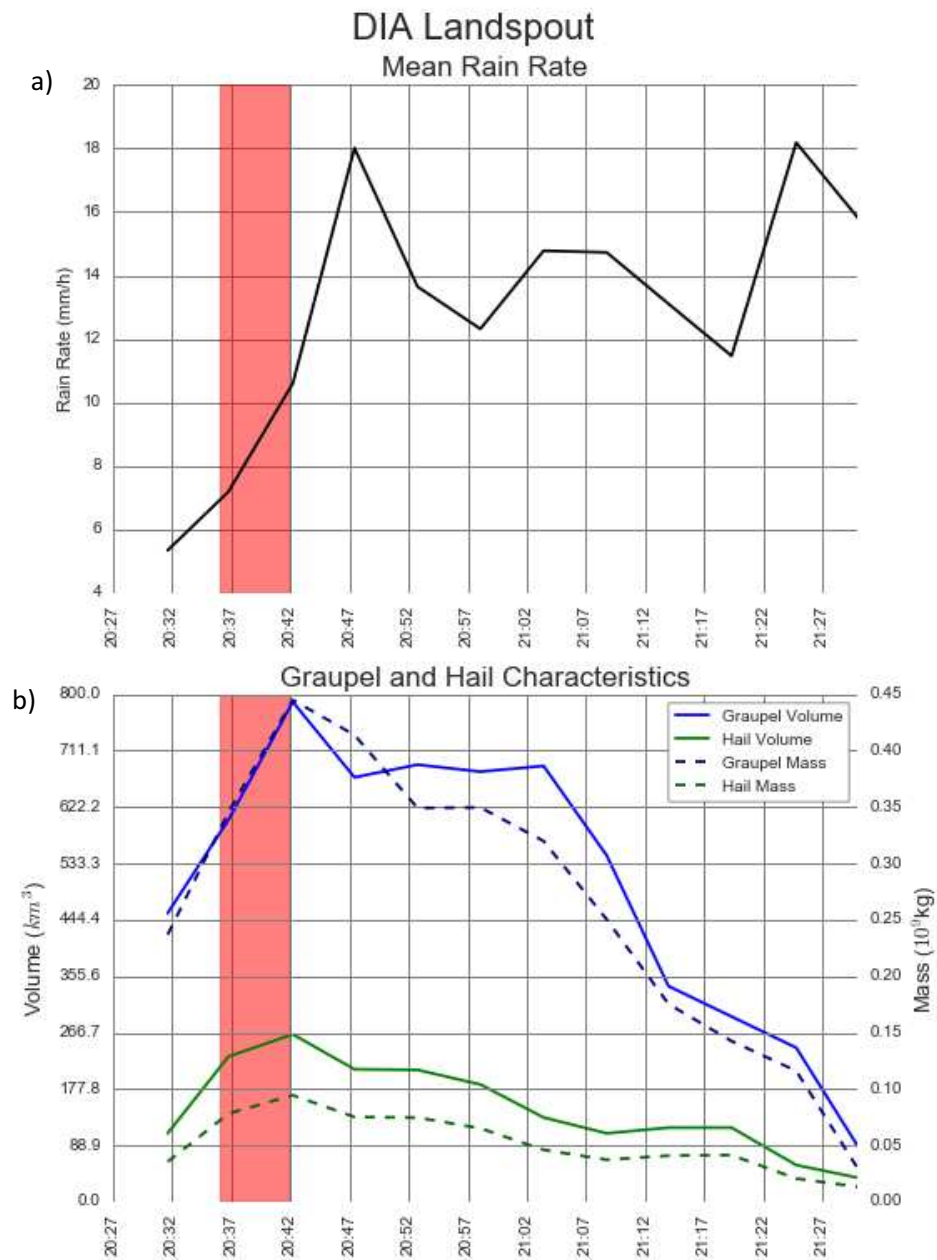


Figure 3.22 Same as 3.5 but for the DIA landspout that occurred on June 21, 2014.

CHAPTER 4: DISCUSSION AND CONCLUSIONS

4.1 LIGHTNING IN TORNADIC SUPERCELLS

Previous studies have found that the total lightning flash rate often “jumps” or increases rapidly before the onset of severe weather (e.g. Williams et al. 1999; Goodman et al. 2005; Schultz et al. 2009, 2011, 2015; Gatlin and Goodman 2010; Stano et al. 2014). The flash rate increases as a result of an updraft intensifying as a storm strengthens and matures to the point that it is able to produce severe weather. This relationship between severe weather and lightning flash rate is useful in nowcasting situations when lightning data are available (Darden et al. 2010; White et al. 2012). Soon, total lightning data will be available for the entire western hemisphere thanks to the launch of the Geostationary Lightning Mapper (GLM; Goodman et al. 2013) on the GOES-16 satellite. GLM will provide 60 s updated data continuously over the entire western hemisphere. Improving our understanding of how all types of severe weather directly relate to the evolution of lightning in a storm will help forecasters best utilize the new data provided by GLM.

Supercellular tornadoes form in association with a mesocyclone within a mature supercell. Analysis of the Berthoud and Denver storms in this thesis suggests that in scenarios that multicellular convection organizes into a supercell, this organization may be accompanied by a sharp rise in flash rate. The Denver supercell began as a cluster of thunderstorms with total lightning flash rates near 20 flashes min^{-1} . The flash rates increased in conjunction with the organization of these storms. A peak in flash rate occurred when a mesocyclone first became evident within the Doppler radar radial velocity field. The Berthoud supercell showed the same trait. In this case, the Berthoud supercell had weakened significantly during the touchdown of

the Berthoud tornado and lost its defined mesocyclone. The former supercell appeared to maintain more of a multicellular structure. As the storm reorganized to maintain a defined mesocyclone, the flash rates showed a rapid increase. This relationship corroborates other studies (e.g. Darden et al. 2010; Stano et al. 2014) and could serve useful to forecasters in determining which storm clusters will organize into more intense convection.

Charge separation that produces lightning results from the collision of graupel and ice crystals in the presence of supercooled liquid water (e.g., Reynolds et al. 1957). Intensifying updrafts lead to the formation of more graupel and ice crystals as more liquid water is made available for riming growth. Should larger supercooled drops be lofted into the mixed phase region, they rapidly freeze to become graupel particles. Intensifying updrafts also lead to more turbulence, which may promote more collisions between falling graupel and rising ice crystals, leading to more charge separation. These processes cause flash rates to increase as a mesocyclone develops. However, this characteristic is not limited to the formation of a mesocyclone. This analysis demonstrates that this process occurs any time that an updraft intensifies sufficiently to produce ice and electrification, resulting in lightning flash rates increasing regardless of the storm's maturity.

As the updraft of the Berthoud supercell first intensifies before the onset of the tornado, the flash rates increase rapidly from 20 flashes min^{-1} to nearly 270 flashes min^{-1} over a 25-minute period. This prolific increase in total lightning flash rate is accompanied by increasing graupel and hail volumes, a decreasing minimum brightness temperature over the overshooting top, and an increase in the height of the VHF sources. The increase in graupel volume is consistent with updraft intensification, as discussed above. The increase in updraft intensity also causes the location of the VHF sources to rise with it, creating the increase in VHF source density heights.

Rising VHF sources do not appear each time an updraft intensifies. The Berthoud storm displays other periods of intensifying updrafts that correspond to rises in total lightning flash rate and increases in graupel volume, but these time periods do not show an increase in the VHF source density height. For example, just before the Berthoud tornado lifts, the total lightning flash rate of the Berthoud supercell increases rapidly. The VHF sources show no strong upward ascent. They appear to “spread out” such that the sources are not highly concentrated at a specific elevation. This likely occurs because updraft intensification promotes enhanced turbulence which helps to create smaller pockets of charge along the periphery of the updraft. More frequent, smaller flashes occur causing the VHF sources to appear throughout the atmospheric column (Bruning and MacGorman 2013).

Similar to the beginning stages of the Berthoud supercell, the Denver storm shows an initial rise in the VHF sources in conjunction with increasing flash rates as the storm first develops, prior to the touchdown of the first tornado. There is a spreading out of the source densities as the flash rates recover after dropping significantly during the onset of the first tornado, but any other increase in flash rate is not accompanied by significant changes in the VHF source densities. The Cullman case on the other hand, shows VHF sources varying in height with the changes in the total lightning flash rate. Increases in flash rate are accompanied by increases in VHF source heights, while decreases in flash rate are accompanied by the descent of VHF source heights. The Limestone case did not display this relationship. The first increase in flash rate that occurs prior to the formation of the first tornado shows no change in the source density heights. The VHF sources remain confined to a relatively thin layer through the duration of the storm. There is a slight spreading out of the VHF sources with height as the flash rate increases during the onset of the second tornado.

In each supercell analyzed, the flash rate consistently increases during radar indicated periods of updraft intensification, consistent with the relationship between intensifying updrafts and increasing flash rates found in past studies. Steiger et al. (2005, 2007) found that when storms were most intense, lightning occurred at the highest altitudes during the storm's history. This study shows that this is not always true. The Denver and Berthoud supercells show an increase in the source density height as the storm develops, but the VHF sources do not vary during re-intensification of the storms. The Cullman supercell tends to support Steiger et al. (2005, 2007) as the VHF source rise and fall with increasing flash rates and thus intensifying updrafts. The Limestone supercell is unique in that the source heights decrease as the storm intensifies. The storm reverses polarity from normal to anomalous, perhaps associated with higher amounts of supercooled water aloft, thereby promoting positive charge on graupel. Though the evolution of these storms do not fully support the findings in Steiger et al., it is interesting to note that the Cullman storm, which produced the highest echo top heights and most severe tornado, displayed VHF sources well above 10 km. The other cases showed VHF sources remaining near 10 km.

Multiple studies have indicated that the total lightning flash rate typically decreases during the onset of a tornado (Williams et al. 1999; Steiger et al. 2005, 2007; White et al. 2012). These studies agree that this decrease in flash rate is likely a result of the updraft weakening and a downdraft forming since supercellular tornadoes require a downdraft to advect vertical vorticity to the surface where it can be stretched by the updraft. Steiger et al. (2005, 2007) suggests that lightning decreases in height as downdrafts form before tornadogenesis. This study supports this notion along with the idea that flash rates decrease during tornadogenesis. The Berthoud tornado is an ideal example. Just prior to the onset of the tornado, the flash rates

decrease sharply and there is a descent of the VHF source densities throughout the duration of the tornado. Decreasing graupel volume in conjunction with increasing rain rates are consistent with the formation of a downdraft.

