Examining Conditions Supporting the Development of Anomalous Charge Structures in Supercell Thunderstorms in the Southeastern United States

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Abstract  Hypotheses regarding favorable conditions for anomalous charging have primarily resulted from studies within the Great Plains region of the United States, where the efficiency of warm precipitation processes is thought to be fundamental. Rare observations of anomalous charge structures in the Southeastern region challenge existing conceptual models used to explain anomalous charging. As a rigorous evaluation of conditions that support anomalous charge structures, environmental characteristics and bulk kinematic and microphysical properties of two normal and two anomalous supercell thunderstorms observed in the Southeast were compared. Within the anomalous supercells, greater quantities of precipitation ice were identified at higher altitudes and colder temperatures, suggesting a greater depth of riming growth and increased vertical transport of rimed hydrometeors. Deeper anomalous supercell updrafts were larger and stronger in the upper mixed-phase and glaciated regions. However, normal supercells were characterized by more robust low-level updrafts, resulting in comparable warm cloud residence times that suggested warm precipitation processes were not necessarily less efficient in the anomalous supercells. Indications of enhanced mixed-phase liquid water content in favor of anomalous charging were observed in the anomalous supercells, though contrasts in related environmental parameters were not as large as observed in other comparative studies. Anomalous supercell environments were characterized by increased instability, shallower warm cloud depth, as well as lower relative humidity in the 700–500 mb layer. Evidence of impacts from dry air in anomalous storm structures suggested that water vapor content may have affected particle-scale charge transfer in support of anomalous charge structure development.

1. Introduction

Understanding of the catalysts leading to the development of anomalous charge structures (ACSs) over normal charge structures (NCSs) in thunderstorms remains incomplete. A dominant layer of net positive charge in the lower to middle mixed-phase region (the layer between 0° and −40°C) characterizes ACSs while net negative charge similarly characterizes NCSs (e.g., Bruning et al., 2014; Fuchs et al., 2015; Kuhlman et al., 2006; MacGorman et al., 2005; Rust et al., 2005; Stolzenburg et al., 1998; Wiens et al., 2005; E. R. Williams, 1989, 2005). Results of studies addressing the structural and environmental relationships between dominant thunderstorm charge structure polarities are often varied. Leading hypotheses suggest that ACSs arise from conditions that support increased liquid water content (LWC) in the mixed-phase updraft (Carey et al., 2003; Carey & Buffalo, 2007; Chmielewski et al., 2018; Fuchs et al., 2015, 2018; Lang & Rutledge, 2011; E. R. Williams et al., 2005). However, causative roles of environmental conditions and thunderstorm structures in ACS development are unresolved, as is the relative sensitivity of ACSs to a variety of specific conditions.

A number of different applications are affected by implications of relationships between the environment, storm structure, and charge structure. For instance, it has been observed that in some ACSs, lightning flash occurrences at lower altitudes than in their normal counterparts (Fuchs et al., 2016; Fuchs & Rutledge, 2018). Lightning flash locations influence how and where nitrogen oxide (NOx) and subsequent ozone are produced (e.g., Barth et al., 2015; Chmielewski et al., 2018; Davis et al., 2019) and impact as well the optical detection of lightning in deep convection from space (e.g., Brunner & Bitzer, 2020; Fuchs & Rutledge, 2018;
Fuchs et al., 2016; Murphy & Said, 2020; Rutledge et al., 2020). Further, anomalous storms have been shown to exhibit later development and more limited quantities of cloud-to-ground (CG) compared with in-cloud (IC) lightning flashes (e.g., Carey & Rutledge, 1998; Lang & Rutledge, 2002; MacGorman et al., 2011; Tessendorf, Rutledge, & Wiens, 2007; Wiens et al., 2005), implying variations in risk relationships associated with CG lightning flashes in normal versus anomalous storms (Chmielewski et al., 2018). Proposed relationships between ACSs and thunderstorm microphysics also carry implications for other microphysical processes, including hail production (Fuchs et al., 2018). Therefore, ACS relationships need to be well understood in order to draw appropriate inferences about lightning detection and other storm characteristics.

Most thunderstorms with ACSs have been documented in the Great Plains and Midwest regions of the US (e.g., Bluestein & MacGorman, 1998; Branick & Doswell III, 1992; Carey & Buffalo, 2007; Carey et al., 2003; Chmielewski et al., 2018; Curran & Rust, 1992; Eddy et al., 2021; Fuchs et al., 2015, 2018; Gilmore & Wicker, 2002; Lang & Rutledge, 2002, 2006, 2011; Logan, 2018; Lyons et al., 1998; MacGorman & Nielsen, 1991; MacGorman & Burgess, 1994; Reap & MacGorman, 1989; Seimon, 1993; Smith et al., 2000; E. R. Williams et al., 2005), and therefore, most relationships between ACSs, thunderstorm structures, and environmental conditions have been derived from observations from those regions. Several studies examining contrasts between thunderstorm structures and environments in storms with ACSs and NCs (i.e., “anomalous storms” and “normal storms”) have compared anomalous thunderstorms in the Great Plains with normal thunderstorms in the Eastern Atlantic and Southeastern US (Fuchs & Rutledge, 2018; Fuchs et al., 2016) and the tropics (Lang & Rutledge, 2002). However, contrasts arising from differences between interregional climates are difficult to separate from controls on charge structure polarity. Similar issues in interpretation could arise in the convolution of storm morphology in comparisons. Addressing aspects of these problems, Chmielewski et al. (2018) compared NCs and ACSs observed in ordinary convection in West Texas over a 3-h period, identifying likely complex relationships in controlling environmental parameters as well as consistent differences associated with environmental moisture parameters. However, regional variability between charge structures inferred from total lightning data and environmental conditions in intense deep convection has not been similarly addressed.

The first documented cases of ACSs and associated supercell thunderstorm structures and environments in the Southeastern US were detailed in Stough and Carey (2020). Therein, the Southeastern ACSs were established as consistent with Great Plains-based conceptual models of charge structure, particularly within the context of time-varying mixed-phase and kinematic structures. However, departures in environmental parameters typically associated with ACSs were noted. Comparisons between rare anomalous and more typical normal supercells within the Southeast allow the opportunity for a more critical evaluation of hypotheses concerning the kinematic, microphysical, and environmental support for ACSs in an atypical parameter space. Additionally, restriction of the comparison to supercells within the same region reduces complex effects of interregional climate differences and variability in intensity between storm modes that may have affected previous ACS studies.

1.1. Thunderstorm Electrification

Thunderstorm electrification requires first the transfer of charge between hydrometeors and then the organization of these particles into net charge regions. The greatest magnitudes of particle-scale charge transfer are thought to result from rebounding collisions between non-precipitation-sized (small) and riming (large) ice in the presence of supercooled cloud water droplets, referred to as the ice–ice collisional non-inductive charging (NIC) mechanism (Jayaratne et al., 1983; Reynolds et al., 1957; Saunders et al., 2006; Takahashi, 1978). Current understanding of charge transfer suggests that the particle undergoing more rapid depositional growth is characterized by a greater thickness of what is referred to as a semi-liquid layer, resulting in a greater number of hydroxide (OH) ions over its surface (M. B. Baker & Dash, 1994). More negative ions are thought to transfer from the particle experiencing faster depositional growth during rebounding collisions, leaving the faster-growing particle with positive charge and imparting the slower-growing particle with negative charge (B. Baker et al., 1987; M. B. Baker & Dash, 1989, 1994). Under normal conditions at cooler temperatures and lower LWC, non-riming small ice particles (i.e., ice crystals or aggregates) become positively charged during rebounding collisions as a result of their faster depositional growth over
smaller surface areas while riming ice (i.e., graupel or dry hail) particles become negatively charged as a result of slower or less efficient growth (B. Baker et al., 1987; M. B. Baker & Dash, 1994).

Laboratory studies have shown that the magnitude and polarity of charge transfer are controlled by the velocity of the particle collisions, riming rate associated with effective LWC, ice crystal and cloud water size spectra, and cloud and particle temperatures (Avila & Pereyra, 2000; Emersic & Saunders, 2010; Jayaratne et al., 1983; Saunders & Peck, 1998; Saunders et al., 2006), pointing to the importance of riming efficiency in enhancing the depositional growth rate and positive charging of graupel (e.g., B. Baker et al., 1987; Mitzeva et al., 2005; Saunders et al., 2006). Limited laboratory studies have also suggested that some of the discrepancies in LWC and reversal temperature at which positive charging rather than negative charging was observed may have resulted from unaccounted variations in supersaturation (e.g., Berdeklis & List, 2001; Emersic & Saunders, 2010, 2020; Saunders et al., 2006). Theoretical models also indicate that the saturation ratio influences the growth of small ice relative to graupel such that graupel charges positively at lower values of effective LWC in the presence of reduced supersaturation (Mitzeva et al., 2005; Saunders et al., 2001; Tsenova et al., 2010).

The simplified tripole model of a normal charge structure only accounts for gravitational sedimentation of recently charged hydrometeors into collective regions of like charge. Neglecting effects of three-dimensional flows and factors such as charge deposition by lightning channels, such simplified charge structure explanations are most applicable near a storm's updraft (e.g., Bruning et al., 2010; Stolzenburg et al., 1998). At temperatures warmer than the charge reversal temperature in the lower mixed-phase region of a storm's updraft, sedimenting positively charged graupel composes the smaller lower positive charge region of the tripole. At temperatures cooler than the charge reversal temperature in the middle mixed-phase region, negatively charged graupel combines with advedted negatively charged small ice resulting from rebounding collisions with positively charged graupel below, forming the main negative charge center. Positively charged non-precipitation-sized ice resulting from rebounding collisions with the negatively charged graupel is advected in the updraft and forms the larger upper positive charge region. Whereas NCSs are thought to arise under standard conditions in which LWC is limited at temperatures colder than approximately −10 °C (e.g., Saunders et al., 2006; Takahashi, 1978), ACSs are suggested to develop when the LWC in the mixed-phase region of the storm increases to sufficiently lower the charge reversal temperature, deepening the layer over which a dominant positive charge region is observed and seemingly replacing the dominant negative charge region (e.g., Bruning et al., 2014).

