Analysis of Thundersnow Storms over Northern Colorado

MATTHEW R. KUMJIAN
Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

WIEBKE DEIERLING
Research Applications Laboratory, National Center for Atmospheric Research,* Boulder, Colorado

(Manuscript received 13 January 2015, in final form 12 August 2015)

ABSTRACT

Lightning flashes during snowstorms occur infrequently compared to warm-season convection. The rarity of such thundersnow events poses an additional hazard because the lightning is unexpected. Because cloud electrification in thundersnow storms leads to relatively few lightning discharges, studying thundersnow events may offer insights into mechanisms for charging and possible thresholds required for lightning discharges. Observations of four northern Colorado thundersnow events that occurred during the 2012/13 winter are presented. Four thundersnow events in one season strongly disagrees with previous climatologies that used surface reports, implying thundersnow may be more common than previously thought. Total lightning information from the Colorado Lightning Mapping Array and data from conterminous United States lightning detection networks are examined to investigate the snowstorms’ electrical properties and to compare them to typical warm-season thunderstorms. Data from polarimetric WSR-88Ds near Denver, Colorado, and Cheyenne, Wyoming, are used to reveal the storms’ microphysical structure and determine operationally relevant signatures related to storm electrification. Most lightning occurred within convective cells containing graupel and pristine ice. However, one flash occurred in a stratiform snowband, apparently triggered by a tower. Depolarization streaks were observed in the radar data prior to the flash, indicating electric fields strong enough to orient pristine ice crystals. Direct comparisons of similar lightning- and nonlightning-producing convective cells reveal that though both cells likely produced graupel, the lightning-producing cell had larger values of specific differential phase $K_{DP}$ and polarimetric radar–derived ice mass. Compared to warm-season thunderstorms, the analyzed thundersnow storms had similar electrical properties but lower flash rates and smaller vertical depths, suggesting they are weaker, ordinary thunderstorms lacking any warm (>0°C) cloud depth.

1. Introduction

Thundersnow events are rare phenomena that are said to occur when snow-bearing storms produce lightning and thunder. Because lightning in winter storms is less common and may seem counterintuitive, such events may take forecasters and the public by surprise and may pose an unexpected hazard. Though people tend to spend less time outdoors in winter months, lightning-related incidents have occurred at ski resorts, which host a variety of outdoor activities (e.g., Berger 1998; Zook 2014). Electrified winter clouds can pose an aviation hazard, as highly electrified areas have a high probability of causing aircraft- or helicopter-triggered lightning (e.g., Mäkelä et al. 2013; Wilkinson et al. 2013). Additionally, other winter weather hazards such as heavy snow have been associated with thundersnow events (Crowe et al. 2006). Thus, better understanding these unique storms could serve to mitigate potential risks.

Schultz and Vavrek (2009) provide a good review of historical observations of thundersnow events, which are limited because of their rarity. In the United States, early climatological studies of thundersnow events used surface reports. For example, Curran and Pearson (1971) investigated 76 thundersnow cases in the United States and found that only 1.3% of cool-season
thunderstorms (i.e., those occurring between October and May) produced snow, and that only 0.07% of snowfall observations were associated with lightning or thunder. Further, Market et al. (2002) analyzed 3-hourly METARs to construct a climatology of thundersnow events. They reported only three incidences of thundersnow in northern Colorado over a 30-yr period from 1961 to 1990, and an average annual occurrence of only 6.3 events across the conterminous United States (CONUS). Their climatology was not meant to be exhaustive and, instead, probably better characterizes the broader synoptic distribution of thundersnow events, as the use of surface reports for thundersnow detection is limited by the spatial coverage and resolution of surface observing stations, population density, etc.

These studies were performed before data from nationwide lightning detection networks were readily available. These networks provide high-resolution spatiotemporal information on lightning events across the CONUS (e.g., Orville 1991; Cummins et al. 1998; Orville and Huffines 2001; Holle 2014). More recent studies have made use of such lightning detection network data to investigate various aspects of thundersnow events (e.g., Pettegrew 2008; Market and Becker 2009; Rauber et al. 2014). For example, Pettegrew (2008) analyzed Vaisala’s National Lightning Detection Network (NLDN) data from 14 central United States thundersnow cases between October 2006 and April 2007, and found that only 1.4% of flashes during these events were associated with winter precipitation.

Such CONUS networks mostly detect cloud-to-ground (CG) lightning, which only makes up a fraction of all lightning discharges (Cummins et al. 1998; Boccippio et al. 2000). In recent years, however, the detection efficiency of total lightning from these networks has been improving (e.g., Rudlosky and Fuelberg 2010). Additionally, the growing availability of regional total lightning detection systems has enhanced our detection capabilities. Such systems include the Lightning Mapping Array (LMA), developed by the New Mexico Institute of Mining and Technology (NMIMT), which detects both CG and in-cloud (IC) flashes with high detection efficiency and accuracy (Rison et al. 1999; Krebbiel et al. 2000; Thomas et al. 2004; Lang et al. 2004) within a 100–200-km range. Thus, with these systems it is possible to document events that previously may have gone undetected.

Previous studies have also explored the necessary atmospheric ingredients for thundersnow storms. Schultz and Vavrek (2009) point out that the same conditions needed for warm-season thunderstorms must be present in thundersnow events: moisture, lift, and an unstable temperature profile. Curran and Pearson (1971) showed that the mean environment of their thundersnow cases was supportive of elevated convection, with a stable boundary layer topped by a near-neutral thermal profile. Market et al. (2006) found similar results for thundersnow events in the central United States, with the most-unstable level roughly 30–50 hPa above the top of the low-level temperature inversion. Rauber et al. (2014) found that lightning in midwestern winter cyclones tended to occur in regions of convective instability in the comma-head region, associated with dry upper-tropospheric air overrunning moister air from the Gulf of Mexico. They suggest that conditional symmetric instability (CSI) likely did not play a role in driving the convection. In contrast to typical summer convective storms, thundersnow storms are found to have lower convective available potential energy (CAPE) (e.g., Market et al. 2006; Mäkelä et al. 2013) and, consequently, weaker updrafts and much lower vertical cloud extent.

