Observations of two sprite-producing storms in Colorado

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Abstract

Two sprite-producing thunderstorms were observed on 8 and 25 June 2012 in northeastern Colorado by a combination of low-light cameras, a lightning mapping array, polarimetric and Doppler radars, the National Lightning Detection Network, and charge moment change measurements. The 8 June event evolved from a tornadic hailstorm to a larger multicellular system that produced 21 observed positive sprites in 2 h. The majority of sprites occurred during a lull in convective strength, as measured by total flash rate, flash energy, and radar echo volume. Mean flash area spiked multiple times during this period; however, total flash rates still exceeded 60 min⁻¹, and portions of the storm featured a complex anomalous charge structure, with midlevel positive charge near −20°C. The storm produced predominantly positive cloud-to-ground lightning. All sprite-parent flashes occurred on the northeastern flank of the storm, where strong westerly upper level flow was consistent with advection of charged precipitation away from convection, providing a pathway for stratiform lightning. The 25 June event was another multicellular hailstorm with an anomalous charge structure that produced 26 positive sprites in less than 1 h. The sprites again occurred during a convective lull, with relatively weaker reflectivity and lower total flash rate but relatively larger mean flash area. However, all sprite parents occurred in or near convection and tapped charge layers in adjacent anvil cloud. The results demonstrate the sprite production by convective ground strokes in anomalously charged storms and also indicate that sprite production and convective vigor are inversely related in mature storms.

1. Introduction

After more than 25 years of research on sprites, stemming from their fortuitous capture on video by Franz et al. [1990], a general understanding of how they are related to thunderstorm structure and evolution has begun to coalesce. Sprites are most likely to be associated with powerful positive cloud-to-ground (+CG) lightning strokes [Boccippio et al., 1995; Bell et al., 1998; Williams, 1998], especially those associated with large charge moment changes (CMCs) [Huang et al., 1999; Hu et al., 2002; Cummer and Lyons, 2005; Hiraki and Fukunishi, 2006; Qin et al., 2012]. Production of such strokes is favored to occur in stratiform precipitation regions, such as might be found in mesoscale convective systems (MCSs) [Lyons, 1996; Lyons et al., 2003; Lyons, 2006; Williams and Yair, 2006]. Research has shown that sprite production appears to be related to the development and intensification of the stratiform region [Lyons, 1996, 2006; Williams and Yair, 2006; Soula et al., 2009; Lang et al., 2010]; however, it also appears that a robust convective line or region is required since many sprite-parent CGs originate there, followed by propagation into the stratiform precipitation [Lang et al., 2010; van der Velde et al., 2010, 2014; Soula et al., 2015]. The charge layers involving these flashes appear to be fed by charged ice particles that detrain from the convective line, which then descend and further develop in the weak mesoscale updrafts typically found inside stratiform precipitation regions [Carey et al., 2005; Ely et al., 2008].

However, there are notable exceptions to this scenario. Some sprites are caused by −CGs [Barrington-Leigh et al., 1999, 2001; Taylor et al., 2008; Soula et al., 2009; Li et al., 2012; Lu et al., 2012], and these appear to not be strictly tied to stratiform precipitation; they often occur in deep convection [Lang et al., 2013]. Some sprites may even be caused by purely intracloud (IC) lightning [Neubert et al., 2005], and in any event it has been established that sprite evolution is strongly related to the development of in-cloud portions of the parent flash [van der Velde et al., 2006, 2010, 2014; Lu et al., 2013]. Furthermore, positive sprites can occur over +CGs in convective regions, and there is some limited evidence that many of these are related to the
presence of anomalous charge structures [Lyons et al., 2008; Lang et al., 2010]; that is, the presence of positive charge layers near midlevels (e.g., $-20^\circ\text{C}$) rather than the more common negative charge [Lang et al., 2004; Rust et al., 2005; Wiens et al., 2005; Bruning et al., 2014; Fuchs et al., 2015]. Moreover, it is not strictly the case that an MCS is required for sprites; they have been observed over smaller (i.e., sub-MCS) storms [Lyons et al., 2008; São Sabbas et al., 2009; van der Velde et al., 2014], including winter storms [Takahashi et al., 2003; Suzuki et al., 2011].

These exceptions demonstrate the limits of our current understanding concerning the meteorology of sprites. Moreover, studies of most sprite-producing storms often lack a comprehensive data set including detailed information on storm kinematics and microphysics, such as can be provided by polarimetric and multi-Doppler radar networks. In addition, while increasing emphasis has been placed on three-dimensional (3-D) mapping of individual sprite-parent CGs [Lyons et al., 2003; Lang et al., 2010, 2011; Lu et al., 2013; van der Velde et al., 2014], relatively little focus has been placed on the evolution of total lightning relative to sprite production, nor has storm tracking been implemented to characterize the full temporal evolution of the convection in sprite-producing systems. This would be especially helpful with smaller, sub-MCS storms where ambiguity in convective state can be reduced due to the presence of fewer cells. Overall, this information deficit prevents the development of a fully realized theory of sprite production relative to thunderstorm structure and evolution.