1.2. Suggested Impacts of Warm Precipitation Efficiency on Charge Structure Polarity

Pathways providing for a deeper region of increased LWC in the mixed-phase updraft in support of high-LWC riming and positive charging have been suggested as those which limit depletion of cloud water through warm rain processes, including shallow warm cloud depths (WCDs), robust updrafts, and high concentrations of cloud condensation nuclei (CCN; e.g., Carey & Buffalo, 2007; Fuchs et al., 2015, 2018; Kuhlman et al., 2006; Lang & Rutledge, 2011; Lang et al., 2016; MacGorman et al., 2005; Mansell & Ziegler, 2013; Tessendorf, Wiens, & Rutledge, 2007; Wiens et al., 2005; E. R. Williams, 2001; E. Williams et al., 2002; E. R. Williams & Stanfill, 2002; E. R. Williams et al., 2005). Shallow WCDs are thought to reduce the depth over which warm rain can grow and deplete cloud water (e.g., Carey & Buffalo, 2007; Fuchs & Rutledge, 2018; E. R. Williams et al., 2005). Similarly, fast updrafts not only transport LWC deeper into the mixed-phase region but also reduce the time over which cloud LWC can be depleted by warm precipitation growth (e.g., Carey & Buffalo, 2007; Fuchs et al., 2018). Large updrafts supported by high cloud base heights (CBHs) also limit entrainment, maintaining a more robust updraft capable of transporting increased LWC, and may limit precipitation recirculation in the mixed-phase region that would increase competition for LWC (e.g., Fuchs et al., 2018; MacGorman et al., 2005, 2011; E. R. Williams et al., 2005). High instability metrics, particularly near cloud base and in the lower mixed-phase region, are thought to support these updraft characteristics conducive to ACS development. Other environmental characteristics thought to support ACS development by reducing warm rain efficiency include increased CCN and dry layers near cloud base and in the warm cloud region. Each is suggested to create competition among growing drops for water vapor in a way that reduces the efficiency of warm rain growth and promotes the availability of LWC in the mixed-phase region (e.g., Carey & Buffalo, 2007; Chmielewski et al., 2018; Fuchs et al., 2018; Lang & Rutledge, 2002). The size...
of cloud droplet populations has also been suggested to affect the sign of graupel charging, where it has been observed that smaller droplets favor positive charging of graupel (Avila & Pereyra, 2000). Not all of these environmental conditions have been observed simultaneously in documented anomalous storms nor do they consistently differentiate anomalous and normal storm environments (e.g., Carey & Buffalo, 2007; Chmielewski et al., 2018; Eddy et al., 2021; Fuchs et al., 2018; Lang & Rutledge, 2011). Rather, it is thought that only some combination and degree of these conditions may be required for ACS development (e.g., Carey & Buffalo, 2007; Chmielewski et al., 2018; Eddy et al., 2021; Fuchs et al., 2018; Lang & Rutledge, 2011).

These microphysical and kinematic effects thought to promote high LWC riming in the mixed-phase region have an assortment of environmental roots that have been explored in detail (e.g., Carey & Buffalo, 2007; Fuchs et al., 2015, 2018; Lang & Rutledge, 2011). While warm cloud convective available potential energy (CAPE) may be greater in normal storms, anomalous storms often occur in environments with greater CAPE or normalized CAPE (NCAPE) in the mixed-phase region (Carey & Buffalo, 2007). Environmental measures such as the lifted condensation level (LCL) are used to derive the CBH and the WCD through which warm precipitation growth occurs. Environments in which anomalous storms form have been characterized by higher LCLs and CBHs associated with shallower WCDs and broader updrafts (Carey & Buffalo, 2007; E. R. Williams et al., 2005).

Additionally, the combination of environmental contributions to ACSs may be varied and complex, potentially competing with other environmental conditions. For instance, dry air near cloud base has been noted in anomalous environments (e.g., Carey & Buffalo, 2007; Chmielewski et al., 2018) with various hypothesized impacts. That is, it has been suggested that dry air entrainment may limit condensational growth of cloud droplets, inhibit growth of rain, and effectively increase the availability of small cloud droplets for enhanced ice growth processes, despite also reducing instability and suppressing the updraft (Chmielewski et al., 2018). Owing to these overlaps and complexities, no single parameter may be used as a discriminatory factor in all situations supporting anomalous storms. However, combinations of parameters may be more informative. For instance, warm cloud residence time (WCT, Fuchs et al., 2018) utilizes WCD, warm-cloud updraft velocities, and representative particle fall speeds. This combination may more effectively identify conditions conducive to enhanced mixed-phase LWC that support positive charging of riming hydrometeors.

1.3. Motivation of the Present Study

Kinematic and microphysical relationships with charge structures are compared between a sample of two normal and two anomalous supercells observed in the Southeastern US with specialized lightning and radar networks. While some aspects of the temporal evolution of each storm are discussed for context, the storm-total nature of these properties are emphasized in order to draw effective comparisons with results from studies of Great Plains supercells. The environmental characteristics associated with each supercell are also contrasted in the context of their structures to examine the conditions that supported the unusual development of the ACSs. It is hypothesized that conditions favoring the requisite mixed-phase microphysical state in anomalous storms may manifest as discernible differences from normal storms in kinematic and microphysical structure. If consistent with conceptual models of anomalous charging based on LWC aspects of NIC theory, elements differentiating Southeastern ACSs in environmental data are expected to include those that promote robust updrafts, limit warm precipitation efficiency, and support the microphysical parameter space in favor of positive charging of riming hydrometeors. However, environmental distinctions between normal and anomalous thunderstorms in the Southeast may be more subtle and may not be observed at similar magnitudes as Great Plains anomalous environments. Any departures from relationships gleaned from Great Plains observations may also allude to additional factors that promote anomalous charging, including water vapor considerations in particle-scale charging (e.g., Berdeklis & List, 2001; Mitzeva et al., 2005; Tsanov et al., 2010). Observations of Southeastern anomalous supercells and comparisons against their normal counterparts are expected to enhance understanding of the relative emphases of environmental contributions to and requirements of NIC-based ACS development as well as cloud electrification. Results may also benefit applications to which lightning flash characteristics associated with ACSs are pertinent, including optical detection of lightning, accuracy in NOx modeling, and risk relationships associated with the timing of the onset of CG lightning flashes.
2. Data and Methods

Two normal supercell thunderstorms were observed on February 6, 2008 and April 11, 2008 and two anomalous supercells were observed on April 10, 2009 and April 22, 2017 in North AL and South Central Tennessee (TN). Total lightning and radar observations of the supercells of interest to this study were obtained either during local research operations (February 6, 2008, April 11, 2008, and April 10, 2009 cases) or larger multi-agency field projects (April 22, 2017 case, Padula et al., 2017; Rasmussen, 2015). This study expands upon the work of Stough and Carey (2020) in which the environments and spatial and temporal aspects of the co-evolving electrical, kinematic, and microphysical structures of the anomalous storms were documented in detail. As such, this study incorporates many of the same datasets, processing methods, and analysis techniques. A brief summary of these data and methods is provided with emphasis on the normal supercell data analyzed for the first time herein. Total lightning data, including detection of both CG and IC lightning flashes, were obtained from the North Alabama Lightning Mapping Array (NALMA; Koshak et al., 2004; Rison et al., 1999) while CG lightning flash data were obtained from the US National Lightning Detection Network (NLDN; Cummins & Murphy, 2009). Radar data were obtained from the KHTX S-band operational Weather Surveillance Radar–1988 Doppler (WSR-88D; Crum & Alberty, 1993; Doviak et al., 2000) and the polarimetric C-band Advanced Radar for Meteorological and Operational Research (ARMOR; Mecikalski et al., 2015). Reconstructed soundings from hourly Rapid Update Cycle (RUC, prior to 2012; Benjamin et al., 2004) or Rapid Refresh (RAP, after 2012; Benjamin et al., 2006) model analysis data were similarly used to characterize the pre-convective environment in each case. Aerosol data obtained from the Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2; Gelaro et al., 2017), were used as a loose proxy for environmental CCN. Adapted from Stough and Carey (2020), Figure 1 documents the radar analysis domain, the NALMA network, as well as the tracks along which the four supercells of interest were analyzed. Additional information about the synoptic to meso-α conditions under which each storm developed, model environmental sounding profiles, and a radar overview of each storm are available as Supporting Information S1.

2.1. Storm Identification

Objective storm tracking was utilized to isolate and identify precipitation, lightning, and kinematics associated with each supercell as it propagated. Storm footprints were identified based on a feature with a minimum size of 20 km² at a specified minimum reflectivity at the height of −10°C and tracked through time using the Warning Decision Support System–Integrated Information (Lakshmanan et al., 2007) software. These footprints were expanded into tracking boxes using subjectively selected latitudinal and longitudinal radii to encompass and isolate the full storm area. Tracking information pertaining to each supercell is included in Table 1. The tracking method is completely detailed in Stough et al. (2017).

2.2. Radar Observations

Radar data were obtained from the S-band KHTX WSR-88D located in Hytop, AL, and the C-band ARMOR located in Huntsville, AL (Figure 1). While KHTX had received a dual-polarization upgrade before 2017, dual-polarization data were available only from the ARMOR during the 2008 and 2009 cases. To maintain consistency, dual-polarization data from the ARMOR alone were used for hydrometeor identification (HID) and calculation of microphysical properties in all storms. Differences between quantities resulting from polarimetric products from KHTX and the ARMOR were found to be insignificant and inconsequential to interpreted results from the study (e.g., Mecikalski et al., 2015). Doppler velocity data from both radars were utilized for dual-Doppler vertical wind retrievals in each case.

Prior to analysis, ARMOR data were corrected for attenuation and differential attenuation (Bringi et al., 2001; Mecikalski et al., 2015). Specific differential phase (KDP) was calculated following Bringi et al. (2001) as described in Mecikalski et al. (2015). ARMOR data were subsequently manually quality controlled (Oye et al., 1995; Stough & Carey, 2020) and ARMOR and KHTX Doppler velocity data were also dealiased (Helmus & Collis, 2016; James & Houze, 2001; Oye et al., 1995; Stough & Carey, 2020).
Processed, corrected radar data were gridded to a common Cartesian coordinate system with a Cressman weighting scheme using the Python Atmospheric Radiation Measurement Radar Toolkit (Py-ART, Helmus & Collis, 2016). The common grid size was 250 km by 250 km in the horizontal and extended 20 km above ground level (AGL) in the vertical with the ARMOR location serving as the grid center. Grid spacing was 1 km in the horizontal and vertical dimensions, while the minimum radius of influence of 870 m expanded by a factor of 0.025 and 0.04 m for every 1 m increase in horizontal and vertical distance from the radar, respectively.