In addition to an environment supporting convection, cold (<0°C) air throughout the precipitation-bearing layer is required to produce snow at the surface. The composite sounding in Market et al. (2006) was cold enough throughout the lower troposphere to support snow. The surface temperature in thundersnow events often is found to be very near 0°C (e.g., Schultz 1999; Hunter et al. 2001; Stuart 2001). Market et al. (2002) found the mean surface temperature in their thundersnow cases to be about −1°C. Model soundings from the long-lived thundersnow event studied by Market et al. (2007) were consistent with these previous findings, as well.

Thunderstorm electrification leading to lightning is typically associated with storms exhibiting a strong updraft and a robust mixed-phase region (Williams and Lhermitte 1983; Dye et al. 1988; Zipser and Lutz 1994; Wiens et al. 2005; Deierling et al. 2008). Moreover, collisions between hydrometeors such as graupel and ice crystals in the presence of supercooled liquid water (SLW) is the basis for significant charging to take place via the noninductive charging mechanism that is thought to play a dominant role in lightning production (Takahashi 1978; Saunders 1993; Saunders and Peck 1998; MacGorman and Rust 1998; Takahashi and Miyawaki 2002). However, in the absence of appreciable SLW, noninductive charging may also occur (e.g., Dye and Willett 2007; Kuhlman et al. 2009), albeit at lesser charging rates. Furthermore, other charging mechanisms including inductive charging may also contribute to cloud electrification (e.g., MacGorman and Rust 1998). As thundersnow cases documented in previous studies are associated with lower CAPE, yielding lower updraft strength and vertical cloud extent, one expects...
that the availability of SLW, charging, and thus resultant flash rates should also be lower compared to their warm-season convective storm counterparts. For reference, flash rates in warm-season convective storms may range from several flashes per minute to hundreds of flashes per minute in high-end severe cases (e.g., Williams et al. 2005; Deierling et al. 2008; Fuchs et al. 2015). In contrast, total lightning flash rates in a variety of thundersnow storms have yet to be quantified.

Several authors have also investigated the precipitation structure of thundersnow events using radar. Pettegrew et al. (2009) used data from a single-polarization WSR-88D to investigate a thundersnow event over eastern Iowa and north-central Illinois. The echo top of the storm in their case study never exceeded ~3.7 km AGL, and its maximum values of reflectivity factor at horizontal polarization $Z_H$ never surpassed 40–45 dBZ. These maximum $Z_H$ values were confined to the lowest levels, consistent with particles growing by aggregation and/or riming as they descend. Market and Becker (2009) analyzed the spatial relationship between lightning and low-level $Z_H$ in thundersnow events, using data from the NLDN and preupgraded WSR-88D network. They found that the lightning flashes tended to be identified downstream of the largest $Z_H$.

However, $Z_H$ alone provides only limited insights into microphysical structures and processes. In contrast, the use of radar polarimetry provides more information about the bulk microphysical structure of thundersnow storms. A series of studies of an electrified storm over the Sea of Japan made use of C-band dual-polarization radar data (Fukao et al. 1991; Maekawa et al. 1992, 1993). These authors noted the collocation of radar-inferred graupel and ice crystals at the $-10^\circ C$ level preceding lightning flashes. They also found that the lightning-producing cell exhibited $Z_H > 40$ dBZ, whereas other nearby cells did not. The lightning production seemed to be tied to an increase in the inferred graupel content. However, these authors did not have available the full complement of polarimetric variables and derived products routinely obtained from today’s operational WSR-88D network.

Comprehensive information about the finescale electrical and microphysical structure of thundersnow is not available to date. In this paper, we report on four events that occurred in northern Colorado over a 5.5-month period during the 2012/13 cold season. For the first time, the electrical and microphysical thundersnow events are analyzed using lightning data from both the LMA and CONUS networks, and using dual-polarization WSR-88D data. During these four events, a total of 16 individual storms produced lightning. Though unclear whether the 2012/13 season was anomalous, it does hint at the possibility that thundersnow events are more common than previously documented [i.e., only 6.3 events on average annually in the CONUS; Market et al. (2002)], as observations from regional lightning detection networks such as the LMA detect CG and IC (i.e., total) lightning more comprehensively than do the CONUS networks used in the previous studies mentioned above.

This study aims to answer the following questions:

(i) How often are these storms detected by CONUS versus LMA networks?
(ii) What types of dual-polarization radar signatures in these Colorado thundersnow storms may be relevant to operational forecasters?
(iii) What are the electrical properties of these storms, and how do they compare to ordinary warm-season convective storms?

If the LMA data reveal events that are missed by the CONUS networks, this implies that previous climatologies constructed using only surface reports or CONUS network data may underestimate the frequency of occurrence of thundersnow. Any dual-polarization radar signatures, if found more widely in thundersnow events around the country, could provide increased situational awareness about a particular storm’s propensity to produce an otherwise unexpected lightning flash and attendant winter weather hazards such as heavy snow. Also, such observations can provide information about the bulk microphysical structure of lightning-producing convective snowstorms. In addition, studying low-flash-rate thundersnow events could be scientifically advantageous, as such events are marginal in nature and could be used to better define thresholds necessary for lightning flashes to occur. Finally, analysis of the microphysical and electrical structure of these storms and the environments in which they form can provide operational guidance for situations conducive to thundersnow.

The next section provides an overview of the instrumentation and data used to analyze thundersnow events in this study. Section 3 presents an overview of the cases, and an analysis of the LMA and CONUS lightning data, followed by the detailed polarimetric radar data analysis in section 4. The paper finishes with a discussion and summary of the main conclusions in section 5.

2. Instrumentation and data

a. Dual-polarization WSR-88Ds

The National Weather Service WSR-88D network recently has undergone an upgrade to have dual-polarization capabilities. In addition to the conventional moments of $Z_H$, Doppler velocity $V_r$, and Doppler...
spectrum width $\sigma_w$, the polarimetric radars provide the differential reflectivity $Z_{DR}$; differential propagation phase $\Phi_{DP}$, from which the specific differential phase $K_{DP}$ is computed; and the copolar correlation coefficient ($\rho_{hv}$ or CC). Descriptions of these polarimetric variables and their informative content can be found in Doviak and Zrnić (1993), Zrnić and Ryzhkov (1999), Straka et al. (2000), Bringi and Chandrasekar (2001), Ryzhkov et al. (2005), and Kumjian (2013a–c), among others.