The Denver supercell shows an extreme decrease in flash rate prior to the onset of the first tornado. The VHF sources begin to decrease in height only after the tornado is on the ground. The VHF sources show no definitive trend once the second and third tornado form, but there is an apparent rise in flash rate and then decrease prior to the onset of both of these tornadoes. Only a portion of the Cullman supercell could be analyzed, but the onset of the second tornado also supports the notion that lightning data is suggesting downdraft formation. The second tornado in the analysis touches down while the flash rates are decreasing. There is also a descent in the source densities as tornadogenesis begins. The first Cullman tornado begins before the analysis time, so information was not attained about the flash rates and its initial formation.

The Limestone supercell shows slightly different trends when it comes to total lightning flash rates and VHF source heights. The first tornado reported within the analysis period touches down right as the peak in flash rate is reached. The second tornado occurs while the flash rates increase to a peak and then begin to fall. The VHF sources show a very gradual descent during the period of tornadogenesis, but the descent is not significant enough to discern any obvious trend. Evidence of a downdraft occurs as graupel volume decreases and rain rate increases throughout the duration of tornadogenesis, but the flash rate does not follow the trends found in the other three supercellular cases. Part of this may be due to error in spotter-reported times in which a tornado was on the ground.

Though the Limestone supercell shows differing trends, it is noteworthy that three of the four cases analyzed depict signs of downdraft formation within the lightning data. Awareness of these trends could serve useful to forecasters during nowcasting situations. It is interesting to note that the trends in flash information were not as robust during secondary tornadogenesis during the Denver supercell. This is likely a result of the fact that secondary tornadogenesis is easier to achieve because the environment left behind by original tornadogenesis supports the formation of horizontal vorticity that can be tilted vertically, and subsequently advected to the surface and stretched to form subsequent tornadoes (Davies-Jones et al. 2001).

Decreases in graupel volume also were found to coincide with the formation of a downdraft, which may serve useful to forecasters if sufficient radar data along with HID information is available. Using graupel volume calculations in conjunction with total lightning flash rates would provide forecasters with valuable information that could be used to predict the formation of a downdraft. Analysis completed in this thesis suggest that rises and falls in graupel volume are less abrupt than rises and falls in the minute by minute flash rate. Graupel volumes also tend to display a less noisy curve. As a result, this analysis shows that a decrease in graupel volume above the freezing level may be a more reliable way of determining when a downdraft forms. Ideally, witnessing a decrease in graupel volume at the same time as a decrease in lightning flash rate would be the most reliable way of determining downdraft formation. Radar data is not always available to forecasters due to power outages or blockage from high terrain. With the launch of GOES-16 and the availability of total lightning data from GLM, it is important to define relationships between lightning information and storm dynamics as GLM will continually provide lightning coverage in areas that may not have access to weather radars.

No trend in the production of CG flashes during tornadogenesis is indicated in this study. The apparent lack of relationship between tornado formation and CG production has been well reported in the literature (Perez et al. 1997; MacGorman et al. 1989; Keighton et al. 1991; MacGorman and Nielsen 1991, Strader and Ashley 2014). Both the Denver and Berthoud supercells show the formation of an EOSO at the end of their lifetime. The Berthoud supercell displays a decrease in flash rate and a decrease in VHF source density heights as the number of +CG's quickly increases. The Denver supercell does not show a decrease in flash rate, but there is a descent in the VHF source heights as the +CG rate increases substantially. The EOSO in both of these cases is linked to the fallout of precipitation-sized ice particles, as indicated by radar cross sections. The Berthoud case also shows a decrease in the graupel volume, most likely resulting from graupel fallout. As the graupel falls through the cloud, negatively charged hydrometeors fall out of the cloud, leaving behind a large amount of positively charged hydrometeors that fall closer to the surface (indicated by the descent in VHF sources). This descent in the positive charge layer likely promotes the formation of +CG flashes.

An interesting final note about this analysis is the difference between the Alabama and Colorado supercells. Colorado is located in the High Plains region which is characterized by high cloud base heights in a relatively dry, warm environment. High cloud base heights promote the formation of broad updrafts as parcels have more time to expand as they move upward. These high cloud base heights result in small warm cloud depths, which promote enhanced ice formation in the reduction of robust warm rain processes. These traits are clearly indicated in comparisons of graupel volume. Increases in graupel volume indicate enhanced graupel production. Graupel production in the Colorado cases is significantly larger than the Alabama cases. This enhanced graupel production corresponds to larger flash rates in the Colorado storms

overall as well. Regardless of these differences, the lightning trends are similar in both locations, suggesting that information that can be gained from lightning trends holds constant across varying environmental backgrounds.

4.2 LIGHTNING IN LANDSPOUTS

Two case studies of landspouts completed over the Colorado region depict identical lightning characteristics during the materialization of a landspout at the surface. As discussed in Chapter 1, landspouts form differently than supercellular tornadoes. They do not require a downdraft to form. Rather, they require vertical vorticity to already be present at the surface. When an updraft of a developing storm moves over a region of enhanced surface vertical vorticity, the vorticity is stretched and advected towards the inflow region, creating enhanced vertical vorticity throughout the atmospheric column (Wakimoto and Wilson 1989). This enhanced vorticity can lead to the formation of a landspout.

In both the DIA and Fort Lupton landspout, the flash rate is increasing when the landspout touches down. The increasing flash rate is consistent with the idea that landspouts form as updrafts move over a region of vertical vorticity and intensify. The graupel and hail volume are both increasing during the onset of the landspout, further supporting the idea that the updraft is intensifying. Graupel volumes are significantly larger than the hail volume for both of these cases. These storms also produce a significantly smaller number of flashes compared to the Colorado supercells described above. This is likely a result of these storms being weaker and shorter lived. The VHF sources remain confined to a small area in the storm due to presumably weaker updrafts being unable to loft hydrometeors to the same altitudes as supercellular updrafts. The smaller flash rates also lead to less production of VHF sources, resulting in defined regions of VHF source density.

The Fort Lupton storm merged with other convection before its dissipation could be observed. In the DIA landspout storm, the landspout lifts just prior to the peak in flash rates. The flash rates then decrease steadily as graupel and hail volumes decrease and rain rates remain large. Landspouts are sustained by updrafts, so weakening of the updraft due to precipitation loading likely led to the landspout's demise. Forecasters may find utility in monitoring storms for increasing flash rates when they are aware of vertical vorticity already being present at the surface, such as by the presence of a DVCZ. Monitoring storm intensification in such situations through monitoring flash rates could serve useful in determining which storms could potentially produce landspouts that day.

CHAPTER 5: SUMMARY

Overall, this study has demonstrated that lightning data can reveal information about the dynamics within a tornadic storm. Updraft intensification is accompanied by increases in total lightning flash rate, and sometimes an increase in the VHF source heights. The organization of a cluster of thunderstorms into a supercell is also characterized by increasing total lightning flash rates and increasing VHF source heights. Tornadogenesis in supercells is typically characterized by decreasing flash rates and descending VHF source densities. This descent likely occurs as a result of updraft weakening and downdraft formation. This idea is supported by indications of decreasing graupel volume and increasing rain rates during the majority of the tornado touchdowns analyzed. Downdrafts are necessary for supercellular tornado formation because they advect vertical vorticity to the inflow region of the storm, where the vorticity can be stretched to create enhanced vertical vorticity at the surface. It is logical that the formation of these downdrafts would be evident in lightning characteristics since lightning flash rates are heavily dependent on the strength of an updraft.

Landspout storms do not require the formation of a downdraft, rather they are characterized by updrafts intensifying over regions of already present vertical vorticity. Landspouts analyzed in this study are characterized by increasing flash rates. The tornadoes lift before flash rates begin to decrease. Understanding how flash rates and other lightning characteristics relate to the formation of tornadoes is important for nowcasting purposes. Knowing that landspouts form during increases in flash rate could help forecasters focus their energy on individual storms showing signs of intensification in regions where vertical vorticity is already present. Forecasters can also find use in knowing that lightning information can

sometimes point towards the formation of a downdraft, supporting supercellular tornado formation. Increasing total lightning flash rates with increasing updraft strength is also a very useful relationship as intensifying updrafts can lead to any type of severe weather.

More case studies need to be completed to develop a more robust understanding of how lightning can be used to diagnose dynamic evolution within potentially tornadic storms, but it is promising to see trends within such a small number of cases. With the imminent release of GLM data to atmospheric scientists, larger scale studies can be completed to assist in this effort as lightning data will be available for the entire western hemisphere.

CHAPTER 6: THE GEOSTATIONARY LIGHTING MAPPER

6.1 INTRODUCTION

The Geostationary Lightning Mapper (GLM) is the first instrument to detect lightning from the geostationary orbit. GLM was created to assist forecasters in nowcasting severe weather situations and in detecting regions where lightning is located, with an emphasis on total lightning detection and more detailed description of lightning outside of surface networks. Eventually, GLM will also provide researchers with a long term climatological database for lightning information (Goodman et al. 2013). Onboard GOES-16, launched in November of 2016, the GLM is currently in its calibration/validation stages in preparation for operational commission. The calibration/validation process has involved a field campaign whose goal was to target storms occurring in regions that have VHF-based Lightning Mapping Arrays (LMAs). The campaign aimed to assess the GLM flash detection efficiency in various types of storms by comparing GLM flash detections to flash detections made by other instruments, including LMAs.

As part of these calibration/validation activities, we participated in the 8 May 2017 flight campaign mission that targeted the Colorado LMA on a fortuitously severe weather day. During this mission, the NASA ER-2 flew over many strong hail producing storms, collecting data with the Fly's Eye GLM Simulator (FEGS) and numerous other sensors from 22:00 to 01:00 UTC. This instrument detects lightning in the same way that GLM does, but at a smaller capacity since it is only able to see regions overflown by the NASA-ER2. The CSU-CHILL radar was operated on this day to provide high quality polarimetric and Doppler observations of convection. Sector scans were performed over storms targeted by the ER-2. PPI sector scans

topped the storms such that the anvil region could also be observed and eventually compared to information gained from the Advanced Baseline Imager (ABI; Schmit et al. 2005) also on board of GOES-16. High resolution RHI scans were taken to display the complex structure and strength of each targeted storm.

Data collected by FEGS onboard the NASA ER-2 and data collected from CSU-CHILL provide ample data sources to scientists analyzing the initial results of the GLM instrument. LMA data, NLDN data, as well as Earth Networks Total Lightning Network (ENTLN) data were recorded and are available for analysis as well. To understand how well GLM performs in various types of intense storms, this study takes a “deep dive” look at three individual storms during field campaign operations on 8 May 2017. By analyzing GLM data and comparing it to other forms of lightning information on a storm cell basis, changes in GLM flash detection efficiency can be directly related to changes in other parameters within a storm.