Figure 1. Domain over which supercells were analyzed. Advanced radar for meteorological and operational research (ARMOR) and KHTX radar sites (dark blue and dark red diamonds), permanent and supplemental North Alabama lightning mapping array (NALMA) sensors (green circles and orange crosses) with notes on special operating conditions pertaining to 2008 and 2017 cases (black and gold dots), the 30° beam-crossing area of the dual-Doppler domain established between ARMOR and KHTX (purple rings), and storm centroid tracks (red and blue dashed lines) are shown. For scale, note that each dual-Doppler lobe diameter is 140.6 km. Adapted from Figure 1 in Stough and Carey (2020).
Table 1

Normal and Anomalous Supercell Analysis Information

| Case date     | Analysis period [UTC] | Min. tracking reflectivity [dBZ] | W., E., S., N. expansion [km] |
|---------------|-----------------------|----------------------------------|------------------------------|
| February 6, 2008 | 1002–1123             | 20                               | 10, 35, 20, 25               |
| April 11, 2008 | 1844–1956             | 30                               | 10, 20, 5, 15                |
| April 10, 2009 | 1712–1825             | 30                               | 10, 25, 12, 10               |
| April 22, 2017 | 2056–2206             | 20                               | 5, 25, 10, 15                |

Note. Adapted from Table 1 in Stough and Carey (2020).

Vertical motion in each storm was retrieved utilizing the three-dimensional variation (3DVAR) method (Potvin, Shapiro, & Xue, 2012; Shapiro et al., 2009), available from the open-source Pythnic Direct Data Assimilation (PyDDA) package (Jackson et al., 2019). Through 3DVAR, an analyzed three-dimensional wind field is retrieved via minimization of a cost function which incorporates Doppler velocity observations from two or more radars, mass conservation, and smoothness constraints. Details of application of the 3DVAR technique to these storms are elaborated upon in Stough and Carey (2020). The 3DVAR technique offers the benefit of avoiding error propagation associated with integration of mass continuity through the depth of the retrieval and related boundary condition assumptions utilized in traditional methods (Potvin, Betten, et al., 2012).

Storm analyses were conducted while each storm was within the 30° beam-crossing area of the ARMOR-KHTX dual-Doppler domain (Figure 1), the temporal bounds of which are documented in Table 1. Note that gaps in sampling from the ARMOR during the 2017 case resulted in extended periods of missing retrievals near the times of 2140, 2150, and 2155 UTC in the latter portion of the analysis period.

Regions of precipitation ice relevant to charge structure considerations were assessed using a fuzzy-logic HID algorithm implemented in the CSU RadarTools Python package (Dolan & Rutledge, 2009; Dolan et al., 2013; Lang et al., 2016; Mroz et al., 2017). Precipitation ice mass (PIM) within regions of updraft \( \geq 5 \text{ m s}^{-1} \) was identified where hydrometeor types were classified as big drops or melting hail, hail, high density (HD) graupel, and low density (LD) graupel following reflectivity-mass relationships utilized by Deierling et al. (2008) and Heymsfield and Miller (1988). Deierling et al. (2008) includes PIM data from North Alabama and the Plains, representing a variety of storm modes and intensities, that may be useful for familiarization with PIM in the context of lightning research.

2.3. Lightning Observations

2.3.1. Total Lightning Data

Total lightning data were used to characterize the charge structure of active electrification within the mixed-phase updraft of each supercell. The NALMA was comprised of between 10 and 12 active sensors located in North AL and South Central TN between 2008 and 2009 and was supplemented by additional mobile sensors in 2017 during field project operations in the region (Figure 1). Lightning data used in this study were analyzed within 125 km of the network center; a range within which minimum source location accuracy and source and flash detection efficiencies are considered acceptable (Chmielewski & Bruning, 2016; Fuchs et al., 2016; Koshak et al., 2004; Thomas et al., 2004).

2.3.2. Lightning Flash Identification

Very high frequency (VHF) source data from the NALMA were clustered into individual lightning flashes using the Python-based Imatools algorithm (Fuchs et al., 2015, 2016). Prior to flash clustering, VHF source location and timing information were standardized. Spatial and temporal standardization criteria used in each case were 3.0 km and 3.0 s, consistent with physical properties of flash propagation. A maximum flash duration of 3.0 s was also imposed. For the 2008 and 2009 data, a minimum of 6 detecting sensors was required for each source included in a cluster. As the 2017 NALMA consisted of more sensors, a 7-sensor minimum was required in order to reduce the inclusion of non-lightning noise in clustered flashes. Total lightning flash rates were calculated over the analysis period of each supercell by taking the 1 min average flash count in 2 min periods (e.g., Schultz et al., 2009), using the flashes identified within the tracking boxes.

2.3.3. Polarity Classification

Net charge structures were inferred from aggregated classified VHF sources determined using temporal and altitudinal properties of individual lightning flashes. Source polarities were objectively classified using an application of the Density-Based Spatial Clustering of Applications with Noise (DBSCAN, Ester et al., 1996) algorithm that leverages the properties of VHF emission and the model of bi-directional flash propagation.
(e.g., Bruning & MacGorman, 2013; Coleman et al., 2003; MacGorman et al., 1981; Mazur & Ruhnke, 1993; Rison et al., 1999; Thomas et al., 2001; E. R. Williams, 1985). The details of this charge classification method and how it compares with more traditional techniques are described in Stough and Carey (2020). Briefly, DBSCAN was implemented to identify clusters of VHF sources apart from isolated VHF sources in standardized temporal and altitudinal space. Because positive breakdown in negative charge regions is not as noisy in the VHF range, characteristic isolated sources were labeled as negative sources. Clusters of sources that met a set of size criteria were considered to be consistent with noisier negative breakdown in positive charge regions and labeled as positive sources. All of the sources in a lightning flash and the flash itself remained unclassified if too few source clusters met the size criteria to be considered consistent with the model of flash propagation in the VHF.

To most closely match inferred charge to microphysical and kinematic conditions at the time the mixed-phase region was sampled by the radar, classified sources were selected from flashes that occurred within the two-minute period beginning two minutes after the start of each ARMOR volume. Note that gaps in charge analysis in the 2017 case are related to aforementioned missing portions of dual-Doppler retrievals associated with gaps in ARMOR sampling. Sources that occurred during each two-minute period were gridded to match the radar grid structure. Source densities within 5 m s\(^{-1}\) updrafts were then retained to consider the charge associated with most recent NIC in the updraft. Comparisons between the relative vertical frequencies of source densities in the updraft were used to differentiate regions of net positive and net negative charge. Using the vertical frequencies instead of the raw source densities mitigated the effects of stronger emission of negative breakdown in the VHF that results in more numerous positive source detections and more sparse negative source detections. For the purposes of this study, “NCS” or “ACS” will refer to a charge structure in which the dominant charge region in the mid-level updraft has been inferred as negative or positive, respectively.

2.3.4. CG Lightning Flash Data

CG data from the NLDN data include information about the timing, location, number of strokes in, polarity, and peak current of detected lightning flashes. CG lightning flash detection efficiency of ≥90% has been reported within the continental US during the period of data considered in this study (Biagi et al., 2007; Cummins & Murphy, 2009; Murphy et al., 2020). While NLDN data have differentiated IC and CG lightning flash detections since 2006, studies have shown that IC lightning flash detections are occasionally misclassified as CGs with errors most often found in lightning flashes labeled as CGs with positive peak current of <15 kA and negative peak current of >-10 kA (Biagi et al., 2007; Fleenor et al., 2009; Zhu et al., 2016). As a result, NLDN CG data used in this study were filtered to remove lightning flashes with peak currents of between −10 and 15 kA.

Using both NALMA-derived total lightning flash rate and NLDN-derived CG flash rate, the storm total ratios of IC to CG flashes (IC:CG ratio) were calculated. The total percentage of CGs that were positive was also calculated using NLDN data. With the exception of these two quantities, all other lightning properties, including total lightning flash rate, were determined exclusively using NALMA total lightning observations. Hereafter, “lightning” and “lightning flash rate” will be used to refer to “total lightning” and “total lightning flash rate” for simplicity.

2.4. Environmental Data

Model analysis data were utilized to characterize the pre-convective thermodynamic environment associated with each storm in the absence of proximal radiosonde data. Data from the 13-km and 20-km RUC model were used for 2008 and 2009 environmental analysis, respectively, while data from the 13-km RAP model were used for 2017 environmental analysis. Soundings reconstructed from model analysis data were assessed at the locations of each storm approximately an hour prior to the beginning of each storm analysis period. Analysis times, reconstructed sounding location coordinates, and relevant quantities from this analysis are documented in Table 2. It should be noted that environmental temperature data were used as required inputs to the 3DVAR and HID algorithms. For the purposes of all other analyses, temperature profiles...
originated from a virtual temperature-based pseudoadiabatic parcel. Each of these profiles can be thought of as boundaries on the most likely conditions experienced by the storm, where the pseudoadiabatic parcel profile may more closely represent conditions within a moist updraft of a supercell (Davies-Jones, 1974; Davies-Jones & Henderson, 1975). Pseudoadiabatic parcel profile temperatures are used for subsequent discussion unless otherwise specified.