Data from the WSR-88D polarimetric stations near Denver, Colorado (KFTG), and Cheyenne, Wyoming (KCYS), are used in this study. In addition to the polarimetric variables, the output of the operational hydrometeor classification algorithm (HCA) is used. The current HCA combines the informative contents of each polarimetric radar variable and determines the type of scatterer dominating the returned signals in each radar sampling volume (Park et al. 2009). Currently, 1 of 10 possible classes is assigned: light-to-moderate rain, heavy rain, big drops, rain mixed with hail, graupel, wet snow, dry snow aggregates, ice crystals, biological scatterers, and ground clutter and/or anomalous propagation.

b. Colorado Lightning Mapping Array

Three-dimensional total lightning data collected by the Colorado Lightning Mapping Array (COLMA) were used in this study to document the total flash rate from the thundersnow events as well as to deduce the location and polarity of charge layers within storm cells. During the period of study, the COLMA consisted of 15 stations within a 100-km radius equipped with global positioning system (GPS) sensors at the locations $x_i$, $y_i$, and $z_i$, where the subscript integer $i$ indicates the station number. The sensors record the time of arrival $t_i$ of received VHF sources emitted from lightning at around 60–66 MHz (Thomas et al. 2004). Lightning emits broadband radiation; breakdown processes of IC and CG lightning discharges such as negative leaders and streamers radiate strongly at VHF, making it possible to map IC and CG lightning in great detail at this frequency band. As such, the LMA provides comprehensive measurements of IC and CG lightning. In principle, if four or more stations detect a source, the three-dimensional position $x$, $y$, and $z$, and time $t$ of this source can be determined using the system of equations defined by Thomas et al. (2004):

$$c(t-t_i) = \left[ (x-x_i)^2 + (y-y_i)^2 + (z-z_i)^2 \right]^{1/2},$$

where $c$ is the speed of light. Six or more stations are required to allow a VHF source to be counted as a detection. The source locations are retrieved using a least squares $\chi^2$ minimization technique (e.g., Hiscox et al. 1984).

c. CONUS networks

Lightning captured by lightning detection networks covering the CONUS are used herein as well. These include Earth Networks Total Lightning Network (ENTLN), the United States Precision Lightning Network (USPLN) owned by Weather Services International (WSI), and Vaisala’s NLDN. Generally, these networks detect a high percentage of CG lightning and a lower percentage of IC lightning, as they mostly operate in the low-frequency (LF) band. Specifically, the NLDN consists of a combination of time-of-arrival sensors and wideband magnetic direction finder (MDF) sensors, collectively referred to as the Improved Accuracy through Combined Technology (IMPACT) sensors. They detect the electromagnetic radiation with frequencies around 10 kHz emitted from long transient current events such as CG lightning return strokes. The NLDN also detects a lower percentage of IC lightning. The two-dimensional, horizontal lightning locations $(x, y$ co-ordinates) are retrieved using a generalization of the least squares $\chi^2$ minimization technique discussed by Hiscox et al. (1984). NLDN measurements used herein include information about the location and time, polarity, signal strength, multiplicity (number of return strokes in a flash), and type (e.g., return stroke or IC event) of the measured lightning. Detailed descriptions of the NLDN can be found in Cummins et al. (1998), Cummins and Murphy (2009), and on the Vaisala website (http://www.vaisala.com/en/products/thunderstormandlightningdetectionsystems/Pages/NLDN.aspx). The ENTLN (http://earthnetworks.com/OurNetworks/LightningNetwork.aspx) uses broadband sensors operating in the LF and high-frequency (HF) bands between 1 kHz and 12 MHz (Liu and Heckman 2011). Similar to the NLDN, the ENTLN also detects and locates CG and IC lightning events based on the time of arrival of their waveforms. The HF measurements provide improved detection efficiency for IC lightning. The USPLN (http://www.uspln.com/uspln.html) is also a time-of-arrival system using broadband sensors with a bandwidth from 1.5 to 450 kHz. Similarly, the USPLN detects CG lightning and some IC lightning events.

All of these networks measure parts of lightning flashes (such as return strokes of CG flashes, etc.), which we will call lightning events. Each of the networks classifies an event as a CG or IC part of a lightning flash, and possibly what events belong to the flash. CONUS lightning data used in this study include the date, time, polarity, signal strength, and type of a lightning event. Flash information was also available from the NLDN.
and ENTLN. We determined flashes from USPLN-measured lightning events by applying the criteria described in Cummins et al. (1998). Note that the detection efficiency, classification of IC and CG events, and flash determination can vary between the networks depending on factors such as network station density and the way detected lightning signals are processed.

3. The thundersnow events

a. Environment and overview of the cases

The synoptic conditions for each thundersnow case reveal similarities (Fig. 1). On 11 November 2012, 28 January 2013, and 9 April 2013, a large-scale, positively tilted 500-hPa trough was located to the west of Colorado, over the Rocky Mountains. On 25 October 2012, the trough axis had just passed through Colorado and was lifting out. A ridge was located over the eastern United States in all four cases. In each case, a surface low was located to the south or east of the COLMA domain, and a cold frontal passage had just occurred, providing surface temperatures very near 0°C. The surface features provided northerly or northeasterly upslope flow for the Colorado Front Range and Palmer Divide (e.g., Rasmussen et al. 1995; Kumjian et al. 2014; Schrom et al. 2015), overlaid by larger-scale southwesterly flow aloft. The proximity of the upper-level trough would support large-scale ascent across the region, whereas topographically forced ascent was likely for some events, as discussed below.

Figure 2 shows the nearest soundings from each of the four cases. The lowest levels in the first three cases (Figs. 2a–c) exhibited temperatures >0°C; however, strong cold advection associated with the frontal passage soon decreased temperatures throughout the atmosphere to below 0°C. For example, at 0000 UTC 25 October 2012, the surface temperature near Denver was about 10°C (Fig. 2a), whereas near Cheyenne it was already below −1°C (not shown). By 0200 UTC, Denver was reporting snow and a surface temperature of 0.5°C (not shown). The soundings are quite consistent with those associated with thundersnow cases presented in previous studies (e.g., Curran and Pearson 1971; Market et al. 2006, 2007), with a low-level inversion or stable layer overtopped with nearly pseudoadiabatic lapse rates and small dewpoint depressions.
In each of the soundings, layers of potential instability (i.e., where the vertical gradient of equivalent potential temperature $\Delta \theta_e/\Delta z$ is negative) are found above the low-level stable region, also consistent with Market et al. (2006). Typically, these layers were associated with temperatures between $-13^\circ$ and $-30^\circ$C. Using high-resolution soundings (cf. Fig. 2d), Kumjian et al. (2014) found potential instability layers that were not observed by the nearby coarser operational soundings; thus, it is plausible that the magnitude or depth of the potential instability in these cases is underestimated by the operational soundings.