6.2 DATA

6.2.1 GEOSTATIONARY LIGHTNING MAPPER

The GLM detects optical emissions produced by lightning that emerge at the top and/or sides of clouds (depending on satellite viewing geometry). Characterized by a 1372 x 1300 pixel Charge Coupled Device (CCD) focal array, the spatial resolution of the GLM is 8 km at nadir expanding to 14 km at the edge of the field of view (Goodman et al. 2013). Product latency is less than 20 seconds. Various filtering techniques are utilized to filter out unwanted light so that lightning can be seen even during broad daylight when cloud tops may appear brighter than the emissions produced by lightning. Four unique filters are used on board the space-craft before ground processing further refines the data (Goodman et al. 2012).

The first spatial filter matches the instantaneous field of view (IFOV) to the typical size of lightning illumination over a cloud. This filtering helps to improve the signal to noise ratio by preventing pixels from being under filled with light. A narrow band interference filter centered at 777.4 nm is used to enhance the optical emission produced by lightning (Goodman et al. 2012). The GLM also uses time integration to filter out unwanted data. The GLM measures how long charge, or emitted light, accumulates with time on the CCD focal array between readouts of the data. Lightning emissions are quick and transient while cloud illumination from sunlight is typically steady at the short (2 ms) sampling intervals. By integrating over a 2 ms time period, the signal to noise ratio is improved by filtering out the steady-state signal of cloud brightness (Goodman et al. 2012). The final filtering technique creates an estimated background signal produced from each previous time frame and subtracts the background signal from the next time frame so that gradual changes in the background do not affect the output produced by the GLM (Goodman et al. 2012).

Filtered data are output into “events”. Events are made up of individual lightning pulses that illuminate an individual pixel on the focal plane (Goodman et al. 2013). While data filtering helps to reduce false events, some false alarms remain. More than one lightning pulse can also occur within the same pixel and time integration frame, also leading to error. The Lightning Cluster-Filter Algorithm (LCFA) described in the Algorithm Theoretical Basic Document (ATBD) clusters events together into “groups” and groups into “flashes” so the information is conceptually useful to forecasters and researchers (Goodman et al. 2012). Groups are made up of illuminated pixels that are adjacent to one another and that occur within the same time integration frame (Goodman et al. 2013). The groups are then clustered together to create flashes. A flash is comprised of a cluster of groups that occur within 330 ms of one another and

are separated by no more than 16.5 km spatially (Goodman et al. 2013). A flash can be made up of many events and groups, or it can be made up of one event and group. An example of this grouping is shown in Figure 6.1. The ATBD also mentions that there is potential for multiple flashes to be labeled as one flash due to the generous temporal and spatial thresholds used to define a flash, compared to the finer space/time scales upon which lightning pulses actually occur.

6.2.2 LIGHTNING MAPPING ARRAY

An LMA detects electromagnetic radiation emitted by lightning at frequencies between 60-66 MHz (Rison et al. 1999, Krehbiel et al. 2000). LMAs use GPS time of arrival techniques to geolocate burst radiation from lightning discharges. These sources of radiation can be grouped into flashes utilizing a flash clustering algorithm described in Fuchs et al. (2015) and summarized in Chapter 2. The flash clustering algorithm uses spatial and temporal thresholds based on physical expectations of lightning size and duration, to cluster sources into flashes. Typically for the Colorado LMA, thresholds are set such that sources within a specific flash cannot be separated by more than 3 km and be no more than 150 ms apart in time. For this study, these values are expanded to more generous terms to account for the generous thresholds set for GLM flash clustering. Accordingly, the temporal and spatial thresholds are set to 330 ms and 16.5 km respectively.

6.2.3 NATIONAL LIGHTNING DETECTION NETWORK

The NLDN detects radiation emitted by lightning in the VLF (3-30 kHz)/LF (30-300 kHz), which is much different than the range of frequencies utilized for the LMA (Cummins et al. 1998; Cummins and Murphy 2009). Lower frequencies do not detect the small flashes that are typically seen by the LMA, including IC flashes. Rather, they are very good at detecting

large CG flashes and longer IC flashes. For this reason, NLDN data were used solely for detecting CG flashes and their corresponding polarity throughout this study. IC and CG flashes are differentiated through analysis of the waveform of the radiation recorded. Sometimes this analysis erroneously classifies IC as CG flashes. To ensure that CG flashes used in this study are truly CG flashes, all flashes classified as CG that recorded peak currents smaller than 15 kA were reclassified as IC flashes, and therefore excluded from analysis. If an IC flash created a peak current larger than 25 kA, then the flash was recorded as CG (recommendations per personal communication with Dr. Timothy Lang).

6.2.4 EARTH NETWORKS TOTAL LIGHTNING NETWORK

Earth Networks Total Lightning Network (ENTLN) detects radiation emitted by lightning at frequencies ranging from 1 Hz to 12 MHz (Liu and Heckman 2011). This network uses time of arrival techniques to locate pulses of lightning and then groups the pulses into flashes. ENTLN is a near global network that detects both IC and CG flashes across many countries. ENTLN sensors are separated by hundreds of kilometers, causing their accuracy in detecting the altitude of a flash to be poor compared to an LMA (Liu and Heckman 2011). For the sake of this study, only the latitude and longitude location and time of an ENTLN flash is used in comparing the flashes observed by different systems.

6.2.5 NATIONAL MOSAIC AND MULTI-SENSOR QPE SYSTEM

The National Mosaic and Multi-Sensor Quantitative Precipitation Estimate System (NMQ) ingests level 2 base data from the NWS WSR-88D's as well as the Canadian weather radar network and grids the reflectivity field to a 3-dimensional Cartesian grid such that radar information for the entire CONUS domain is visible on a single display. This system also uses multiple sensor systems such as radar and rain gauges to produce a quantitative precipitation

estimate that is useful for flash flood forecasting. This study utilizes the 3-dimensional gridded reflectivity field to analyze the storms targeted during the 8 May 2017 field campaign mission. Many of the storms from this day occurred along the edge of the KFTG WSR-88D range. By using NMQ data instead of information from one WSR-88D, gaps in storm coverage are avoided.

6.3 ANALYSIS METHODS

6.3.1 *CLEAR*

The CSU Lightning, Environment, Aerosol, and Radar (CLEAR; Lang and Rutledge 2011) framework was implemented to track the storms targeted on 8 May 2017. Due to these storms passing directly over the cone of silence of the CSU-CHILL radar, NMQ radar data was used instead. As described in Chapter 2, CLEAR uses reflectivity and spatial thresholds to identify convective features and track them over time. For this study, storms were required to have a 35 dBZ contour at least 20 km² in area as well as a 45 dBZ contour of at least 10 km² in area to be identified as a convective entity.

6.3.2 *FLASH ATTRIBUTION*

The flash clustering algorithm mentioned above, fully explained in Fuchs et al. (2015), also attributes individual flashes to storm cells defined by the CLEAR framework. In this study, LMA, ENTLN, and NLDN flashes are attributed to each identified storm cell. Flashes that occur within the storm cell 35 dBZ contour or within 10 km of the storm cell are attributed to the storm. GLM flashes, sources, and events are also attributed to each storm. If a GLM flash occurs within the storm cell or within 15 km of the storm cell, the flash is attributed. The GLM attribution parameter is slightly more generous than those defined for the other lighting networks due to known errors in the non-operational GLM flash locations. The generous attribution

parameter may cause a few flashes from other storms to be attributed to the storms of interest, but analysis periods were chosen for time periods in which the storms remained relatively isolated for this reason.

6.3.3 *LIGHTNING ANALYSIS*

The main objective of this study is to compare the flash rates produced by LMA and ENTLN networks to those produced by the GLM to determine how well the GLM is performing in its preliminary stages. Flash rates are determined by binning cell attributed flashes of a storm of interest into one minute intervals and counting the number of flashes in each bin. This methodology is applied in determining the minute by minute GLM event and group counts, as well as the CG flash polarity counts. To determine how well GLM detects lightning compared to high resolution information gained by LMAs, the GLM/LMA ratio is determined by dividing each GLM flash rate count by the LMA flash rate that corresponds to the same time interval. By doing so, a GLM detection efficiency is created with LMA flashes (the latter considered as truth).

In an effort to understand why GLM detection efficiencies fluctuate throughout a storm's lifetime, the flash areas and flash altitudes recorded in the LMA data are further analyzed. Because LMAs detect individual sources of radiation emitted by lightning, LMA data reveals flash information in 3 dimensions. The flash area is determined by projecting the sources of a flash onto a 2D plane and then creating a convex hull around the outer points of the flash (Bruning and MacGorman 2013). The area within the convex hull is defined as the flash area. A visualization of this idea is found in Figure 6.2. The flash altitude is taken as the altitude of the flash centroid. The median flash altitude and flash area for each radar volume time span is recorded and displayed in a time series to depict how the flash areas and altitudes vary with time.

The median value was chosen instead of the mean so that outlying points did not skew the results.

6.4 RESULTS AND DISCUSSION

6.4.1 FORT MORGAN NORMAL POLARITY STORM

Figure 6.3 displays the lightning characteristics for a one-hour period of a storm observed by the NASA ER-2 on 8 May 2017. This storm was long lived, forming before the field campaign mission started and dissipating after the NASA ER-2 began its return to base. The analysis period of this storm was chosen because during this period the storm remained isolated from other convection. Choosing time periods in which convection is isolated ensures flashes attributed to a storm are not influenced by other lightning producing convection. Figure 6.3a indicates that the LMA produced flash rates significantly larger than any other lightning detecting system (Earth-based networks and GLM). ENTLN derived flash rates were over a factor of two less than LMA. GLM flash rates were lower yet. The LMA displayed a peak in flash rate near $210 \text{ flashes min}^{-1}$ at about 23:12 UTC. Both ENTLN and the GLM show a rise and fall in flash rate similar to the LMA, but their peaks occur at significantly different times and the peaks are substantially less than LMA. ENTLN flash rates peak around $85 \text{ flashes min}^{-1}$ at 23:07 UTC and the GLM flash rates peak at about $55 \text{ flashes min}^{-1}$ at 23:16 UTC.

The GLM/LMA flash rate ratio is also displayed in Figure 6.3a to emphasize the time periods in which the GLM detection efficiency with respect to LMA flashes was largest. The ratio averages around 0.2. The highest detection efficiency occurs at 23:16 UTC when the GLM flash rate peaks, reaching a detection efficiency of 31%. Figure 6.3b shows the GLM event rates and group rates for this analysis period. The events and groups follow a similar trend to the GLM flash rates, but produce much larger values. It is interesting to note that the GLM events

and groups follow the LMA flash rates more closely than the GLM flash rate itself. At the beginning of the time period and the end of the time period, significantly more events and groups are produced than LMA flashes, but from 22:57 to 23:12 UTC the GLM groups and events follow the LMA flash rate trend and values rather closely.