The MERRA-2 reanalysis data set was used to assess the concentration of aerosols present in the environment on each day. MERRA-2 data are available every three hours over a domain with 5° longitude and 1° latitude spacing. Data were restricted to the most likely species to have contributed as CCN, though these data should still be treated as a coarse approximation. Vertical profiles of aerosol mixing ratio data corresponding to dust with particle radii of between 0.1 and 1.8 μm; sea salt with particle radii of between 0.1 and 1.5 μm; and hydrophilic black carbon, hydrophilic organic carbon, and sulphate with particle radii of 0.35 μm were selected from the analysis period prior to the time each storm was first sampled. These data were converted to mass concentration, summed in the vertical, and averaged over 12 data points associated with the northern extent of the sampling domain for the anomalous storms or the southern extent of the sampling domain for the normal storms (Figure 1). The final profiles used are representative of analysis locations bounded within the longitudinal range of −87.5°W and −85.625°W and the latitudinal range between 35.0° N and 36.0° N for the anomalous cases to the north or the latitudinal range of 33.5° N and 34.5° N for the normal cases to the south.

| Table 2 | Environmental Parameters Obtained From a Model Sounding at the Location of Each Normal and Anomalous Supercell an Hour Prior to Its Analysis Period |
|---------|---------------------------------------------------------------------------------|
| Model analysis sounding time | February 6, 2008 | April 11, 2008 | April 10, 2009 | April 22, 2017 |
| Model analysis sounding location | 0900 UTC | 1800 UTC | 1600 UTC | 2000 UTC |
| Surface temperature (°C) | 19.5 | 24.9 | 20.6 | 22.2 |
| Surface dew point temperature (°C) | 17.8 | 18.9 | 16.6 | 18.2 |
| Height of env. 0°C (ML; m) | 3,861 | 4,365 | 3,105 | 3,617 |
| Height of env. −40°C (m) | 9,640 | 9,810 | 8,177 | 9,449 |
| Wet bulb zero height (m) | 3,403 | 3,619 | 2,336 | 3,192 |
| PW in sfc. to 400 hPa layer (cm) | 3.5 | 3.8 | 2.3 | 3.2 |
| Mean mixing ratio in lowest 100 hPa (g kg⁻¹) | 11 | 13 | 11 | 12 |
| Midlevel RH (700–500 hPa layer; %) | 55 | 41 | 21 | 36 |
| Mean RH through full depth (%) | 68 | 54 | 41 | 54 |
| LCL height (m) | 585 | 1,022 | 643 | 900 |
| WCD = ML−LCL (m) | 3,280 | 3,340 | 2,460 | 2,720 |
| CAPE (J kg⁻¹) | 447 | 1,214 | 2,123 | 1,453 |
| NCAPE (m s⁻²) | 0.06 | 0.11 | 0.19 | 0.14 |
| 0–6 km AGL shear (m s⁻¹) | 29 | 25 | 32 | 25 |

Note: Note that the height of the LCL and instability metrics of CAPE and NCAPE were derived from surface-based parcels. Adapted from Table 2 in Stough and Carey (2020).

Abbreviations: AGL, above ground level; CAPE, convective available potential energy; LCL, lifted condensation level; NCAPE, normalized convective available potential energy; RH, relative humidity; WCD, warm cloud depth.
The difference between the relative vertical frequencies of positive and negative sources observed in each storm highlights two primary charge layers near the updraft in each case. In the normal supercells (Figures 2a and 2b), the lower dominant charge layer in the lower to middle mixed-phase region was inferred as negative and a positive charge layer was inferred aloft in the upper mixed-phase and glaciated regions. Contrasting structure was observed in the anomalous supercells (Figures 2c and 2d), in which the lower dominant charge layer in the lower to middle mixed-phase region was inferred as positive, beneath a layer inferred as negative in the colder mixed-phase and glaciated regions. In general, net charge layers within the updraft were better defined in the normal supercells (Figures 2a and 2b) and anomalous April 10, 2009 (Figure 2c) than in the anomalous April 22, 2017 supercell (Figure 2d). Relative differences in positive and negative source vertical frequency were significantly larger in the anomalous April 10, 2009 supercell (Figure 2c) than in the normal April 11, 2008 supercell (Figure 2b).
negative charge layers were more diffuse in the anomalous April 22, 2017 supercell, suggesting charge structure complexity arising from vertical overlap in charge regions and horizontal charge region heterogeneity (Stough & Carey, 2020).

The charge layers in the anomalous April 10, 2009 supercell exhibited more vertical variation over time than was observed in the other storms. Initiation and development through approximately 1739 UTC were captured during the analysis period of the April 10, 2009 storm whereas all other storms were analyzed after they had already reached maturity (Stough & Carey, 2020). As the updraft developed in the April 10, 2009 supercell, the flash rate increased (Schultz et al., 2015, 2017) and the net charge layers associated with charged hydrometeor populations were lofted to higher altitudes within the storm (Stough & Carey, 2020). Charge layers remained relatively distinct in the anomalous supercells as they retained robust updrafts through the end of their analysis periods (discussed in Section 4.1), evidenced by increasing or maintenance of high lightning flash rates (Figures 2c and 2d). Unlike in the anomalous supercells, initial decaying phases were captured in the analysis periods of each of the normal supercells as reflected by steadily declining lightning flash rates (Figures 2a and 2b). As these storms weakened, fewer lightning flashes were observed from which charge structure could be inferred. As a result, the charge polarity in a given layer became more poorly defined after 1037 UTC in the normal February 6, 2008 supercell (Figure 2a) and after 1941 UTC in the normal April 11, 2008 supercell (Figure 2b). After 1107 UTC in the February 6, 2008 case, very few flashes occurred and were mostly located outside of the updraft (not shown), resulting in sparse, strong polarity signals (Figure 2a).

In addition to charge structure differences, the four storms exhibited differences in lightning flash properties (Table 3). Peak total lightning flash rates were greater in the anomalous supercells, consistent with observations in the literature (e.g., Tessendorf et al., 2005). In particular, the maximum lightning flash rate in the April 22, 2017 supercell was two to three times as large as observed in the other three storms. This is consistent with the occurrence of numerous small lightning flashes (not shown) as well as aforementioned charge structure complexity (Bruning & MacGorman, 2013). As Zhang and Cummins (2020) discussed, small lightning flashes may reduce Geostationary Lightning Mapper detection efficiency, potentially contributing to reduced detection efficiency in anomalous storms by space-borne optical lightning detection instrumentation (Murphy & Said, 2020; Rutledge et al., 2020; Thiel et al., 2020).

Storm total CG lightning properties are considered in order to place these storms in the context of historical studies of thunderstorm polarity that almost exclusively used CG data prior to greater availability of total lightning data (e.g., Carey & Buffalo, 2007; Carey et al., 2003; MacGorman & Burgess, 1994; Rust & MacGorman, 2002; E. R. Williams & Stanfill, 2002; E. R. Williams et al., 2005). Not only were the maximum total lightning flash rates higher in the anomalous supercells, the IC:CG ratios were considerably greater at 60.5 and 139.6 compared with 7.0 and 15.2 observed in the normal supercells (Table 3). These values are consistent with elevated IC:CG ratios observed in Great Plains anomalous storms (e.g., Boccippio et al., 2001; Carey & Rutledge, 1998; MacGorman et al., 1989). Further, the percentages of positive CGs observed in the anomalous storms were 20% and 60% compared with 5% and 6% positive CG lightning flashes observed in the normal storms. While the percentage of positive CGs observed in the anomalous April 10, 2009 supercell

### Table 3

|                     | Normal         | Anomalous     |
|---------------------|----------------|---------------|
|                     | February 6, 2008 | April 11, 2008 | April 10, 2009 | April 22, 2017 |
| Peak total lightning flash rate (min⁻¹) | 62             | 48            | 79            | 266            |
| IC:CG ratio          | 7.0            | 15.2          | 60.5          | 139.6          |
| Percent positive CG (%) | 6              | 5             | 20            | 60             |

Note. Total lightning properties, including IC lightning, are reported as observed by the NALMA, whereas CG properties are reported as observed by the NLDN.

Abbreviations: CG, cloud-to-ground; IC, in-cloud; NALMA, North Alabama Lightning Mapping Array; NLDN, National Lightning Detection Network.

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was lower than observed in the anomalous April 22, 2017 supercell, it was still greater than the typical maximum value of 10% positive CGs within a normal storm (e.g., Carey et al., 2003). These observations are generally consistent with Great Plains anomalous storms in which higher fractions of positive CG versus negative CG lightning flashes have been observed (e.g., Carey & Buffalo, 2007; Carey et al., 2003; Lang & Rutledge, 2002; E. R. Williams et al., 1999).

4. Supercell Structural and Environmental Comparisons

The following section discusses comparisons of the supercells’ kinematic and microphysical structures to identify common features in Southeastern anomalous storms that differentiate them from normal storms. Combined with information from environmental analyses, structural observations are used to explore support for the microphysical processes thought to be required for the development of ACSs.

4.1. Updraft Structure

4.1.1. Updraft Size

Changing updraft volumes in each supercell (Figures 3a and 3b) reflected changes in intensity consistent with trends observed in the time series of vertical charge structures (Figure 2). Of the four storms, the anomalous April 10, 2009 supercell exhibited the most change over the course of the storm analysis period as its development and maturation were observed. As the anomalous April 10, 2009 supercell intensified, it reached relative maximum 5 m s⁻¹ and 10 m s⁻¹ updraft volumes that were at least 497 km² and 522 km² greater than those of the other storms (Figures 3a and 3b). The updraft volumes in the normal April 11, 2008 supercell were noticeably smaller than in the other storms over most of its analysis period. Its lightning flash rate decreased after 1915 UTC (Figure 2b), approximately 30 min into the analysis period, and reached a relative minimum below 10 min⁻¹ at 1941 UTC, approximately 55 min into the analysis period. The overall charge structure began to exhibit more variability at 1941 UTC as well (Figure 2b), as the 10 m s⁻¹ updraft volume decreased over time to 3 km³ (Figure 3b). Whereas decreasing kinematic support was evident in the normal April 11, 2008 supercell, the low lightning flash rates and obscured gross charge structure observed in the normal February 6, 2008 supercell (Figure 2a) coincided with a larger 5 m s⁻¹ updraft volume comparable to that observed in the anomalous April 22, 2017 supercell (Figure 3a). However, its 10 m s⁻¹ updraft volume generally decreased during the second half of the analysis period, reaching a relative minimum of 249 km³ that was approximately 153 km³ smaller than the minimum observed in the anomalous April 22, 2017 supercell (Figure 3b). Even with some slight increase at the end of the analysis period, the 10 m s⁻¹ updraft volume in the normal February 6, 2008 supercell remained smaller than that of the anomalous April 22, 2017 supercell.

4.1.2. Updraft Speed

The altitudinal frequencies of vertical velocity in each storm in Figures 3c–3f provide additional information about the range of velocity structures observed between the four supercells. The greatest updraft velocities were observed in the anomalous April 10, 2009 supercell near 50 m s⁻¹ (Figure 3e), while the weakest maximum updraft velocities were observed in the normal April 11, 2008 supercell with maxima near 25 m s⁻¹ (Figure 3d). These extrema are consistent with the trends in total updraft size observed in each storm. Just as updraft sizes were generally most similar in the normal February 6, 2008 and anomalous April 22, 2017 supercells (Figures 3a and 3b), velocities in each exhibited similar extrema with maximum values near 30 m s⁻¹ (Figures 3c and 3f).