The analysis domain is shown in Fig. 3. Locations of all LMA sources from each case are overlaid, as are the locations of the COLMA stations, the WSR-88Ds near KFTG and KCYS, and the National Weather Service Denver sounding site (DNR). One immediately sees a preferred location for lightning activity south of the LMA network, seemingly anchored to the terrain feature known as the Palmer Divide. The northerly or northeasterly upslope low-level flow in each case would favor enhanced uplift at the Palmer Divide. This topographically forced ascent could provide a focused lifting mechanism required to release the potential instability. In the absence of other lifting mechanisms, the Palmer Divide then could serve as a focus for convective initiation and subsequent lightning activity. The presence of broader large-scale ascent associated with upper-level troughs in each case could also have helped provide the lift necessary to release the potential instability.

Other features of note in Fig. 3 include linear tracks of static discharges seen by the LMA that are produced by aircraft departing from or arriving at Denver International Airport (i.e., LMA source locations increased or...
produced 10 flashes over a period of about 30 min. In contrast, cells 2 and 4 only produced one flash, and cell 6 only produced two flashes. Six additional cells produced lightning flashes over the next 5 h (not shown in Fig. 4), with most of these not producing more than four flashes in their lifetime. One short-lived cell (cell 9) was more active, producing 15 flashes within 15 min. Another (cell 8) produced 19 flashes over about 90 min. The 11 November case (Fig. 4b) had the most electrically active storm of the dataset, with 28 flashes observed over about 1 h. Two cells produced flashes on 28 January (Fig. 4c), and one produced five flashes in about 10 min on 9 April 2013 (Fig. 4d).

Flash characteristics from each case are summarized in Table 1. The total number of flashes was determined by accumulating lightning data from all networks and matching them in space (within 1–2 km) and time (within 1 ms, the temporal resolution of CONUS network data). IC and CG events were determined based on the available data, including CONUS network classifications (e.g., Cummins et al. 1998) and interpretation of 3D LMA data (e.g., Thomas et al. 2003). The maximum flash rate (based on total flashes) in any 10-min period is also provided. Four cases produced more CG flashes than IC flashes, though one of these cases (cell 1, 25 October 2012) was at the edge of the LMA detection range, and thus not all flashes may have been detected. Another flash (cell 2, 25 October 2012) likely was tower initiated, as discussed further in section 4b. Otherwise, the majority of the cases produced more IC than CG lightning. Interestingly, for most cases the three CONUS networks exhibited variability in the number, type, and polarity of flashes detected, owing to differences in their design and performance, especially for cells with high IC-to-CG ratios. Note that there are ambiguities in deciding if flashes were IC or CG events based on the LMA data, particularly for flashes that initiate and progress very close to the ground and thus are limited by the line-of-sight restrictions of the LMA detections. Therefore, we caution that these flash-type classifications contain uncertainty.

Flash rates were determined from a 10-min moving window throughout each cell’s lifetime. The maximum flash rate in any 10-min window throughout a cell’s lifetime is shown in Table 1. Most cells had maximum flash rates <1.0 min$^{-1}$; the maximum flash rate for all events was from the 11 November 2012 cell, but was only

---

1 The farther distance from the LMA network also explains why cells 1 and 8 on 25 Oct 2012 had strokes detected by the CONUS networks but not the LMA.
Compared to flash rates in warm-season thunderstorms (e.g., Williams et al. 2005; Deierling et al. 2008), flash rates in these thundersnow cases were much lower, exemplifying the more marginal nature of these storms. Fuchs et al. (2015) determined that 0-dBZ echo-top heights in Colorado summertime thunderstorms varied between 12 and 18 km, whereas the thundersnow storm echo-top heights varied between 8 and 12 km for most cases (and did not exceed 14 km). This suggests less vigorous upward motion than warm-season convection in this region. In addition, observed temperature profiles suggest no warm-cloud (>-0°C) depth associated with these storms. These environmental factors are consistent with the observed lower flash rates (e.g., Williams et al. 2005; Fuchs et al. 2015). The charge structure of these cells was also examined. As such, the vertical distribution of LMA sources is shown in Fig. 5. The heights of the charge regions varied by case, occurring anywhere between 2 and 10 km AGL. Manual flash-by-flash analysis allows us to infer charge regions based on the propagation of LMA-detected VHF sources (e.g., Wiens et al. 2005). These analyses suggested that, in many of the cases, negative charge was located at lower altitudes and positive charge at higher altitudes in these storms. This positive-over-negative charge structure could be achieved if the faster-falling riming particles (e.g., graupel or rimed aggregates) gained negative charge in these storms at temperatures <0°C and intermediate values of SLW (e.g., Saunders and Peck 1998; Takahashi and Miyawaki 2002). Similar processes contribute to cloud electrification in warm-season thunderstorms; however, warm-season storms often appear to be dynamically more vigorous than these thundersnow events (i.e., greater updraft velocities

![Fig. 4. Time series of flash counts for each case: (a) 25 Oct 2012, (b) 11 Nov 2012, (c) 28 Jan 2013, and (d) 9 Apr 2013. The same horizontal and vertical axis ranges are used to facilitate comparison. Multiple cells during one event are shown in different color lines.](image-url)
leading to higher echo-top heights). We caution that the flashes did not always reveal such a simple charge structure, and longer-lived events exhibited evolving structures throughout their lifetimes. Thus, it is difficult to make any broad generalizations about the charge structure of thundersnow storms without a much larger dataset.