Figure 6.3c displays the density of LMA sources binned by height throughout the analysis period. There are two distinct altitudes of enhanced source density throughout the analysis period. One maximum occurs between 10 and 12 km in altitude while the other occurs between 4 and 6 km in altitude. This bimodal source structure is a classic signature that the storm maintained a normal polarity charge structure. Electrical breakdown in a region of positive charge produces more VHF radiation compared to positive break down within negative charge, marking the location of positive charge by enhanced VHF sources. Since a normal tripole charge structure is characterized by a negative charge layer sandwiched between an upper and lower positive charge layer, a bimodal source density structure would suggest that a normal tripole charge structure is present during this time period. This idea is confirmed in Figure 6.3d. The storm produced predominantly negative CG's throughout the entire analysis period.

Analysis of Figure 6.4a and Figure 6.4b show that the GLM/LMA ratio is increasing as the median flash altitude and median flash area are both increasing at approximately 23:14 UTC. It appears that a combination of larger flash areas and higher flash heights may lead to better detection of lightning by GLM. Larger flashes have the opportunity to illuminate more GLM pixels, making these flashes likely easier for GLM to detect. In theory, flashes that occur at higher altitudes would be subject to less scattering and attenuation of optical emissions since light produced by lightning must penetrate surrounding ice to reach the cloud tops. It is logical that detection efficiencies should increase as the flash altitude increases, resulting from a smaller

amount of ice being present between a lightning flash and the top of the cloud. Analysis of the ice water path between flashes and the cloud top would be a useful parameter to explore this idea and should be pursued in future analysis.

Figure 6.4c shows that this storm consistently maintained reflectivity values of 55 dBZ. The maximum echo heights do not appear to have a strong correlation with the GLM/LMA flash ratio, but it is interesting to note that the slight rise in echo heights from 23:03 to 23:06 UTC corresponds to a period of increasing flash rates in each of the lightning detecting networks. Rising echo heights typically correspond to a strengthening updraft leading to the formation of more ice particles. This strengthening updraft likely caused the flash rates to increase and the flash size to decrease (Bruning and MacGorman 2013). During this period, the event and group rates closely follow the LMA flash rates and the LMA flash areas are small compared to the rest of the analysis period. Before and after this time period, the event and group rates are significantly larger than the LMA flash rates and the LMA displays median flash areas that are larger than those found from 22:57 to 23:12 UTC. This trend indicates that GLM events and groups are clustered together into spatially large flashes as the LMA-inferred flash area gets larger, as would be expected if the GLM clustering algorithm is performing well. When flashes are small they likely only take up one pixel, causing the events and group rates to be similar to one another and similar to the LMA flash rates.

6.4.2 GREELEY ANOMALOUS POLARITY CASE

A hail producing storm formed prior to the NASA ER-2 flight and persisted throughout the mission time period. This hail producing storm passed directly over the CSU-CHILL radar and produced a great deal of hail and rain in the process. The analysis period begins at the beginning of the ER-2 mission (22:00 UTC) and continues until about 23:00 UTC. The analysis

terminates at this point because this Greeley storm began to merge with other convection. Figure 6.5a displays the flash rates determined by each lightning detecting network or instrument for this storm. The LMA shows clearly defined increases and decreases in flash rate with flash rates reaching nearly 300 flashes min^{-1} . ENTLN displays much smaller flash rates, with values reaching only 65 flashes min^{-1} . The ENTLN flash rate trends are also very different from those displayed by the LMA. ENTLN flash rates gradually decrease for the first half of the analysis period and then remain relatively constant. GLM shows flash rates only reaching 30 flash min^{-1} with no definitive trend, unlike LMA. Because GLM does not show any definitive trends in flash rate, the GLM/LMA ratio maximizes when the LMA flash rates are smallest. The detection efficiency appears to be much smaller for this case, with GLM/LMA ratios averaging around 0.1 and never reaching higher than 0.2.

Figure 6.5b displays the event rates and group rates produced by GLM for this case. The events and groups are of the same magnitude or smaller than the LMA flash rates throughout the period. This analysis differs greatly from observations in the Fort Morgan normal polarity storm. The Fort Morgan storm displayed GLM group rates and event rates becoming significantly larger as flash sizes increased. In this case, the event and group rates appear noisier and show no definitive upward or downward trend, similar to behavior of the GLM flash rates. The source density plot in Figure 6.5c shows that this storm maintained a clearly anomalous charge structure. The main positive charge region, depicted by the highest source density in the column, remains between 4 and 6 km with no upper layer of positive charge being displayed. The anomalous structure is confirmed by the numerous +CG's produced in this storm shown in Figure 6.5d.

Figure 6.6a displays the median flash altitude alongside the GLM/LMA ratio. The median flash altitude does not fluctuate very much compared to the time series produced for the Fort Morgan normal polarity case. The median flash altitude maximizes near 8 km and reaches a minimum of about 6 km. The GLM/LMA ratio actually shows a slight decrease as the median flash height increases. This is likely a result of flashes remaining small in area during this period (Figure 6.6b) and due to the fact that the flash median height still does not reach the altitudes found in the Fort Morgan case. There are two peaks in the median flash area that directly corresponds to the two peaks in the GLM detection efficiency. These time periods also correspond to local minimums in the LMA flash rate. Overall, this case shows that the flash area may play a role in the detection efficiency of GLM in anomalous storms. Because the flashes occur so low in the cloud and the flash altitude does not increase very significantly, it is logical that larger flashes would have a higher probability of being detected.

The maximum echo heights shown in Figure 6.6c do not show a strong relationship with the GLM detection efficiency and do not show any correlation between the ENTLN and GLM flash rates. The LMA flash rate shows the strongest relationship with this plot. At 22:18 UTC there is an increase in the LMA flash rate that corresponds with increases in the 60 dBZ echo height. There is also a gradual increase in the LMA flash rate from 22:28 through the end of the analysis period that is accompanied by an increase in the 60 dBZ echo height between 22:33 and 22:43 UTC. One thing to note is that this storm is much more intense than the Fort Morgan storm. There was very little if any 60 dBZ elevated area in the Fort Morgan storm, in strong contrast to the Greeley case.

6.4.3 DENVER HAIL STORM

Before the ER-2 aircraft arrival to the Colorado area, a hailstorm traversed Denver producing baseball-sized hail. This storm combined with the other severe storms on this day, produced over a billion dollars in damage (Storm Data- available at <https://www.ncdc.noaa.gov/stormevents/>). The Denver hail storm is analyzed from 22 UTC until its dissipation. Hail reports ceased for this storm just before the analysis period begins. Figure 6.7a displays the flash rates produced by each lightning detecting network or instrument for this case. From the beginning of the analysis period until about 22:24 UTC, the LMA flash rates increase slightly, reaching flash rates of 175 flashes min^{-1} . After 22:24 UTC, the flash rates decrease significantly and then level out before decreasing gradually throughout the rest of the analysis period. Visual analysis of the NMQ radar data shows the Denver hail storm splitting into two cells beginning at 22:20 UTC. The cells continue to split and the south cell weakens and dissipates quickly. Both cells are included in the lightning analysis. The rapid decrease in flash rate is likely a result of these cells splitting and the southern cell weakening.

ENTLN does not show a sharp decrease in flash rate. Rather, the flash rates gradually decrease throughout the analysis period. ENTLN flash rates are much smaller than those produced by the LMA, maximizing near 60 flashes min^{-1} . The GLM flash rates maximize near 40 flashes min^{-1} and show a more pronounced decrease in flash rate compared to ENTLN. The GLM flash rates decrease to zero by 22:36 UTC, and then increase slightly while remaining smaller than 10 flashes min^{-1} thereafter. The GLM/LMA ratio is largest at the beginning of the time period and decreases consistently until 22:36 when the ratio becomes zero since GLM does not record any flashes at this point in time. The ratio then increases to a maximum value of 0.2 and then decreases gradually once again. Figure 6.7b displays the event and group rates

produced by GLM. The groups and events follow the trends identified by the LMA more closely than the GLM flash rate does. The GLM group and event rates begin large, reaching values larger than those portrayed by the LMA flash rate. The events and groups gradually decrease until the storm begins to split, and then they rapidly decrease to near zero. The values then increase once more but remain small, often smaller than the LMA flash rates.

Figure 6.7c displays the source densities for this case. The beginning of the analysis period shows a bimodal structure, characterizing a normal polarity, tripolar charge structure. This is confirmed by the fact that the storm only produces –CG’s flashes during the beginning of the analysis period. When the storm begins to split and the flash rates decrease rapidly, the source densities show a very rapid change from displaying a bimodal density structure to only maintaining a region of large source densities at lower altitude. It is not clear how this occurred without further analysis, but the rapid change appears to be related to the splitting of the storm. Figure 6.8a depicts this rapid change in the median flash heights. The median begins around 8 km and then falls to 6 km by 22:32 UTC. The decrease in GLM/LMA ratio at the beginning of the analysis period closely follows this decrease in median flash altitude. The GLM/LMA ratio increases once again after the median flash altitude levels out, but at this time Figure 6.8b shows that the median flash area has increased. This increase in flash area corresponds to the time period in which the source densities remain low in the cloud.

This storm is rather unique due to the abrupt change in source density and flash rate. The storm switches from a clearly normal polarity storm to one that maintains more of an anomalous charge structure. The GLM detection efficiency with respect to the LMA reflected this rapid change. The largest detection efficiency occurs when the storm appears to maintain a normal polarity structure. The detection efficiency gradually decreases as the storm begins to split and

then never regains the same detection efficiency once the storm displays an anomalous structure. The storm apparently weakens based on a decrease in the reflectivity echo-top heights during the split (Figure 6.8c). The normal polarity portion of this storm was characterized by flashes occurring high in the cloud, likely resulting in more photons exiting the top of the cloud as they have less distance to reach cloud top, leading to the enhanced detection efficiency. As flashes transition to lower in the cloud, the detection efficiency drops significantly. Towards the end of the analysis period the detection efficiency increases once again, but this time the flash area has increased which may have aided in GLM detection of flashes.

6.5 SUMMARY AND CONCLUSIONS

Three preliminary case study analyses were examined for three unique storms observed during the GLM calibration/validation field campaign mission that took place on 8 May 2017. This “gold mine” day featured severe, hail producing, storms of both anomalous and normal polarity charge structure. The Fort Morgan case as well as the beginning periods of the Denver hail case both displayed normal polarity charge structures. These storms produced the largest GLM/LMA ratios, that is, the largest GLM flash detection efficiencies. This is likely a result of flashes occurring at higher altitudes within a cloud. When flashes occur at higher altitudes, photons emitted by lightning do not have as far of a path to travel before they reach the cloud top where GLM can detect them. Smaller paths likely lead to less scattering of light, and thus brighter flashes that are easier for GLM to detect. The Fort Morgan case also showed that increases in the flash area can lead to enhanced flash detection efficiencies during normal polarity storms.