Notched box plots shown in Figure 4 were used to examine the top 25% of the maximum vertical motion observed in each layer (layer-maximum updraft speeds) at each analysis time and the altitudes at which these updraft speeds were observed. Notched box plots illustrate the confidence around the median of a distribution and can be used to identify whether distributions are similar based on the comparative notch location and shape within each box. Considering first the relative distributions of peak updraft speeds in Figure 4a, trends were consistent with updraft speed distributions observed in Figure 3. The peak updraft speeds in
Figure 3. Time series of (a) 5 m s$^{-1}$ and (b) 10 m s$^{-1}$ updraft volumes corresponding to each supercell are shown. Composite altitudinal frequency diagrams of vertical velocity in the (c) normal February 6, 2008 supercell, (d) normal April 11, 2008 supercell, (e) anomalous April 10, 2009 supercell, and (f) anomalous April 22, 2017 supercell are shown in color fill. Approximate mixed-phase region boundaries of 0°C and −40°C are marked (solid black lines) with −10°C increments plotted between (dashed black lines) are shown in (c–f).
the anomalous April 10, 2009 supercell were comparatively faster than observed in the other storms, with a minimum difference of 15.4 m s\(^{-1}\) in the median. Although the notches in the normal February 6, 2008 supercell and the anomalous April 22, 2017 supercell were least offset, none of the notches overlapped, indicating the medians of peak layer-maximum updraft speeds were each significantly different from each other even as some distributions overlapped. The anomalous April 10, 2009 supercell’s distribution in particular exhibited no overlap with that of the other storms, consistent with inferences drawn from Figure 3 that it was characterized by the fastest updrafts in the data set.

4.1.3. Vertical Updraft Structures

The anomalous and normal supercell vertical velocity distributions (Figures 3c–3f) exhibited differences in vertical structure. With indications that the anomalous supercells were deeper, larger frequencies of greater updraft speeds were observed above 9 km in the anomalous supercells (Figures 3e and 3f). Figure 4b shows the distributions of the altitudes of the peak layer-maximum updraft speeds relative to the melting level in each case, facilitating a more direct comparison of the spatial distribution of updraft speeds within the mixed-phase region. Part of the structural differences observed between storms in Figures 3c–3f can be attributed to differences in storm depth, as the normal storms were relatively more shallow than the anomalous supercells with differences in 18 dBZ echo top heights ranging from 1.2 to 2.7 km (Figure 4b). However, analysis of the peak vertical winds indicates that the depths over which relative maxima occurred were not necessarily deeper in anomalous storms (Figure 4b). Rather, the most notable differences were observed in the location of peak layer-maximum updraft speeds relative to the mixed-phase region between the two subsets. The majority of each of the normal storm’s peak layer-maximum updraft speeds were confined to the mixed-phase region, with less than 10% located at heights above \(-40^\circ\)C. By contrast, greater than 25% of peak layer-maximum updraft speeds in the anomalous supercells were observed at heights above \(-40^\circ\)C. Furthermore, less than 5% of each of the anomalous storm’s peak layer-maximum updraft speeds were observed within the warm cloud region compared with 15% or greater of the normal storms’ peak layer-maximum updraft speeds (Figure 4b).
In addition to differences in distributions of updraft speed, differences in updraft area with height were also observed when average vertical profiles of updraft areas were compared (Figure 5). Each anomalous supercell was characterized by larger average 5 m s\(^{-1}\) and 10 m s\(^{-1}\) updraft areas through the full depth of each storm compared with the normal April 11, 2008 supercell (Figures 5b and 5d). However, the relationships were more variable when the anomalous storms were compared with the normal February 6, 2008 storms, in which the 5 m s\(^{-1}\) updraft area was larger throughout most of the middle mixed-phase region between approximately \(-5^\circ\)C and \(-25^\circ\)C than in the anomalous supercells (Figures 5a and 5c). Most notably, the 10 m s\(^{-1}\) updraft area, considered as a demarcation of the updraft core, was greater in the anomalous storms in the mixed-phase region than in the normal storms, particularly above the height of \(-20^\circ\)C in the upper mixed-phase and glaciated regions. Comparing the differences in the vertical distributions of anomalous and normal storm updraft areas in Figure 5, it is evident that the area of the updraft core was larger at higher altitudes in the anomalous storms than in the normal storms in addition to being faster. Although Figure 5 reflects average profiles, additional analyses (not shown) confirmed that trends and implications.
are robust to minor temporal variations in updraft volume observed in Figures 3a and 3b, especially during the mature periods of each supercell.

The anomalous April 10, 2009 supercell was characterized by a stronger updraft in a deeper storm compared with the other supercells, and this greater depth may have contributed to its somewhat larger 10 m s⁻¹ updraft volume compared with the other storms (Figure 3b). With the exception of the normal April 11, 2008 supercell, however, total updraft sizes were comparable between three of the supercells for a substantial portion of their analysis periods, especially considering their 5 m s⁻¹ updrafts (Figures 3a and 3b). Bulk updraft characteristics were particularly comparable between the normal February 6, 2008 supercell and the anomalous April 22, 2017 supercell with regard to the range and depth of vertical velocities (Figure 3). However, the distinct separation in the notches between anomalous and normal box plots in Figure 4b informs that the medians of the vertical distributions of their peak updraft speeds were significantly different at the 95th percentile confidence level, highlighting the contrast in structure of these two individual updrafts and two sets of storms. While this difference could be attributed to disparities in storm size, comparisons of vertical distributions between the normal February 6, 2008 and anomalous April 22, 2017 supercells particularly suggest that structural contrasts were more prominent. Specifically, the distance between the 75th percentile peak velocity altitude and the 18 dBZ echo top height was 1.1 km less in the anomalous April 22, 2017 supercell than in the normal February 6, 2008 supercell (Figure 4b). This difference indicates that peak level-maximum updraft speeds were shifted higher in the anomalous April 22, 2017 supercell relative to storm top and were not only higher because of its higher storm top.

4.2. Mixed-Phase Precipitation Microphysics in the Updraft

The quantities and vertical distributions of PIM in the supercell updrafts (Figure 6) were used for comparison between cases particularly because of the role of precipitation ice in ice-ice collisional NIC and its relationship with the dominant layer of the overall charge structure. Specifically, regions of riming hydrometeors, or greater PIM, are expected to be associated with dominant regions of positive charge in anomalous storms and dominant regions of negative charge in normal storms. In each supercell, total PIM associated with the updraft (Figure 6a) trended similarly with changes in updraft intensity as inferred from charge structure evolution and observed through updraft properties (Figures 2 and 3). However, the anomalous storms exhibited greater PIM quantities than the normal storms, particularly after the anomalous April 10, 2009 supercell reached maturity (Figure 6a). These observations are consistent with the expectation that more efficient riming of graupel particles takes place in higher LWC in anomalous storms, promoting positive charging of the riming hydrometeors. Understanding differences in the characteristics of large ice quantities, types, and vertical distributions between the normal and anomalous storms may provide insight as to how the anomalous supercell charge carrier populations differed.

While hail, HD graupel, and LD graupel are all rimed ice categories, hail and HD graupel in particular represent the larger, denser, and more efficient riming hydrometeors. Appreciable hail fractions were observed in all but the normal February 6, 2008 supercell (Figure 6b), consistent with generally lower frequencies of higher PIM values at all levels in the normal February 6, 2008 supercell compared with the vertical PIM distributions in all storms (Figures 6c–6f). The normal February 6, 2008 supercell was also the only supercell with which large hail reports were not associated, though the early morning hour at which the storm occurred may have presented reporting challenges (e.g., Ortega et al., 2009; Trapp et al., 2006). By contrast, a maximum hail size of 4.44 cm in diameter was reported with each of the other three storms. These observations suggest riming processes in the normal February 6, 2008 supercell were less efficient, possibly resulting from some combination of reduced mixed-phase LWC or a kinematic environment that did not favor sufficient residence time of riming particles for hail growth. In this way, the normal February 6, 2008 supercell also markedly contrasts with the anomalous April 22, 2017 supercell with which it otherwise shared similar bulk kinematic trends over time (Figures 3a and 3b). Hail contributed the most to PIM within the updraft in the anomalous April 22, 2017 supercell with fractions of approximately 50% or greater in the upper mixed-phase region (Figure 6b). This is consistent with a generally higher total updraft PIM quantity over the analysis period (Figure 6a) as well as high values of PIM observed throughout the mixed-phase region (Figure 6f). Coupled with multiple reports of large hail of up to 4.44 cm in diameter, these metrics indicated that efficient riming processes within the anomalous April 22, 2017 supercell likely
Figure 6. Time series of (a) total precipitation ice mass (PIM) within updraft of ≥5 m s⁻¹ corresponding to each supercell are shown. (b) The fractions of PIM within updraft of ≥5 m s⁻¹ at each altitude in each storm attributed to low density (LD) graupel (yellow), high density (HD) graupel (green), and hail or large drops and hail (pink) categories are shown as stacked bar charts. Composite altitudinal frequency diagrams of PIM within updraft of ≥5 m s⁻¹ in the (c) normal February 6, 2008 supercell, (d) normal April 11, 2008 supercell, (e) anomalous April 10, 2009 supercell, and (f) anomalous April 22, 2017 supercell are shown in color fill. Approximate mixed-phase region boundaries of 0°C and −40°C are marked (solid blue lines) with −10°C increments plotted between (dashed blue lines) in (c–f).
extended throughout a substantial depth of the mixed-phase region and may have been supported by sufficient mixed-phase LWC and/or conducive kinematic structure.

The smaller normal April 11, 2008 supercell was characterized by a relatively smaller, weaker updraft as well as the least PIM compared with the other storms (Figure 6a). However, its precipitation ice composition was similar to that of the anomalous April 22, 2017 supercell in the lower mixed-phase region with between 25% and 50% hail in multiple layers (Figure 6b). Comparing the normal supercells, higher frequencies of greater PIM values throughout the mixed-phase region in the normal April 11, 2008 supercell (Figure 6d) was consistent with comparatively higher riming efficiency than indicated in the normal February 6, 2008 supercell (Figure 6c). PIM frequencies within the lower mixed-phase region (heights below the level of −20°C) of the normal February 6, 2008 and anomalous April 22, 2017 supercells (Figures 6c and 6f) exhibited the strongest similarities, indicating similar ice precipitation efficiency at warmer mixed-phase temperatures.