4. Radar analysis

a. Evolution of HCA fields

The upgraded WSR-88D network employs a hydrometeor classification algorithm (Park et al. 2009) that categorizes each radar pixel as 1 of 10 possible classes. In this section, we show illustrative examples of the HCA evolution for two cases: 11 November and 28 January (a similar evolution was observed for cells on 25 October and for 9 April; for brevity these are not shown). Figure 6 provides the evolution of the HCA output at 2.4° elevation using volume scans from 2350 UTC 10 November through 0007 UTC 11 November 2012, collected by the KFTG WSR-88D. Initially, mainly dry snow aggregates are classified in the echo to the southwest of the radar (Fig. 6a, centered at about \( x = -35 \text{ km}, y = -25 \text{ km}, \) where the origin of the coordinate system is the KFTG radar). By 2356 UTC, a small area of graupel is identified (Fig. 6b, marked by the arrow). This HCA-indicated graupel region expands in areal extent in the next two volume scans, more than doubling the number of graupel-identified pixels (Figs. 6c,d). By the end of the 0007 UTC volume scan, a flash is detected by the LMA (VHF sources indicated by the black markers in Fig. 6d). A strikingly similar evolution is observed for the 28 January case (Fig. 7). Again, the echo is dominated by dry snow aggregates initially (Fig. 7a). By the 2131:03 UTC scan (Fig. 7b), a few contiguous pixels of graupel are identified. This region expands considerably in the next two scans (Figs. 7c,d). The first LMA sources occurred shortly after 2142 UTC (Fig. 7d).

Such a rapid appearance and expansion of graupel identifications in both cases is suggestive of convective activity and considerable ongoing riming within a region of snow aggregates and ice crystals, which are precursors to electrification. In most of the cases, the operational HCA output classified regions of graupel prior to the first LMA-indicated flash. Most often this change in classification from snow aggregates to graupel is related to the increase in \( Z_H \) values over those expected for dry snow aggregates; in other words, more weight is assigned to the graupel category as \( Z_H \) increases to over 35–40 dBZ.

From an operational perspective, it appears that a sudden appearance or expansion in the areas classified as graupel should warrant more attention, as conditions are favorable for the development of electrification and possible lightning initiation, as well as potentially heavy snowfall. Crowe et al. (2006) explore correlations between heavy snowfall and thundersnow occurrences using METAR. They report that, if a storm is capable of producing thundersnow, then there is an enhanced likelihood of that system producing heavy snowfall.

<table>
<thead>
<tr>
<th>Date</th>
<th>Cell No.</th>
<th>Total flashes</th>
<th>IC (CG) flashes</th>
<th>LMA detections</th>
<th>CONUS detections</th>
<th>Max flash rate in 10-min window (min⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>25 Oct 2012</td>
<td>1</td>
<td>10</td>
<td>3 (7)</td>
<td>6</td>
<td>4–10</td>
<td>0.4</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>2</td>
<td>1</td>
<td>0 (1)</td>
<td>1</td>
<td>0</td>
<td>0.1</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>3</td>
<td>4</td>
<td>3 (1)</td>
<td>4</td>
<td>0–4</td>
<td>0.5</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>4</td>
<td>1</td>
<td>1 (0)</td>
<td>1</td>
<td>0</td>
<td>0.1</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>5</td>
<td>9</td>
<td>5 (4)</td>
<td>9</td>
<td>1–5</td>
<td>0.6</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>6</td>
<td>2</td>
<td>2 (0)</td>
<td>2</td>
<td>0–1</td>
<td>0.2</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>7</td>
<td>1</td>
<td>1 (0)</td>
<td>1</td>
<td>1</td>
<td>0.1</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>8</td>
<td>19</td>
<td>14 (5)</td>
<td>18</td>
<td>2–12</td>
<td>0.5</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>9</td>
<td>15</td>
<td>12 (3)</td>
<td>15</td>
<td>2–8</td>
<td>1.2</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>10</td>
<td>3</td>
<td>2 (1)</td>
<td>3</td>
<td>1–2</td>
<td>0.3</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>11</td>
<td>4</td>
<td>4 (0)</td>
<td>4</td>
<td>2–3</td>
<td>0.4</td>
</tr>
<tr>
<td>25 Oct 2012</td>
<td>12</td>
<td>1</td>
<td>1 (0)</td>
<td>1</td>
<td>0–1</td>
<td>0.1</td>
</tr>
<tr>
<td>11 Nov 2012</td>
<td>1</td>
<td>28</td>
<td>17 (11)</td>
<td>28</td>
<td>3–8</td>
<td>1.3</td>
</tr>
<tr>
<td>28 Jan 2013</td>
<td>1</td>
<td>1</td>
<td>1 (0)</td>
<td>1</td>
<td>0</td>
<td>0.1</td>
</tr>
<tr>
<td>28 Jan 2013</td>
<td>2</td>
<td>3</td>
<td>1 (2)</td>
<td>3</td>
<td>1–2</td>
<td>0.2</td>
</tr>
<tr>
<td>9 Apr 2013</td>
<td>1</td>
<td>5</td>
<td>2 (3)</td>
<td>5</td>
<td>0–4</td>
<td>0.5</td>
</tr>
</tbody>
</table>
accumulations. The areal expansion of graupel described above for two of the thundersnow events implies a larger region of enhanced $Z_H$ values, which generally suggest larger surface precipitation rates. Further, riming adds mass to hydrometeors and increases their fall speeds, also increasing the ice mass flux to the surface. Given that the lightning-producing cells tended to have larger $Z_H$ (and subsequently larger precipitation rates) than the other cells that did not produce lightning, the association between enhanced precipitation rates and thundersnow supports the findings of Crowe et al. (2006).

However, the appearance of HCA-identified graupel itself is not a necessary or sufficient condition for lightning in the thundersnow cases presented herein. For example, in at least two of the cases (25 October and 11 November) there were other cells in which graupel was identified that did not produce lightning. The 11 November case is investigated in more detail in section 4c below. It is likely that ongoing riming and charging was occurring in these cells. However, the charging apparently was not sufficient to induce a lightning discharge.