The Greeley storm, as well as the end of the Denver hail storm, produced anomalous charge structures. The Greeley storm was a pure anomalous case, with the majority of the CG

flashes produced being positive polarity. This storm displayed a much smaller flash detection efficiency, with values averaging around 0.1, compared to the values of 0.2-0.3 produced by the Fort Morgan case and beginning of the Denver hail case. Increases in flash detection efficiency occurred when flash sizes became larger. Because the majority of flashes occur at low altitudes in a cloud in anomalous storms, it follows that larger flash sizes would have a higher likelihood of being detected by GLM. The end of the Denver hail case supports this idea.

Overall, this analysis shows that the GLM detects fewer flashes than both the LMA and ENTLN. Interestingly, ENTLN detects fewer flashes than the LMA as well. This is likely because each of these lightning detecting networks and instruments detect lightning in different ways. LMAs detect radiation emitted by lightning in the 60-66 MHz range. ENTLN detects lightning in the same way, but detect emissions from 1Hz to 12 MHz. Because these networks detect at different frequencies they likely detect parts of flashes differently, or detect some flashes and not others (e.g., ENTLN likely misses short, compact flashes compared to LMA). The LMA detects radiation at a higher frequency, implying this type of network can better detect shorter flashes, like compact IC's. In comparing the CG and total lightning flash rates, the total lightning flash rates are significantly larger than the CG flash rates, showing that these storms were dominated by IC flashes. Small IC flashes typically occur along the periphery of an updraft when the updraft intensifies. This may be why the LMA flash rates show more definitive trends that match the evolution of the maximum echo heights over time.

The GLM detects optical emissions produced by lightning. In an analysis of flash characteristics detected by a similar instrument called the Lightning Imaging Sensor (LIS) carried by the Tropical Rainfall Measuring Mission (TRMM) satellite, Peterson and Liu (2013) found that flashes were typically brighter in stratiform regions of a storms where less ice is

present and flashes are typically larger. They found flashes were typically dimmer in regions where more ice was present to scatter and attenuate the optical light. The findings of this study are consistent with the results of Peterson and Lui (2013). GLM seems to detect flashes better when they occur higher in altitude where they do not need penetrate as much cloud or in situations where flash sizes are larger.

This preliminary analysis shows that GLM is successfully detecting lightning flashes in both anomalous and normal polarity storms. Based on this limited sample, the detection efficiency is rather low for both types of storm, with anomalous polarity storm flashes being the most difficult for GLM to estimate flash rates within. Further analysis is needed to fully understand how the dynamics and physical structure of each of these storms relates to the detection of GLM flashes. By developing a more complete understanding of GLM's strengths and weaknesses, the calibration/validation process can be used to improve the GLM instrument and to prepare forecasters for the new, invaluable data source.

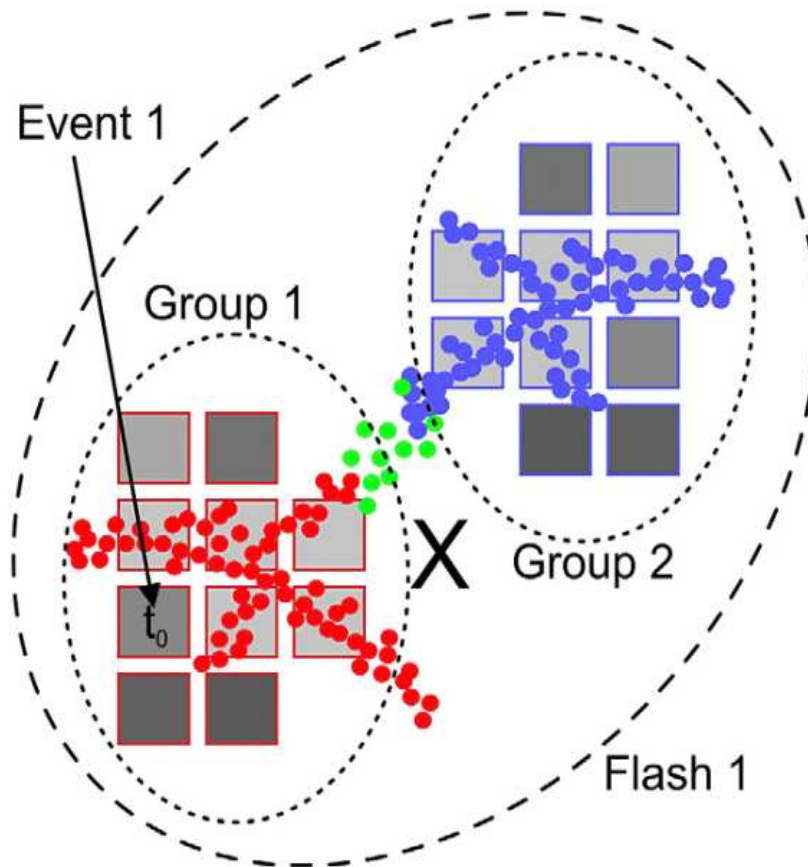


Fig. 5. Illustration of a single GLM flash composed of 2 groups and 20 events relative to a LMA VHF lightning channel. In this example the dots (red, green, blue) are LMA VHF sources and the gray squares are (simulated) GLM data. Time is indicated by color with Red occurring first, Green next, and Blue last. The GLM radiance is indicated by greyscale (darker = greater amplitude). The amplitude weighted flash centroid is indicated by the large X. The time tag for the flash is the time of the first event, labeled t_0 . The two groups (red & blue) are close enough in time/space to be clustered into a single flash (16.5 km & 330 ms). In this example, the green LMA pulses did not create an optical pulse large enough to be detected by the (simulated) GLM (below threshold). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Figure 6.1 Image and caption taken from Goodman et al. 2013 depicting how GLM flashes are created from GLM events and groups

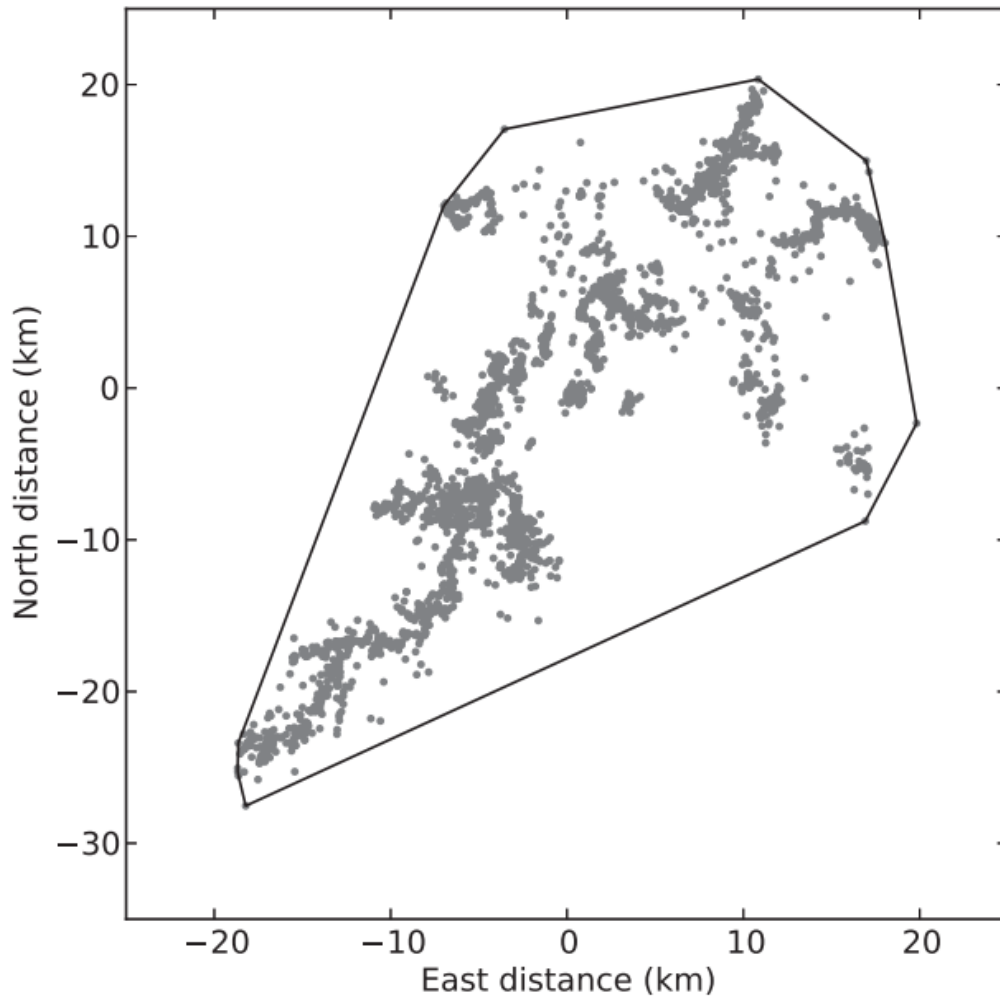


Figure 6.2 Visual depiction of the convex hull method for determining the flash area. Individual dots represents LMA sources. The solid black contour surrounding these dots represents the convex hull drawn around the flash. The area within this contour is recorded as the area of the LMA flash. Image taken from Bruning and MacGorman (2013).