Despite observations of large hail associated with the anomalous April 10, 2009 supercell, LD graupel was the primary contributor to PIM and its average total hail fraction was relatively low at 7%. Compared with all other supercells, the anomalous April 10, 2009 supercell was characterized by lower fractions of the larger, denser riming hydrometeors in the lower mixed-phase region (Figure 6b). Even so, on average, the anomalous April 10, 2009 supercell had $>1.9 \times 10^7$ kg more PIM associated with hail and HD graupel in the lowest 3 km of the mixed-phase region of the updraft compared with the same region in the normal supercells (not shown). Compared with the anomalous April 22, 2017 supercell, the anomalous April 10, 2009 supercell exhibited lower fractions of larger riming hydrometeors throughout the entire mixed-phase region (Figure 6b). Specifically, it had $2.9 \times 10^7$ kg less PIM associated with hail and HD graupel on average in the lowest 3 km of the mixed-phase region of the updraft and $5.9 \times 10^7$ kg less in the combined upper 2 km of the mixed-phase and glaciated regions of the updraft compared with the anomalous April 22, 2017 supercell (not shown). This corresponded with generally higher frequencies of lower PIM values observed throughout the mixed-phase region in the anomalous April 10, 2009 supercell (Figure 6e) compared with lower frequencies of higher PIM values observed in the anomalous April 22, 2017 supercell (Figure 6f). These comparisons indicate that although its relative distribution of larger precipitation ice was lower than in the normal storms, the anomalous April 10, 2009 supercell still contained more large precipitation ice in the form of hail and HD graupel than the normal storms. However, it did not contain as much large precipitation ice as observed in the anomalous April 22, 2017 supercell. The temporal increase in updraft PIM observed in the April 10, 2009 supercell that represented the eventual maximum of the data set could be attributed to the greater capability of its large updraft (Figures 3a and 3b) to support a larger quantity of smaller, less dense graupel. In addition, its lightning flash rates were generally comparable to the normal supercells prior to their weakening phases and significantly lower than the other anomalous supercell (Figure 2). These multiple metrics suggest that in spite of its robust updraft, large riming ice growth processes in the anomalous April 10, 2009 supercell may have been relatively inefficient by comparison.

Comparisons of the vertical profiles of PIM and hydrometeor type alluded to structural differences in precipitation microphysics observed between the two sets of supercells. For instance, despite relatively lower hail fractions observed in the anomalous April 10, 2009 supercell, both anomalous storms exhibited minor hail fractions in the upper mixed-phase region that extended into the glaciated region, whereas the normal storms did not (Figure 6b). The presence of hail as well as the greater depth to which LD graupel was observed contributed to PIM at higher altitudes along with higher frequencies of larger PIM within the glaciated region in the anomalous storms (Figures 6e and 6f). While differences in the depths of PIM observed can partially be attributed to differences in the storm depths, the anomalous storms were characterized by higher frequencies of larger PIM values closer to their respective storm tops than were the normal storms, indicating riming hydrometeors were more vertically extensive therein. This may have been the result of greater depths over which riming growth was supported in anomalous storms or the ability of anomalous storms to loft rimed hydrometeors to higher altitudes given stronger updraft profiles aloft (Figures 3e and 3f).

### 4.3. Environmental Conditions

Coarse environmental data can be considered within the context of the observed kinematic and microphysical structure of each supercell to further understand the differences in gross charge structure. Each of the four storms shared some environmental commonalities, as perhaps expected given their common
formation in the Southeastern US (Table 2). Though other comparative studies have shown substantial differences in these parameters, there were no obvious trends differentiating these Southeastern anomalous and normal supercell environments in surface temperature, surface dew point temperature, mean mixing ratio in the lowest 100 hPa, depth of the free convective layer, or 0–6 km shear. However, parameters that exhibited differences included WCD, instability, metrics of environmental moisture, and a proxy for environmental CCN. It should be noted that while some studies have identified some of these parameters as discriminators between anomalous and normal supercell storm environments (e.g., Carey & Buffalo, 2007; Fuchs et al., 2018), differences in parameters identified in the comparisons discussed herein are not always observed (e.g., Chmielewski et al., 2018; Fuchs et al., 2018; Lang & Rutledge, 2011).

### 4.3.1. CCN Proxy

Aerosol concentration profiles derived from MERRA-2 reanalysis are shown in Figure 7, from which coarse estimates of CCN concentrations were inferred. These estimates indirectly suggest the nature of the cloud and precipitation particle size distributions in each of the storms, though there are no known methods by which CCN can be directly quantified from aerosol data. The aerosol profiles corresponding to the normal April 11, 2008 supercell (Figure 7b) and the anomalous storm environments (Figures 7c and 7d) were more similar, particularly below the height of approximately -10°C and near the inferred CBHs. However, the relatively low aerosol concentration in the normal February 6, 2008 profile (Figure 7a) is consistent with the idea that fewer CCN may have prevented competition for water vapor and did not comparatively reduce warm
precipitation efficiency. By contrast, the anomalous April 10, 2009 supercell aerosol profile indicates that more aerosols were present within the warm cloud region with an absolute maximum just above the height of the LCL, with relatively higher concentrations that persisted into the mixed-phase region (Figure 7c). The normal April 11, 2008 supercell also possessed relatively greater aerosol concentrations throughout the mixed-phase region. However, it did not exhibit a similar relative maximum near the inferred CBH as observed on April 10, 2009 (Figure 7b). Generally, these profiles are consistent with the idea that ample CCN were available to the anomalous supercells, potentially shifting the particle size distribution toward smaller particles and effectively reducing the efficiency of warm rain processes.

4.3.2. Instability

Similar to WCD, results concerning the relationship between instability and ACS development have been varied. Generally, higher NCAPE, or CAPE distributed over specific layers (Blanchard, 1998), has consistently discriminated anomalous from normal storms, and CAPE within the mixed-phase region has been shown to be higher in anomalous storms compared with normal storm environments (Carey & Buffalo, 2007). CAPE values associated with the Southeastern supercells ranged from 447 J kg⁻¹ to 2123 J kg⁻¹, with NCAPE varying between 0.06 m s⁻² and 0.19 m s⁻² (Table 2). For each metric, the greater values corresponded with the anomalous supercells, with nominal minimum differences between the two subsets of storms of 239 J kg⁻¹ and 0.03 m s⁻². Generally, trends toward higher NCAPE in anomalous storm environments were consistent with their more robust updrafts.

4.3.3. Warm Cloud Depth

The shallowest environmental WCD of approximately 2,460 m was associated with the anomalous April 10, 2009 supercell environment, given an LCL height of 643 m and an environmental melting level of 3,105 m (Table 2). The second most shallow WCD of approximately 2,720 m was associated with the anomalous April 22, 2017 supercell environment, resulting from a comparatively high 900 m LCL and an environmental melting level height of 3,617 m. The lowest LCL of 385 m associated with the normal February 6, 2008 supercell environment along with an environmental melting level of 3,861 m resulted in the second deepest WCD of the sample of approximately 3,280 m. Despite having the highest LCL of 1,022 m, the comparatively high environmental melting level height of 4,365 m in the normal April 11, 2008 supercell environment resulted in the deepest WCD of approximately 3,340 m.

These general trends agree with findings in the literature that anomalous supercells sometimes have a more shallow WCD than their normal counterparts, where a shallower WCD likely reduces the depth over which warm precipitation may grow and deplete cloud LWC. However, the anomalous WCDs in these Southeastern supercells were deeper by approximately 1,000 m than values reported in the literature associated with other anomalous supercells (e.g., Carey & Buffalo, 2007; Chmielewski et al., 2018; Fuchs et al., 2018). Although anomalous WCDs were deeper, the minimum 820 m difference between the shallowest anomalous and the deepest normal WCDs was similar to that observed by Fuchs et al. (2018), documented as a difference of 892 m between WCDs in different regions. While shallower WCD has been linked as a discriminator separating anomalous from normal storm environments, others have identified the opposite when examining storms in the same region on the same day (Chmielewski et al., 2018) and storms in the same region on different days (Lang & Rutledge, 2011). Observations of these four Southeastern supercell environments support findings that while a shallower WCD may promote conditions favoring the development of ACSs, they also suggest that the determination of a sufficiently shallow WCD likely varies relative to other environmental conditions that affect precipitation efficiency, cloud LWC, and ice growth processes.

4.3.4. Environmental Moisture

Relationships between the storms based on metrics of environmental moisture differed according to the layers considered. Mixing ratios in the lowest 100 hPa were not substantially different between subsets of storms, varying between 11 g kg⁻¹ and 13 g kg⁻¹ (Table 2). However, relative humidity (RH) in the 700–500 hPa layer trended lower in anomalous storms with values of 21% and 36% compared with values of 41% and 55% in normal storms (Table 2).

Metrics that encompassed more of the full storm depths also tended toward lower moisture in anomalous storm environments. Precipitable water between the surface and 400 hPa was lower in anomalous storms at
2.3 and 3.2 cm compared with 3.5 and 3.8 cm in normal storms (Table 2). RH in the full depth, meanwhile, was 41% and 54% in anomalous storms and 54% and 68% in normal storms (Table 2). The highest RH was observed in the normal February 6, 2008 environment, while the normal April 11, 2008 and anomalous April 22, 2017 environmental RH values were most comparable in the middle of the observed range. RH through any depth considered was consistently lowest in the anomalous April 10, 2009 supercell, which may have also contributed to its inferred relative precipitation inefficiency that extended from the warm cloud region into the mixed-phase region. That is, despite a robust updraft that may have been able to support high-LWC riming (Figure 3e), lower flash rates (Figure 2c) and reduced quantities of large pre-
- precipitation ice (Figure 6e) in the anomalous April 10, 2009 supercell suggest that ice growth processes were impacted, especially compared with the other anomalous supercell.

5. Discussion on Proposed Contributions to Anomalous Charging

The following discussion integrates and summarizes connections in the data to link properties of each supercell to the characteristics that may have influenced their charge structures. Though the rich data set presented herein provided useful indications of kinematics, qualitative lightning activity, and precipitation microphysics, there remain elements of thunderstorm electrification that were not sampled. In the absence of true cloud metrics including LWC and water vapor, the following discussion remains necessarily speculative when the available data are not sufficient to support firm conclusions. These aspects, however, present ample opportunities for future study, which are elaborated upon in the final section.