In addition to such null cases (i.e., HCA-identified graupel but no lightning), there was one flash on 25 October 2012 that occurred in the absence of HCA-identified graupel. In fact, the flash occurred in a snowband that did not exhibit the typical convective appearance of the other lightning-producing cells. However, as discussed in detail in the next subsection, this case did exhibit a polarimetric signature that could have provided forecasters with information about electrification in the cloud.

b. 25 October 2012 depolarization streak

Here, we present data from the 25 October 2012 case, in which a snowband produced a flash (the northernmost...
flash in Fig. 3) in the absence of any obvious convective structure in the radar data. Despite not having a clear convective structure apparent in the radar data or large $Z_H$ values (>35 dBZ) that could be indicative of graupel (Fig. 8a), nor any areas of graupel identified in the HCA output (Fig. 8b), the 25 October snowband case did exhibit a polarimetric radar signature indicative of electrification, beginning at least one volume scan preceding the CG flash. Strong electric fields in clouds can alter the orientation of small ice crystals (e.g., Caylor and Chandrasekar 1996; Ryzhkov and Zrnić 2007). The cloud was sufficiently strongly electrified that these low-inertia ice crystals were oriented at angles off the principal polarization plane axes (i.e., not 0° or 90°). Radars that operate in a mode of simultaneous transmission and reception of horizontally H and vertically V polarized waves (such as the WSR-88Ds) produce an artifact when the beam propagates through canted ice crystals. The canted media lead to depolarization of the signal, resulting in radially oriented streaks of positive or negative $Z_{DR}$ (e.g., Scott et al. 2001; Ryzhkov and Zrnić 2007; Hubbert et al. 2010; Kumjian 2013c). Such a depolarization streak is evident in the 0.92°-elevation scan of $Z_{DR}$ from KCYS at 0035 UTC (Fig. 9b), about 3–4 min prior to the flash (Fig. 9c). Note that it is exceedingly unlikely for the negative $Z_{DR}$ streaks to be caused by differential attenuation in this case, as (i) the precipitation was entirely snow at this time; (ii) the WSR-88D operates at S band, at which specific differential attenuation for snow is negligibly small owing to the very small relative permittivity of ice particles; and (iii) maximum $Z_H$ values are rather low, indicating a lack of very large sizes or concentrations of particles necessary to produce attenuation at S band. The streak also moves in time (cf. Fig. 9), indicating that it is not caused by some sort of partial beam blockage by a ground-based target.

Figure 10 provides an average of radial traces of $Z_H$, $Z_{DR}$, and $\Phi_{DP}$ through the depolarization streak at 0039 UTC. [Note that $\Phi_{DP}$ is not currently displayed in the National Weather Service Advanced Weather Interactive
Processing System (AWIPS) software but is available in archived level-II data. Between range bins 100 and 175, both $Z_H$ and $\Phi_{DP}$ increase, whereas $Z_{DR}$ is relatively low but positive. This suggests particles are getting larger with increasing range, presumably larger aggregates and/or small graupel with intrinsically low $Z_{DR}$. Increasing $\Phi_{DP}$ indicates more pristine, nonspherical particles are present as well because fluffy snow aggregates or graupel largely are invisible to $\Phi_{DP}$. Thus, the combination of each polarimetric variable provides information about a mixture of particles being present simultaneously in the cloud. Interactions among these different hydrometeors is one of the possible mechanisms thought to play a role in charging (e.g., Pruppacher and Klett 1997; MacGorman and Rust 1998; Rakov and Uman 2003). Recent work (e.g., Dye et al. 2007; Dye and Willett 2007; Kuhlman et al. 2009; Tsenova et al. 2014) has also suggested that charging can occur in the absence of graupel or supercooled cloud droplets through ice–ice interactions, and could plausibly explain the electrification in this case.

Notably, $\Phi_{DP}$ values decrease from range bins 175 to 250 along the depolarization streak, whereas adjacent azimuths featured monotonic increases in $\Phi_{DP}$ with range (not shown). The decrease in $\Phi_{DP}$ suggests the presence of nonspherical particles with their maximum dimensions projected more in the vertical plane than the horizontal plane. Such an alignment is possible for conical graupel or small ice crystals aligned in an electric field. However, the fact that $Z_{DR}$ begins to decrease monotonically with range here suggests depolarization, which favors oriented ice crystals that are not perfectly vertically (or horizontally) oriented. The increase in $\Phi_{DP}$ farther downrange (range bins >250) suggests crystals with a more horizontal orientation (e.g., Ryzhkov and Zrnić 2007). Interestingly, the depolarization streaks persisted for another ∼15 min after the flash (cf. Figs. 9d–f), indicating that the cloud was still electrified, albeit insufficiently for a lightning discharge. This example demonstrates that depolarization streaks in $Z_{DR}$ can be useful indicators of electrification, though they do not necessarily indicate that a lightning strike is imminent. Note that many of the other cases of lightning-producing thundersnow storms in our study were isolated cells, so there was a general lack of downrange echoes in which to observe any depolarization streaks.

The vertical structure of the storms can also be informative. A vertical cross section along the azimuth

![Figure 7](image-url)
nearest to the flash (Fig. 11) shows enhanced vertical structure in $Z_H$ centered at a range of about 90 km, coincident with the location of the flash. Though $Z_H$ values in this area remain <30 dBZ throughout, its vertically extensive structure is suggestive of enhanced vertical motion. This is supported by weak midlevel convergence that is present in the Doppler velocity field (not shown).

Figure 12 reveals the three-dimensional structure of the 0039 UTC CG flash and its evolution. The LMA sources indicate the flash initiated below 2 km MSL (within a few hundred meters above ground level) and progressed southward. The second part of the flash is accompanied by an upward-moving leader to the north again, which also initiates very close to the ground. The initial low-level sources (gray markers, near the ground) are within 1–2 km of several wind turbines and a communications tower as observed on Google satellite imagery. Given the uncertainty in VHF source positions by the LMA at this range, it is plausible that this flash was triggered by a man-made object. This would explain the isolated nature of this flash, as well as the seeming lack of a convective appearance in the radar data. Man-made objects such as towers can serve to locally enhance the electric field, possibly to a level sufficient to cause a discharge. In fact, Rakov and Uman (2003) suggest that the large horizontal extent of low-level charge regions common in winter storms favor lightning triggered by such towers.