Fort Morgan Normal Polarity Case

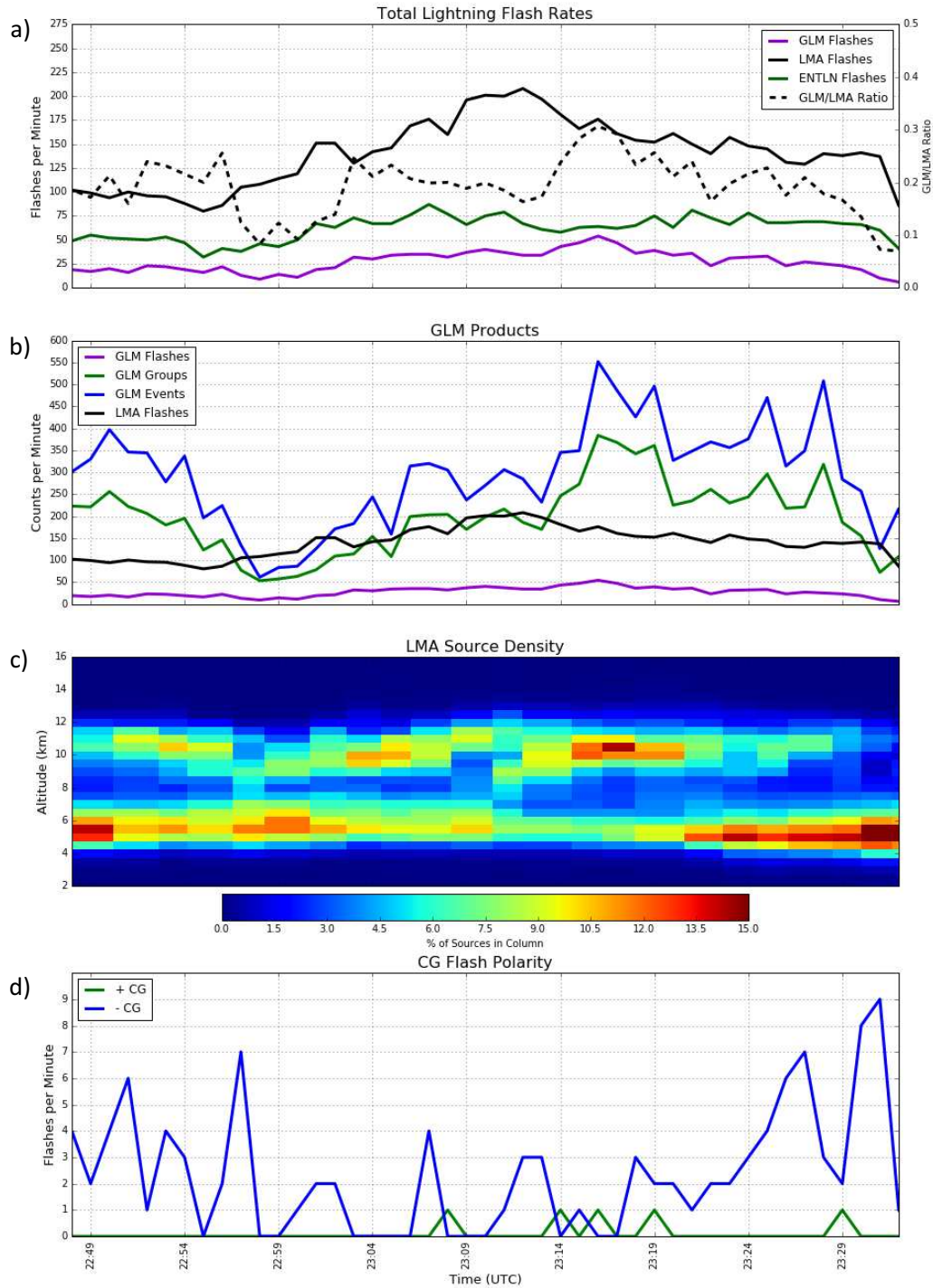


Figure 6.3 displays the lightning characteristics of the Fort Morgan normal polarity case. a) depicts the flash rates produced by the GLM, LMA, and ENTNLN in solid contours while the GLM/LMA ratio is depicted by the dotted line. b) depicts the event rates and group rates produced by GLM as well as the GLM and LMA flash rates for reference. c) LMA source densities are binned by 0.5 km intervals and displayed for each radar volume. Warm colors represent higher source densities. d) NLDN CG flash rates color coded by their polarity.

Fort Morgan Normal Polarity Case

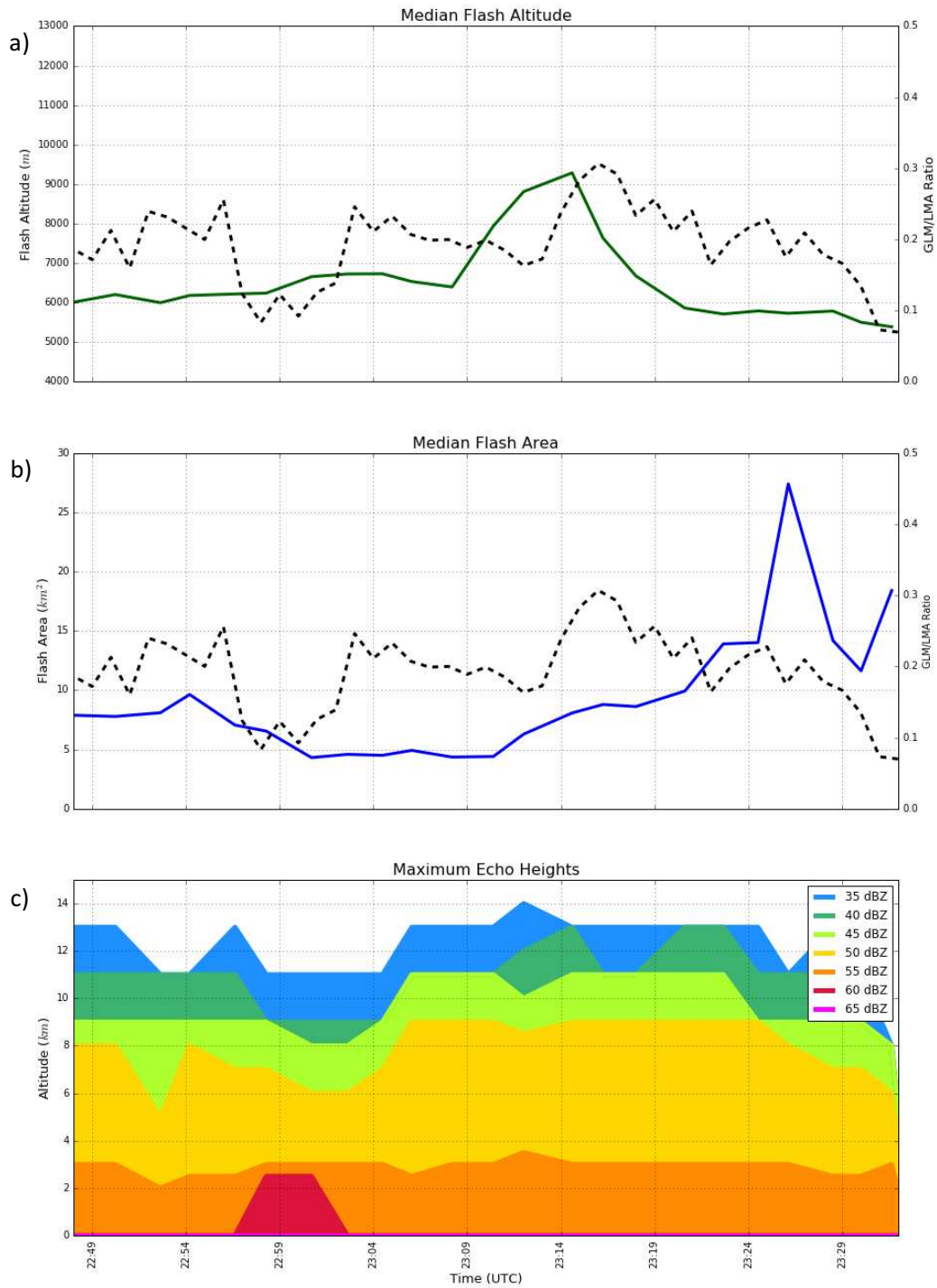


Figure 6.4 displays characteristics of the Fort Morgan normal polarity case. a) displays the median flash heights determined by the LMA over time b) displays the median flash area determined by the LMA over time c) displays the maximum reflectivity heights over time of the 35 dBZ contoured storm.

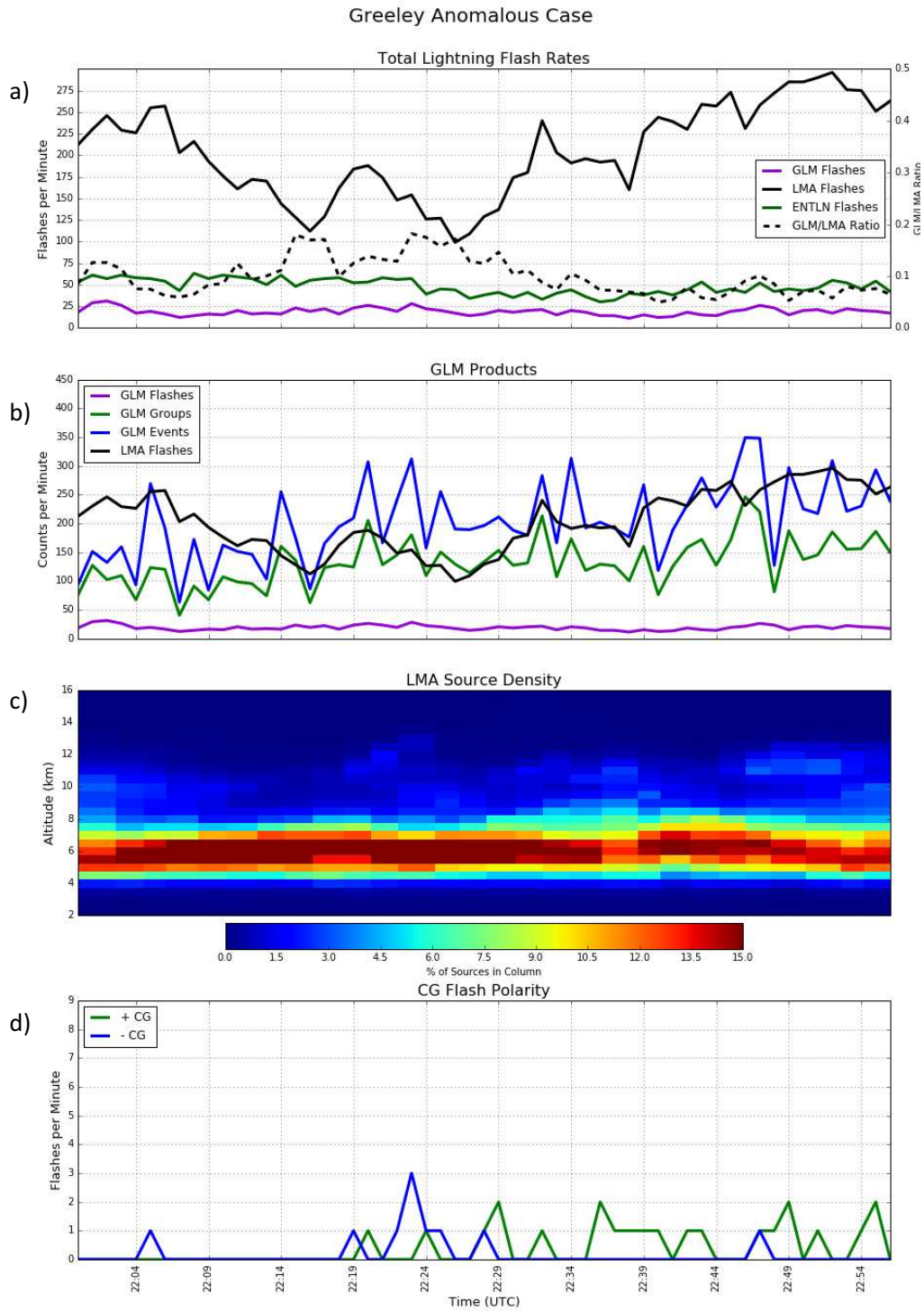


Figure 6.5 Same as 6.3 but for the Greeley anomalous polarity case.