5.1. Evaluation of Updraft Structure Relative to Charging Hypotheses: Low-Level Updraft

More robust updrafts in lower levels are thought to support ACS development by reducing warm precipitation efficiency, allowing cloud LWC to reach the mixed-phase region for more efficient riming growth in favor of increased positive charging of riming hydrometeors. In addition to the benefit of fast updrafts to reduce cloud water residence time in the warm cloud, wide updrafts as well are thought to be less susceptible to entrainment, ultimately preserving LWC and maintaining fast updraft speeds. Generally, higher CBHs are thought to favor broader updrafts (e.g., Carey & Buffalo, 2007; E. R. Williams et al., 2005). Therefore, anomalous storms are thought to be associated with elevated cloud structure as well as broad, fast updrafts in the warm cloud region that contribute to initial reduction of LWC loss to warm precipitation growth. Although the maximum updrafts were located at the top of the mixed-phase region in the anomalous storms, the anomalous April 10, 2009 storm was characterized by higher updraft speeds below the melting level than observed in the other three storms (Figure 3e), supportive of the hypothesis related to the low-level updraft.

The anomalous April 22, 2017 supercell also exhibited a faster, broader updraft in the warm cloud region than the normal April 11, 2008 supercell. However, the normal February 6, 2008 supercell was characterized by both faster updraft speeds and a larger updraft core below the melting level than the anomalous April 22, 2017 supercell (Figures 3c, 3f and 5c). If a large, fast updraft, particularly in the warm cloud region, was the primary requirement supporting conditions favorable for anomalous charging, the February 6, 2008 supercell would have been expected to be anomalous rather than the April 22, 2017 supercell. However, a weaker updraft that comparatively facilitates warm precipitation efficiency may also be compensated by other environmental factors that adequately reduce warm precipitation efficiency. The idea of WCT (Fuchs et al., 2018) addresses this idea. However, it was observed that the distribution of WCTs based on warm updraft speed percentiles and a representative hydrometeor fall speed of 2 m s^{-1} exhibited substantial overlap in the Southeastern anomalous and normal supercells (Figure 8). Moreover, while the tails of the two anomalous WCT distributions were generally lower than the normal storms on the order of 5–8 min, the anomalous storms WCTs were not comparable to those observed in anomalous storms in Colorado by Fuchs et al. (2018). Instead, all more closely resembled the distributions observed in normal storms in AL analyzed by Fuchs et al. (2018). These data indicate that additional environmental factors beyond those that support robust warm cloud updrafts likely contributed to the observed ACSs.
5.2. Evaluation of Updraft Structure Relative to Charging Hypotheses: Mixed-Phase Updraft

Within the mixed-phase region, similar arguments apply as were discussed with respect to low-level updrafts. Large, fast updrafts are more resistant to entrainment of dry air and may be less favorable for particle recirculation which can increase competition for LWC and reduce riming efficiency (Chmielewski et al., 2020; Fuchs & Rutledge, 2018; MacGorman et al., 2005, 2011; E. R. Williams et al., 2005).

The mixed-phase updraft structures in anomalous storms were larger and faster in the middle to upper mixed-phase region compared with the larger, faster updrafts observed in the lower to middle mixed-phase regions of the normal storms. Greater depths of riming and large riming hydrometeors were indicated in the anomalous supercells (Figure 6b) and were attributed to generally more robust upper-level updrafts. In the anomalous supercells, maximum updraft speeds were either aligned with or located above the dominant positive charge region and extended into the glaciated region. At these temperatures, collisional NIC involving riming hydrometeors would have been inactive as a result of homogeneous freezing (although studies have shown that some amount of charging may occur in the absence of LWC, e.g., Dye & Bansemer, 2019; Emersic & Saunders, 2020; Mitzeva et al., 2006). Therefore, the most likely region of positive charging of riming hydrometeors did not strictly coincide with storm maximum updraft speeds, as similarly observed by Chmielewski et al. (2020), and it is not suggested that the ACSs arose because of these elevated maximum updrafts. Rather, these updrafts may have been influenced by conditions that either more directly contributed to or were related to ACS development.

Development and maintenance of fast updrafts at higher levels in anomalous storms may have been supported in part by the additional latent heat release (e.g., Fuchs & Rutledge, 2018) associated with their apparent greater depths of riming or freezing of supercooled liquid water near ~40°C that supported deeper riming in the anomalous storms. However, the relative differences in the low-to mid-level updraft structure between the Southeastern anomalous and normal supercells were collocated with observed dry layers in the anomalous supercells, indicating that dry air entrainment may have impacted updraft profiles of the anomalous supercells. Specifically, it is possible that maximum updrafts were identified at higher altitudes in the anomalous storms because the mid-level updrafts were comparatively reduced by negative buoyancy associated with dry air entrainment. The “top-heavy” updraft profiles of anomalous storms were comparable to updraft profiles in supercell simulations conducted by Grant and van den Heever (2014) in which dry layers were imposed. Compared with control simulations in their study, those that included dry layers centered at 3.5 km AGL (near 650 mb) were able to maintain strong maximum updrafts aloft though low-level velocities were diminished locally where affected by injection of dry air (Grant & van den Heever, 2014). Additionally, results from simulations by Nowotarski et al. (2020) suggest that as much as 30% of the air

Figure 8. (a) Violin plots showing the distributions of warm cloud updraft speeds in each storm. (b) Violin plots showing the distributions of warm cloud residence times (WCTs) in each storm, assuming a particle fall velocity of 2 m s⁻¹.
within the updraft core may originate above the effective inflow layer, which is typically observed within the lowest 3 km. Therefore, similar entrainment above the inflow layer in the anomalous supercells could have introduced dry air from the 700 to 500 mb region to the low-to mid-level updraft, contributing to local buoyancy reduction and seemingly elevated updraft maxima.

5.3. Effects of Entrainment on Charging

Indicators of water vapor content suggested that the anomalous storm environments were drier, particularly just below the melting level and into the lower mixed-phase region (Table 2). Additional detail regarding the dry layers can be found in model reanalysis soundings provided in the Supporting Information S1. Observations of comparably drier layers in the Southeastern anomalous storms in which positive charge layers were inferred are consistent with others reported in the literature (e.g., Carey & Buffalo, 2007; Carey et al., 2003; Chmielewski et al., 2018; Knapp, 1994), including numerous ACSs in low precipitation supercells (e.g., Branick & Doswell III, 1992; Curran & Rust, 1992; Lang et al., 2004; MacGorman & Burgess, 1994; Seimon, 1993; Tessendorf, Wiens, & Rutledge, 2007).

The role of dry air entrainment is often considered from an LWC perspective in the context of ice-ice collisional NIC. As it pertains to precipitation efficiency, several effects of entrainment have been suggested. As discussed in Chmielewski et al. (2018), dry air in lower levels of storms may decrease droplet sizes, reducing precipitation efficiency. However, the possible limiting effects of dry air entrainment to LWC include evaporation of cloud liquid water needed for riming (e.g., Hoffmann, 2020) and erosion of buoyant parcels that would both increase the residence time of LWC in favor of precipitation efficiency and reduce the updraft’s ability to transport larger LWC (e.g., Chmielewski et al., 2018; Fuchs et al., 2018).

The effects of dry air entrainment on the relative diffusional growth of ice particles and its impact to particle-scale charging have not been considered as deeply in the literature on storm-scale electrification. However, it has been shown in theoretical models and from limited laboratory studies that reduced supersaturation with respect to ice relatively increases the magnitude of positive charging of riming precipitation ice for a given LWC over a range of mixed-phase temperatures (Berdeklis & List, 2001; Mitzeva et al., 2005; Saunders et al., 2006; Tsenova et al., 2010).

While factors such as a shallow WCD and fast low-level updraft (in the case of the anomalous April 10, 2009 supercell) may have promoted mixed-phase LWC, the differences in WCT in anomalous storms were not substantial enough to explain the resultant drastic difference in charge structure. It is suggested that entrainment of dry air may have contributed to anomalous charging in these storms in part by reducing the warm rain efficiency and facilitating the availability of LWC in the mixed-phase region. Dry air may have also reduced supersaturation with respect to ice in the charging region, thereby altering relative growth rate relationships and decreasing the LWC required for positive graupel charging. This is not to say that the impacts of dry air are more significant than or preclude the requirement of LWC for anomalous charging. Rather, it is suggested that dry air entrainment and its impact on saturation ratios can augment and increase the effects of requisite available LWC.

The relative dearth of observed Southeastern ACSs suggests that influential environmental factors are likely uncommon to the Southeast. It is unclear how often Southeastern convective environments include mid-level dry air, though it is expected to be somewhat infrequent given proximity to the Gulf of Mexico as a source of deep moisture. The contrast offers a physically consistent explanation for the relatively low frequency of anomalous storm observations in the Southeast compared with the Great Plains where environmental dry layers are more prevalent as a result of elevated mixed layers and/or the proximal dry line (e.g., Chmielewski et al., 2018; Grant & van den Heever, 2014; Ribeiro & Bosart, 2018).

5.4. Summary of Potential Factors Manifesting Apparent Charge Structure

The anomalous April 10, 2009 supercell event was superlative, characterized by a robust updraft and an environment that was most consistent with standards for anomalous charging established in the Great Plains. Including greater concentrations of warm cloud aerosols, conditions were seemingly favorable for the availability of LWC in the mixed-phase updraft. However, it is possible that dry air entrainment contributed to
anomalous charging as well. Having formed in a dry pre-convective environment, the anomalous April 10, 2009 storm was apparently not particularly precipitation-efficient in the warm cloud or mixed-phase regions, given its relatively low lightning flash rate, low hail fraction, and low total PIM compared with the other anomalous April 22, 2017 storm. Its shallow WCD and particularly broad, fast updraft may have compensated sufficiently to provide mixed-phase cloud LWC for riming growth despite deleterious effects of environmental dry air. The low-to mid-level updraft structure indicated impacts from dry air entrainment, which may have also inhibited ice crystal depositional growth relative to riming growth, shifting the LWC requirement for positive graupel charging. Such a shift is thought to lower the LWC threshold to achieve an ACS consisting of deep, positive charging of riming hydrometeors.

The anomalous April 22, 2017 supercell exhibited less definition in charge structure, indicating relatively increased charge structure complexity. This complexity could have been attributed to three-dimensional variability in kinematics as well as cloud microphysical properties. That is, conditions within the updraft may have been variably favorable for both anomalous and normal charging, with predominantly anomalous charging occurring in the net perspective (e.g., Chmielewski et al., 2020). The structural similarities between the anomalous April 22, 2017 storm and the normal storms support this possibility, as do indications of some environmental overlap. It is possible that different factors contributed to comparative net anomalous structure in the April 22, 2017 supercell than contributed to the April 10, 2009 ACS.