Similar observations of flashes in the absence of notable convective structures were made by Bech et al. (2013) in a thundersnow case over Catalonia, Spain. Those authors hypothesized the importance of tall telecommunication towers in leading to CG lightning flashes. Tower-initiated flashes have also been observed in northern Alabama winter cases (C. Schultz 2014, personal communication) and over tall buildings in Chicago, Illinois (Warner et al. 2014). Thus, tall towers may be sufficient to produce a discharge in otherwise submarginal conditions for thundersnow.

c. 10–11 November 2012: A tale of two cells

During 10–11 November 2012, there were numerous isolated convective cells that produced $Z_H$ values >35 dBZ, suggestive of graupel. However, only one of these cells produced lightning flashes. To investigate any possible microphysical differences between the cell that produced lightning and those that did not, we compare the polarimetric radar variables $Z_H$, $Z_{DR}$, and $K_{DP}$ within two contrasting cells: the lightning-producing cell between 0002 and 0106 UTC 11 November, and a cell between 2112 and 2211 UTC 10 November 2012. In both cases, the operational HCA output indicated graupel, suggesting ongoing riming within the cells. Additionally, large (up to 5-mm diameter) conical graupel was observed in the earlier cell near Boulder, Colorado (Fig. 13). Both cells existed in similar environments (the same region and within ~4 h of one another), produced $Z_H > 40$ dBZ and graupel, lasted for extended periods (>1 h), and remained within similar ranges of the radar. However, only the later cell produced lightning flashes. For these reasons, these two cells offer a good opportunity to look for subtle differences in their microphysical characteristics. We will hereafter refer to the lightning-producing cell as the active cell, and the cell that lacked lightning as the inactive cell.

To facilitate a comparison, data from sampling volumes subjectively identified as being associated with each cell are selected. The data from the PPI
sweeps in which the cell is observable are used. At each volume scan time, the data are binned as follows: from 0 to 50 dBZ in 2-dB increments for $Z_H$, from $-1$ to 5 dB in 0.5-dB increments for $Z_{DR}$, and from $-1^\circ$ to $2^\circ$ km$^{-1}$ in 0.25 km$^{-1}$ increments for $K_{DP}$. The frequency distribution is normalized by the maximum frequency at each time such that the peak for each time is equal to one.

Fig. 9. Consecutive fields of $Z_{DR}$ from the 0.92° PPI scan at (a) 0031, (b) 0035, (c) 0039, (d) 0043, (e) 0048, and (f) 0052 UTC 25 Oct 2012. The red arrow points to the location of the observed depolarization streaks. The pink markers show LMA sources from the flash that occurred at 0039:01 UTC.
The results are shown in Fig. 14. The active cell is shown in Fig. 14 (left), with asterisks above indicative of minutes in which lightning occurred. Note the jump in relative frequency of the larger $Z_H$ points ($>30 \text{ dBZ}$) at the second volume scan. This increase in $Z_H$ occurs prior to the first flashes in this case. Compared to the inactive cell, the active cell has very similar peaks in the normalized frequency distribution. However, the active cell has a much broader tail, with larger relative contributions from $Z_H > 40 \text{ dBZ}$. (Indeed, the maximum $Z_H$ in the active cell exceeded 50 dBZ.) This may suggest the presence of more, larger, and/or higher-density graupel in the active cell compared to the inactive cell. These observations all point to better particle growth conditions within the active cell, which may include more favorable vertical velocities, higher liquid water contents, more ice mass, or some combination of these factors.

The distributions of $Z_{DR}$ in both cells reveal a peak at about 0 dB throughout the 1-h time period chosen for analysis. Note that the base-10 logarithm of the frequency distribution is plotted in Fig. 14 in order to bring out subtle changes in the distributions’ tails. Subtle increases in the frequency of high-$Z_{DR}$ volumes are observed between 0010–0030 and 0040–0050 UTC in the active cell. However, in general, the inactive cell produces relatively more high-$Z_{DR}$ volumes. This may be a result of the higher-$Z_H$ values in the active cell: more reflective graupel with intrinsically lower $Z_{DR}$ could dominate the backscatter, weighting $Z_{DR}$ more heavily than in the inactive cell.

The distributions of $K_{DP}$, however, are not weighted by reflectivity, only responding to the anisotropic scatterers within the sampling volume. In winter storms, this means the $K_{DP}$ will provide a measure of the mass content of more pristine crystals (e.g., Ryzhkov et al. 1998; Kennedy and Rutledge 2011; Andrić et al. 2013; Schrom et al. 2015). The normalized $K_{DP}$ distribution in the active cell has an enhanced tail at the second volume scan, just prior to the onset of the first flashes. This is followed by a weaker enhancement that persists for the second half of the analysis period. In contrast, the inactive cell features very few higher-$K_{DP}$ volumes, particularly after the first few volume scans. (In fact, throughout most of the analysis period, there are no volumes with $K_{DP} > 0.5^\circ \text{km}^{-1}$.) This strongly suggests that the inactive cell contained less pristine ice mass than did the active cell. In other words, there were fewer
and/or smaller pristine ice crystals in the inactive cell compared to the active cell.

In summary, the active cell had a broader distribution of $Z_H$ values, including more high-$Z_H$ (>40 dBZ) volumes than did the inactive cell. Additionally, there were more high-$K_{DP}$ volumes in the active cell than the inactive cell. Taken together, this indicates that the active cell had larger, more numerous, and/or higher-density graupel as well as a larger mass content of pristine ice crystals than the inactive cell. This suggests that there may be observable microphysical differences between cells that produce lightning and those that do not.
in similar environments. Obviously, more cases are needed to determine the reliability of these subtle signals, as this analysis only considers two cells. Additionally, the inactive cell was somewhat farther from KFTG than the active cell throughout the analysis period, so the sampling volumes were somewhat larger.

Because the mixture of graupel and pristine ice is a key ingredient for electrification, the combined use of $Z_H$ and $K_{DP}$ offers some promise in quantifying this mixture. Use of C- or X-band polarimetric radars will be especially helpful, as $K_{DP}$ is inversely proportional to radar wavelength for hydrometeors like pristine ice crystals that are in the Rayleigh scattering regime. Thus, subtle differences in $K_{DP}$ will be more easily distinguishable with higher-frequency radars.

In addition to the polarimetric radar variables, other investigators have explored the use of radar-derived products for the detection of storms capable of producing lightning. Figure 15 shows two of these products: volume averaged height-integrated radar reflectivity (VAHIRR; Krider et al. 2006) and a proxy for ice mass (e.g., Carey and Rutledge 2000; Mosier et al. 2011). These are products derived from gridded radar volume scans that have been used in previous lightning forecasting studies. One should not ascribe too much importance to individual values, but rather the relative differences in the two cases are what is important. VAHIRR values in the active cell are larger than those of the inactive cell, which do not exceed ~25 dBZ km. The most striking contrast is found in ice mass, however. Whereas the ice mass in the inactive cell does not exceed about 3.5 g m$^{-3}$, values $>4$ g m$^{-3}$ are consistently found in the active cell throughout the period in which it is producing lightning. In fact, values in excess of 8 g m$^{-3}$ are found during a period of heightened lightning activity. The increased ice mass is consistent with the larger $K_{DP}$ values observed in this case.