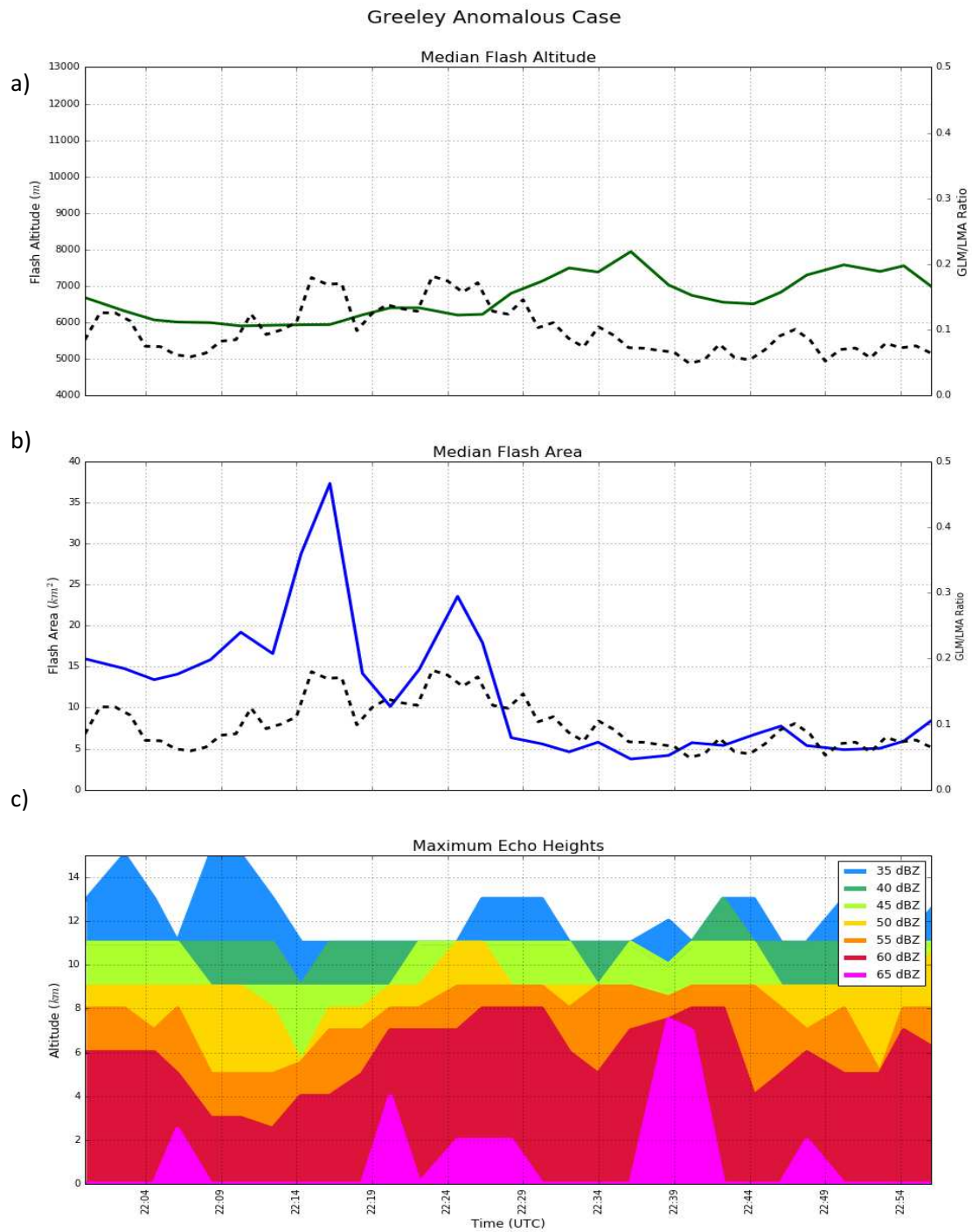


Figure 6.6 Same as 6.4 but for the Greeley anomalous polarity case.

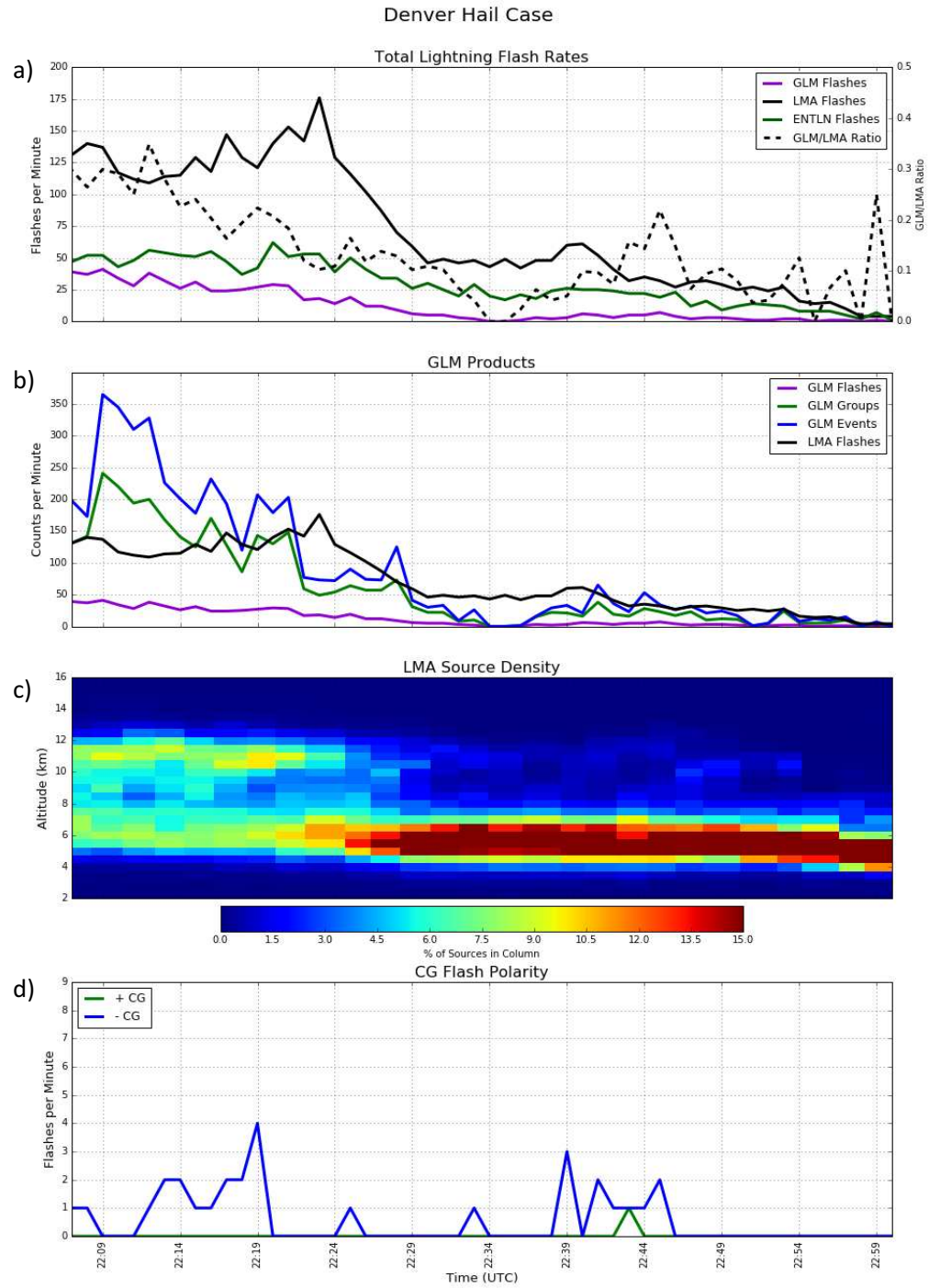


Figure 6.7 Same as 6.3 but for the Denver hail case.