The gross updraft structures were most similar between the normal February 6, 2008 supercell and the anomalous April 22, 2017 supercell (Figures 3a and 3b). The primary differences between the storms included a shallower WCD, increased warm cloud aerosol concentrations, and drier environmental conditions in the anomalous supercell but a faster low-level updraft in the normal supercell. WCTs were comparable between the two storms, indicating similar support for mixed-phase liquid water content from the contrast between WCD and low-level updraft characteristics. However, the normal February 6, 2008 supercell was characterized by substantially less PIM in its structure and produced negligible hail fractions within the updraft. While prolific hail production is not necessary for an anomalous storm (e.g., Tessendorf, Wiens, & Rutledge, 2007), its absence suggests low mixed-phase cloud LWC or limited ice precipitation efficiency. With regard to the normal February 6, 2008 supercell, either condition is physically consistent with decreased charging inferred from the dramatically lower flash rate compared with the anomalous April 22, 2017 supercell. These are indications of relatively increased warm precipitation efficiency in the normal February 6, 2008 supercell, possibly due to decreased competition for droplet growth in the warm cloud region as a result of relatively fewer aerosols and comparatively increased low-level moisture. The relatively drier air in the lower and mid-levels of the April 22, 2017 supercell, by contrast, may have simultaneously reduced warm precipitation efficiency and impacted relative growth rate of ice crystals and graupel in the mixed-phase region, each supportive of greater net anomalous charging.

Metrics of environmental humidity were most similar between the normal April 11, 2008 supercell and the anomalous April 22, 2017 supercell of the four storms, as were the precipitation ice characteristics. However, the WCD was the deepest, the wet bulb zero height was the highest, and the precipitable water was greatest in the normal April 11, 2008 supercell environment, indicating moist low levels and efficient warm rain production despite relatively large aerosol concentrations in the warm cloud region. Even though these indications of warm precipitation efficiency suggested that mixed-phase LWC may have been reduced, relatively high hail fractions in the lower mixed-phase region were observed, indicating that conditions were favorable in some part of the storm for hail growth via riming. However, the observed charge structure suggests that weak kinematic support and factors favoring warm precipitation efficiency precluded sufficient LWC for positive charging of riming hydrometeors through a substantial depth of the storm despite favorable humidity values and evident riming efficiency. Had adequate kinematic support provided sufficient LWC to the mixed-phase region, it is possible that the dry environment similar to or warm cloud aerosol concentrations greater than that observed in the anomalous April 22, 2017 supercell may have favored a similar ACS in the otherwise normal April 11, 2008 supercell.
6. Conclusions

The lightning properties, kinematic structure, precipitation microphysics, and environmental characteristics of four supercell thunderstorms exhibiting normal and anomalous charge structures in the Southeastern US were examined and compared. These properties were used to evaluate the kinematic and microphysical conditions associated with and the environmental support for the inferred net charge structure in each storm. Primary characteristics differentiating the anomalous and normal supercell thunderstorms were identified as follows:

1. The anomalous supercells were characterized by dominant positive charge layers in the middle mixed-phase region where the normal supercells included dominant negative charge layers. Consistent with lightning observations in the Great Plains region, Southeastern anomalous IC:CG ratios were higher and these supercells had higher fractions of positive CG lightning flashes. Further, while both anomalous supercells exhibited higher peak flash rates than the normal supercells, the anomalous April 22, 2017 supercell flash rates were extraordinarily high, consistent with reports in other anomalous supercells. This characteristic coincides with numerous small flashes and is possibly related to reduced detection capabilities associated with anomalous storms by space-borne optical instruments (e.g., Conrad & Schultz, 2019; Zhang & Cummins, 2020).

2. The anomalous supercells were deeper storms with larger, stronger updrafts in the upper mixed-phase regions that extended into the glaciated region. By contrast, the most robust portions of updrafts in normal supercells were confined to the warm cloud and lower to middle mixed-phase regions.

3. The anomalous supercells had larger quantities of precipitation-sized ice in the upper mixed-phase and glaciated regions than normal supercells. These observations indicated more efficient riming growth through a deeper extent and/or greater kinematic support lofting rimed hydrometeors to higher altitudes than occurred in normal supercells.

4. Trends in environmental conditions that favored microphysical processes associated with anomalous charging included shallower WCDs, greater instability, increased warm cloud aerosol concentrations, and reduced measures of moisture in the lower to middle mixed-phase region. However, the values observed and the differences between anomalous and normal storms were not necessarily consistent with Great Plains storm environments. While environmental data were coarse in nature, general magnitudes of these environmental parameters and their differences reinforce the idea that the extent to which each is necessary relies on the combined influence of all factors that affect the mixed-phase microphysical parameter space.

5. Despite increased instability metrics, anomalous storm updrafts were not necessarily comparatively stronger than normal storm updrafts in the warm cloud region. This contributed to comparable WCTs between the two subsets, and as such, warm cloud conditions alone were insufficient to account for anomalous charging. In addition to addressing combined and competing influences of multiple environmental parameters, results suggest that mixed-phase aspects of anomalous charging may extend beyond LWC considerations to include effects of variable water vapor content on relative ice growth rates and associated charging of riming and non-riming ice.

These results further highlight documented complexities concerning the requirements and relative sufficiency of specific environmental conditions for anomalous charging. They also motivate several areas for future study to improve upon the clarity of results presented herein. For instance, the limited number of storms addressed by these analyses was the byproduct of a relatively rare phenomenon of Southeastern anomalous supercell development as well as limited quality observations of comparable storm mode, intensity, and duration. Not only would a larger sample of both anomalous and normal supercells be desirable, but future analyses of similar thunderstorm observations should also seek to broaden the pool of high-intensity normal storms for closer comparison with robust anomalous storms. Similarly, an environmental analysis study may improve understanding of how often the conditions thought to be associated with Southeastern anomalous storms are present, providing further insight regarding their frequency. To complement and advance understanding from observational results, future modeling studies should also continue to assess the sensitivity of ACS development to microphysical parameters. In particular, various environmental impacts to mixed-phase LWC such as instability, CCN and ice nuclei concentrations, RH, and the thermal profile should be addressed. Similarly, it would be of interest to further evaluate the impacts of environmental conditions such as CCN and RH on updraft structure and intensity as well.
The results also raise questions concerning the multi-phase aspect of ice precipitation growth processes in the context of particle-scale charging. For example, are the effects of dry air and supersaturation on relative ice crystal and graupel growth rates sometimes just as important as LWC to ice–ice collisional NIC processes? The potential role of supersaturation in gross charge structures in particular requires further consideration, though tools by which to address this topic are limited as of yet. For instance, laboratory charging studies have not yet tested the effects of water vapor on relative ice particle growth rates over a sufficient range of supersaturation ratio nor has the complete three-dimensional parameter space of temperature, LWC, and supersaturation been addressed in a laboratory setting. These empirical data are necessary to inform development of an appropriate parameterization for numerical simulations that may be used to test hypotheses for which observations are limited. From the observational perspective, current studies are limited by the lack of LWC or water vapor observations in deep convection and difficulty of direct measurement. In their absence, data and techniques that improve current estimates of properties such as CCN are of potential benefit. Surface-based instruments such as ceilometers, sun photometers, and Light Detection and Ranging (LiDAR) may provide aerosol optical depth data from which more specific aerosol concentrations could be derived in limited-area studies. Techniques utilizing satellite data to estimate CCN in convection may be more widely useful as well (e.g., Rosenfeld et al., 2014; Yue et al., 2019). Storm penetrating aircraft and similar efforts exist among the few avenues by which necessary direct measurements of CCN, LWC, and water vapor in convection may be acquired in the future. However, as current techniques are insufficient to explore LWC and supersaturation in thunderstorms at the required scales, laboratory and subsequent modeling studies hold the most immediate promise toward further clarity on their roles in charge structure polarity.

Acronyms

3DVAR Three-dimensional variation
ACS Anomalous charge structure
AGL Above ground level
ARMOR Advanced Radar for Meteorological and Operational Research
CAPE Convective available potential energy
CBH Cloud base height
CCN Cloud condensation nuclei
CG Cloud-to-ground
DBSCAN Density-Based Spatial Clustering of Applications with Noise
HD High density
HID Hydrometeor identification
IC In-cloud
LCL Lifted condensation level
LD Low density
LiDAR Light Detection and Ranging
LMA Lightning Mapping Array
LWC Liquid water content
MALMA North Alabama Lightning Mapping Array
NCAP Normalized convective available potential energy
NCS Normal charge structure
NIC Non-inductive charging
NOx Nitrogen oxide
PIM Precipitation ice mass
RAP Rapid Refresh
RH Relative humidity
RUC Rapid Update Cycle
VHF Very high frequency
WCD Warm cloud depth
WCT Warm cloud residence time

Data Availability Statement

Data from the ARMOR and NALMA corresponding to the 2008 cases and the April 10, 2009 case can be accessed via https://zenodo.org/record/3783694, license: Creative Commons Attribution 4.0 International, Carey and Blakeslee (2020b) and https://zenodo.org/record/3738553, license: Creative Commons Attribution 4.0 International, Carey and Blakeslee (2020a), respectively. Data from the ARMOR used for the April 22, 2017 case analysis can be accessed through the VORTEX-SE 2017 data catalog at the UCAR/NCAR Earth Observing Laboratory data archive (http://catalog.eol.ucar.edu/vortex-se-2017, Carey & Knupp, 2017). Data from the NALMA used for the April 22, 2017 case are available from the NASA Global Hydrology Resource Center Distributed Active Archive Center (https://earthdata.nasa.gov/eosdis/daacs/ghrc, Blakeslee, 2019). Access to the NLNDN data used is available through direct purchase from Vaisala (https://www.vaisala.com/en/products/national-lightning-detection-network-nldn). Other data, including KHTX level-2 data (https://www.ncdc.noaa.gov/data-access/radar-data/nexrad), RUC and RAP model analysis data (https://www.ncdc.noaa.gov/data-access/model-data/model-datasets/rapid-refresh-rap), and MERRA-2 data (https://gmao.gsfc.nasa.gov/reanalysis/MERRA-2-data-access/), are publicly accessible.

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