We performed a similar comparison between the two cells using the level-III product called the enhanced echo-top height (not shown). There were no significant differences between the two cells, highlighting their similarity in storm depth. This also could mean that the vertical resolution of the WSR-88D volume scan is insufficient to observe such subtle differences.

These analyses offer promising clues of subtle differences between otherwise similar convective cells. However, a much larger dataset is required to determine the robustness of these radar-observed differences. Compositing or averaging over a large number of storms may help reveal reliable bulk microphysical differences between convective snowstorms that produce lightning and those that do not.

5. Discussion and conclusions

The four thundersnow events presented in this study display common features in their synoptic-scale setting, thermodynamic environment, radar presentation, and electrical properties. Each case featured a large-scale environment conducive for ascent, with low-level mesoscale features that provided upslope flow. In particular, the Palmer Divide was found to be a preferred location for electrically active snowstorms in Colorado. In each case, the surface temperature was just under 0°C.

With regard to the questions asked in the introduction, we come to the following conclusions:

(i) In general, lightning from the four thundersnow days was detected by both the CONUS and LMA networks. Of the six cells without CG flashes, the CONUS network detected IC flashes in four of them. The two cases that were not detected by the CONUS network were only weakly electrically active, producing only one in-cloud flash. The LMA detected flashes in all 16 cells, though individual flashes were missed when the cells were at the edge of the network’s detectability. The relatively prolific number of thundersnow storms in the 2012/13 snow season in northern Colorado strongly suggests that thundersnow is more common than previously reported in published climatologies that used different datasets (i.e., no combination of LMA and CONUS networks). LMA networks allow for the detection of marginal events with a single or a few IC flashes that generally do not get detected as efficiently by CONUS networks. Future research should investigate other LMA networks in different parts of the world to verify if, in
fact, thundersnow events indeed are more common than previously thought.

(ii) Many of the flash events were associated with localized high-$Z_H$, low-$Z_{DR}$ regions (i.e., presumably convective regions and graupel production). The sudden appearance and expansion of radar gates classified as graupel preceded most of the flashes in these cells. Thus, such a signature in the operational HCA should warrant more attention for the possibility of lightning production, aviation hazards, and heavy snowfall.

In contrast, we documented an isolated flash in regions of presumably snow aggregates. It is clear why regions of radar-inferred graupel production would be favorable for electrification and possible lightning initiation; however, isolated flashes in snowbands are more puzzling. Maximum $Z_H$ values did not much exceed 30 dBZ. Though some degree of riming cannot be ruled out, noninductive charging may have still contributed to the production of a strong enough electric field that allowed for triggered lightning without the presence (or very low concentrations) of supercooled liquid water (e.g., Dye et al. 2007; Dye and Willett 2007). As expected, this rarer and more puzzling case was isolated. Presumably, the electric field was stronger in the more vigorous convective elements during the same event. Despite being only weakly electrified, though, the polarimetric radar data did display a depolarization streak signature prior to the lightning flash. Thus, the presence of depolarization streaks in winter storms should alert forecasters that electric fields are sufficiently strong to affect the orientation of ice crystals (though we emphasize that by itself this signature does not guarantee that a lightning flash
is imminent). As described above, such strong electrification can pose aviation hazards, such as aircraft- or helicopter-triggered lightning (e.g., Mäkelä et al. 2013; Wilkinson et al. 2013).

Additionally, there were instances of graupel detected in the operational HCA in cells in which there was an absence of lightning. It is likely that charging was ongoing, just insufficient in magnitude for a lightning discharge. A comparison of such an inactive cell with a similar lightning-producing (active) cell from 11 November 2012 revealed subtle microphysical differences, with the inactive cell having lower $Z_H$ values and no enhanced $K_{DP}$ values compared to the active cell. In addition, the active cell had somewhat larger VAHIRR values and significantly larger ice mass values than the inactive cell. Enhanced echo tops showed no discernible differences between the two cases.

(iii) Compared to typical warm-season convective storms that may have from several to hundreds of flashes per minute, the thundersnow events studied herein had much lower flash rates (often <1 min$^{-1}$). This is not surprising given the lower-CAPE environments (and thus weaker maximum updraft velocities and cloud vertical extent). Flash-by-flash analysis based on LMA data suggests negative charge below an upper positive charge region in many of the cases. This is very similar to weakly electrified warm-season thunderstorm structures with intermediate amounts of supercooled liquid water. Thus, we suggest that thundersnow storms often are simply at the weaker end of the spectrum of ordinary thunderstorms, with an absence of any $<0^\circ$C temperatures in the lower troposphere.

Aside from being simply a meteorological curiosity, thundersnow events are scientifically advantageous to study because of their marginal nature. Electrification and lightning in deep moist convection are virtually guaranteed in most midlatitude continental storms, whereas thundersnow events are much more marginal.

Fig. 15. As in Fig. 14, but for derived quantities of (top) VAHIRR (dBZ km) and (bottom) ice mass (g m$^{-3}$). The ice mass frequency distributions are shown in logarithmic scale.
If some electrification threshold exists that may distinguish lightning-producing storms from those that do not produce lightning, thundersnow storms are certainly closer to that threshold than most deep moist convective storms. Electric field measurements inside thundersnow clouds may help better define this threshold, if it exists.

Acknowledgments. The National Center for Atmospheric Research (NCAR) is sponsored by the National Science Foundation (NSF). Funding for the first author comes from NSF Grant AGS-1143948. Partial support for the first author came from the NCAR Advanced Study Program. We thank Jim Dye and Matthias Steiner (NCAR) for valuable discussions on this research, and Paul Krehbiel and Bill Rison (New Mexico Institute of Mining and Technology) for the Colorado LMA lightning data. We also thank Earth Networks, Vaisala, and WSI for the use of their respective lightning data in our research. The three anonymous reviewers are thanked for their detailed, constructive criticisms and suggestions that greatly improved the manuscript.

REFERENCES


Unauthenticated | Downloaded 07/15/24 08:02 PM UTC


Straka, J. M., D. S. Zrnić, and A. V. Ryzhkov, 2000: Bulk hydrometeor classification and quantification using polarimetric...