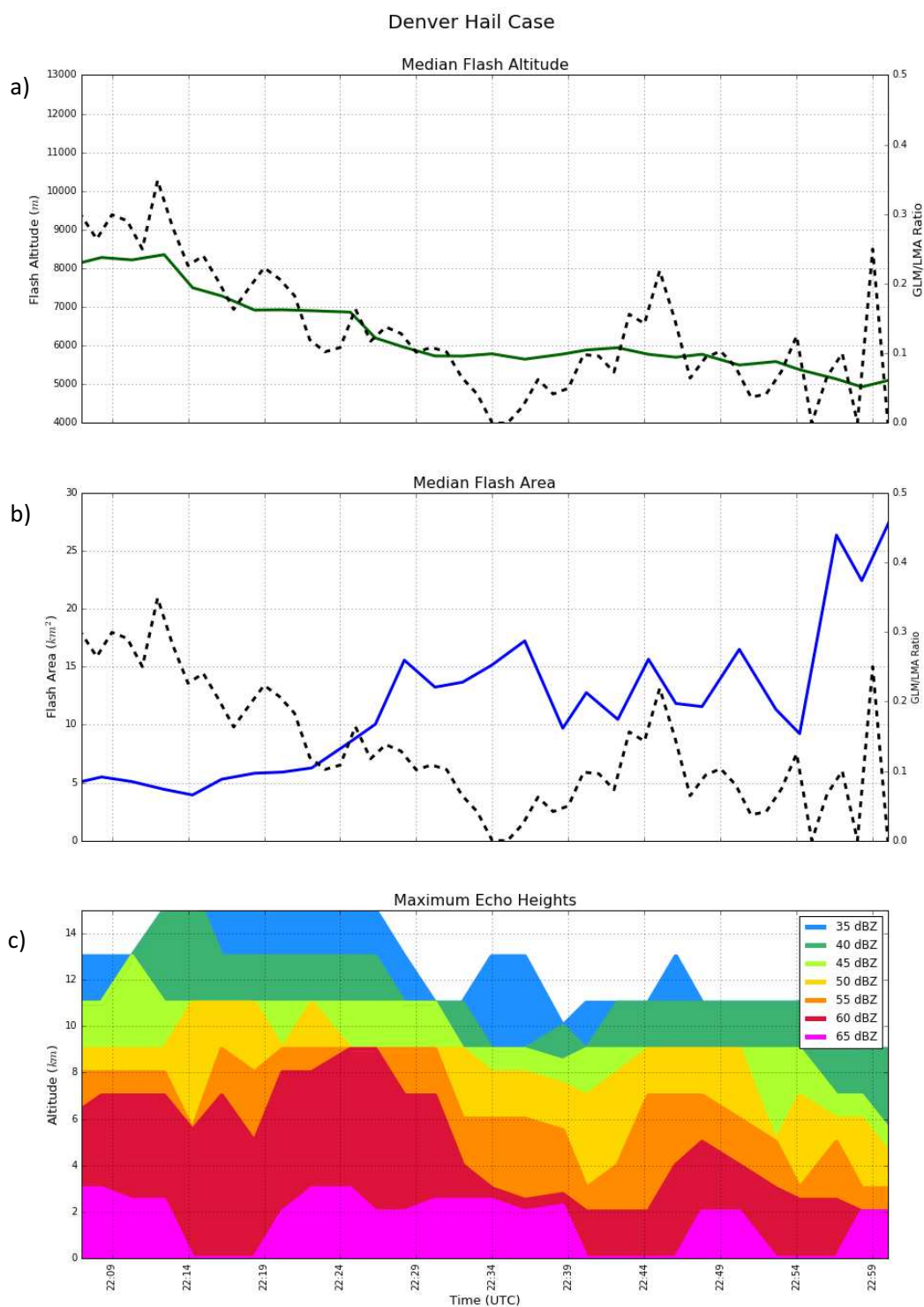


Figure 6.8 Same as 6.4 but for the Denver hail case.

CHAPTER 7: SYNTHESIS

The Geostationary Lightning Mapper (GLM) will provide scientists and forecasters with continuous access to total lightning data over most of the western hemisphere once GLM becomes operational. In preparation for the capabilities of GLM, this study related lightning behavior to the dynamics found within tornadic storms. Lightning flash rates typically increase rapidly, or “jump”, before severe weather is reported at the surface (Williams et al. 1999; Gatlin and Goodman 2010; Schultz et al. 2011, 2015, 2017). This is true because total lightning is closely linked to the updraft characteristics within a storm (Goodman et al. 1988; Dye et al. 1989; Carey and Rutledge 1994; Williams et al. 1999; Lang and Rutledge 2002; Wiens et al. 2005; Tessendorf et al. 2007; Deierling et al. 2008; Schultz et al. 2015, 2017).

Multiple studies have found that the total lightning flash rate jumps prior to tornadogenesis as a result of updraft intensification (Williams et al. 1999; Buechler et al. 2000; Goodman et al. 2005; Steiger et al. 2005; Steiger et al. 2007; Darden et al. 2010; White et al. 2012; Stano et al. 2014). Updrafts intensify prior to tornadogenesis because an updraft is necessary to tilt horizontal vorticity found at the surface into the vertical, where it can generate a mesocyclone (Davies-Jones 1984). A downdraft is then necessary to advect vertical vorticity to the surface so that a tornado can develop (Markowski 2002; Davies-Jones et al. 2001; Davies-Jones 2014). Many studies find that total lightning flash rates typically decrease just prior to the onset of a tornado at the surface, perhaps in relation to this downdraft formation (Williams et al. 1999; Buechler et al. 2000; Steiger et al. 2005, 2007; Darden et al. 2010; White et al. 2012; Stano et al. 2014). Analysis of four tornadic supercells within this study affirmed these findings.

First, two tornadic supercells were analyzed within the domain of the Colorado LMA. One storm produced an EF3 tornado while the other produced three, short-lived EF0 tornadoes. Prior to each tornado report, the total lightning flash rates showed a rapid increase associated with an updraft intensification. Updraft intensifications were inferred to occur when the graupel volume above freezing level in the storm increased rapidly. Intensifying updrafts lead to more supercooled liquid water in the mixed phase region, promoting this graupel growth and thus enhanced flash rates. The presence of KDP and ZDR columns in radar cross sections of these storms confirmed that the updrafts were lofting oblate, liquid drops above the freezing level to aid graupel production. LMA source density heights increased in conjunction with increasing flash rates in two of the four Colorado tornadoes analyzed. These increasing source heights indicated large scale ascent of the positive charge layer.

After rapidly increasing, the flash rates showed a peak and then decreased prior to the start of each tornado report. Decreasing graupel volumes occurred in conjunction with decreasing flash rates, suggesting that graupel fell out of the storm during these time periods. Increasing rain rates at this time are consistent with a water-laden downdraft. These indications of downdraft formation were supported by the realization a tornado. Two of the four cases also exhibited a descent in source density heights, indicating large scale descent of the positive charge layer as hydrometeors fell through the cloud. This finding is similar to Steiger et al. (2005, 2007) where authors find that flash heights descend during tornado reports.

To determine if the ambient environment affects the relationship between lightning behavior and storm dynamics, two tornadic supercells were analyzed within range of the North Alabama LMA. The two Alabama supercells produced smaller flash rates than the Colorado supercells, but the relationship between lightning behavior and the dynamics that led to

tornadogenesis within each storm were similar. Increasing flash rates occurred in conjunction with intensifying updrafts, while flash rates typically decreased prior to or during tornadogenesis. One case produced an EF4 tornado that occurred after an increase, peak, and decrease in both the flash rates and source density heights. The second Alabama case displayed flash rates increasing, peaking, and decreasing around the time of tornadogenesis, but not necessarily prior to it.

The case study analysis of lightning behavior in four tornadic supercells showed good agreement with previous literature. The total lightning flash rates revealed information about what was physically happening in the storm. Increasing flash rates indicated updraft intensification while decreasing flash rates suggested that a downdraft was forming to help advect vertical vorticity to the surface. Lightning flash rates were better correlated to storm dynamics than the source density heights. Though the flash rates differed between the two environmental regimes, the relationship between the internal dynamics of the storms and their lightning behavior did not vary. This is a positive result since forecasters want to use GLM to assist in nowcasting across the entire western hemisphere.

Results from this study and from previous literature indicate that flash rates typically show an increase, peak, and then decrease prior to tornadogenesis. This trend is not necessarily related to the formation of a tornado itself; but instead, it is related to the storm dynamics that allow a supercellular tornado to form. To contrast supercellular tornados and lightning variability, two landspout tornadoes are analyzed within the domain of the Colorado LMA and compared to the supercell cases. Landspouts form when vertical vorticity is already present at the surface. An updraft from a developing storm then stretches the vertical vorticity and pulls it upward through the atmospheric column, allowing for a landspout to form (Wakimoto and

Wilson 1989). Because an updraft is required to maintain a landspout, one would expect flash rates to continue to increase during landspout formation.

Analysis of the landspout cases revealed that the total lightning flash rates generally increased during both tornado reports. Increasing graupel and hail volumes, as well as increasing rain rates occurred in conjunction with the increasing flash rates, indicating that the storms were in a developmental stage during tornadogenesis. Flash rates peaked and then decreased after the landspout lifted in both cases. This decrease in flash rate is likely associated with the updraft weakening, or even the formation of a downdraft (NOTE—in an air mass like storm the downdraft forms and chokes off the updraft). A weakened updraft is detrimental to a landspout since the updraft is responsible for pulling vertical vorticity through the atmospheric column. One case displayed decreasing graupel and hail volumes in conjunction with increasing rain rates after the landspout lifted. These microphysical characteristics occurred in conjunction with decreasing flash rates, suggesting that the flash rates were revealing the formation of a downdraft that led to the termination of the landspout. The source density heights showed no defined trends in either landspout case.

Results of this study indicate that total lightning flash rates can be used to understand dynamical and physical processes in a storm. Few regions across the United States have access to LMA data, so GLM will provide total lightning to many forecasters for the first time. Of course, GLM must properly detect lighting for the instrument to be truly useful. GLM detects optical emissions produced by lightning as seen at cloud top. GLM is currently in its calibration/validation stages and undergoing initial analysis so that problems with the instrument can be pinpointed and corrected before GLM data becomes operational.

Results from limited case study analysis over the Colorado LMA during the GLM calibration/validation field campaign suggest GLM detection efficiencies of 10-30% when LMA data is considered “truth”. These values are significantly smaller than the instrument requirement of 70% detection efficiency. This initial analysis includes three cases studies characterized by a normal polarity storm, anomalous polarity storm, and a storm that displayed both normal and anomalous polarity characteristics. The normal polarity cases displayed the highest detection efficiencies, likely as a result of flashes occurring at higher altitudes compared to anomalous storms. When flashes occur at higher altitudes, they have less ice to penetrate to reach the cloud top, reducing the scattering and attenuation of optical light. Results also indicate the flash size may play a role in detection efficiency, such that larger flashes are more easily detected by GLM.

One case displayed a purely anomalous charge structure characterized by flashes and source density heights occurring at low level in the cloud, as well as the smallest detection efficiency. Radar reflectivity displayed 55 dBZ echo heights reaching 8 km agl for most of the analysis period. There was no correlation between flash height and detection efficiency, likely as a result of flash heights occurring low in the cloud, below large values of radar reflectivity. Perhaps more ice was present between the flashes and cloud top, leading to enhanced attenuation and scattering of optical emission when compared to the normal polarity cases. This will be a subject of a future study. The flash area appeared to play the largest role in flash detection efficiency. The time period in which a separate case displayed anomalous polarity characteristics also showed smaller detection efficiencies, likely as a result of flashes occurring at lower altitudes in the cloud once again.

While the GLM flash rates showed very small values in comparison to the LMA flash rates, GLM events and groups that comprise a GLM flash were closer in magnitude to the LMA flash rates throughout most of the analysis periods. This warrants further investigation of the clustering algorithm used to assemble GLM groups and events into flashes. The lower detection efficiencies found in the anomalous polarity cases likely result from large amounts of ice being present between a flash and cloud top, leading to more attenuation and scattering than would be found in a normal polarity storm where flashes occur at higher altitudes.

Though GLM is still undergoing various improvements, this preliminary analysis suggests GLM may under count lightning in anomalous storms if scattering and attenuation of optical emissions is truly the problem. GLM does not produce the same flash rates as an LMA in the normal polarity cases analyzed, but it does show similar trends in flash rate indicating that GLM will likely be useful for understanding storm dynamics in normal polarity storms. It is important to remember that these conclusions are based on a very small number of case studies over one region. Continued analysis of anomalous storms and normal polarity storms is needed to fully understand where GLM performs best and where GLM needs to be improved. Regardless of the issues presented in this analysis, GLM shows promise in providing forecasters with a useful tool to monitor storm development and lightning location.

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