In order to make further progress on these questions, this study presents observations of two sprite-producing storms that occurred in Colorado during the Deep Convective Clouds and Chemistry (DC3) field campaign [Barth et al., 2014]. The first occurred on 8 June 2012 and produced 21 observed sprites over a 2 h period (0500–0700 UTC). The second storm, on 25 June, produced 26 observed sprites over a 1 h period (0415–0515 UTC). These storms were observed by a network of sprite-observing camera systems, multiple Doppler radars (including one polarimetric radar), a 3-D lightning mapping array (LMA), and national CG- and CMC-observing networks. This enabled one of the most comprehensive examinations of the relationships between sprite production and thunderstorm evolution to date. These storms featured anomalous charge structures, and the observations clearly indicated the influence of convective strength on sprite production in mature storms. The data also will show how the microphysical, kinematic, and electrical evolution of these storms strongly influenced the structure, location, and timing of sprite-parent CGs.

### 2. Data and Methodology

#### 2.1. Sprite Cameras

Optical observations of sprites were made on 8 and 25 June 2012 from multiple locations via a variety of cameras. The main camera type that was used in this study was the Watec 902H, which is reviewed extensively in Lu et al. [2013]. These are steerable, low-light video cameras. They are monochrome and record either continuously or via an automated trigger on an optical transient. Time stamping is provided by a connected Global Positioning System antenna. For the 8 and 25 June cases, the cameras were steered manually and recorded observations continuously. Sprites were identified during later manual analysis of the video. A modified digital single-lens reflex (DSLR) camera with the near-infrared filter removed was used to perform automatic, long-exposure color photography of sprites and other TLEs.

On 8 June 2012, Watec cameras near Lamy, New Mexico, Bennett, Colorado, and on Cerro Pelado mountain, New Mexico, operated and captured sprites. On 25 June 2012, a DSLR and Watec at Lamy and a Watec in Bennett were available. Watecs captured all but one sprite on that day, and the DSLR captured the other sprite. See Figure 1 in Lu et al. [2013] for a map of the sprite-observing network. Phantom v7.3 high-speed imagers were available in the network, but not used in this study.

Time-stamped sprite video and imagery were matched to parent CG lightning (data set described later) via comparison of times and locations (i.e., azimuth from a given camera) of occurrence, following standard practices [e.g., Lu et al., 2013]. Timing for the DSLR-only sprite was matched to a +CG in the camera’s field of view that occurred during the 3 s exposure window. Sprite durations were available via manual video analysis on 8 June, but this analysis was not done for 25 June. This study mainly used the sprite observations to establish time and approximate location of occurrence, relative to storm evolution.
2.2. Radars

There were two research radars and two operational radars of most relevance to this study. The Colorado State University-University of Chicago-Illinois State Water Survey (CSU-CHILL) radar is a dual-wavelength, S/X-band, polarimetric Doppler radar with an 8.5 m, dual-offset, Gregorian antenna [Bringi et al., 2011; Junyent et al., 2015], located near Greeley, Colorado. CSU-Pawnee was an S-band Doppler radar situated 47.7 km NW of CSU-CHILL. Together, they formed a dual-Doppler network that has been frequently used to study combined kinematic and microphysical characteristics of storms [e.g., Lang and Rutledge, 2002; Dolan and Rutledge, 2007; Basarab et al., 2015]. Additional coverage was provided by U.S. National Weather Service Doppler radars near Denver, Colorado (KFTG), and Cheyenne, Wyoming (KCYS). These S-band Doppler radars can often contribute to multi-Doppler syntheses because of their close proximity to CSU-CHILL and CSU-Pawnee [Dolan and Rutledge, 2007].

The CSU radars were operated from ~2100 UTC on 7 June 2012 through ~0440 UTC on 8 June. A major focus of the scanning was coordinated plan position indicator (PPI) sectors to support multi-Doppler syntheses. In addition, CSU-CHILL performed frequent range-height indicator (RHI) scan volumes during this time period. On 24–25 June 2012, CSU-CHILL radar was the only research radar operated, and it primarily scanned storms in RHI mode from ~2030 UTC to ~0450 UTC. When performing RHIs, CSU-CHILL would focus on volumetric scanning with 1–6 min updates, as opposed to high temporal resolution updates along a single azimuth. On both days KFTG and KCYS scanned full 360° PPI volumes, from 0.5° up to 19° elevation.

Multi-Doppler synthesis was performed for the 8 June 2012 case. Prior to gridding and analysis, velocity data from all radars were first manually dealiased, and nonmeteorological artifacts were removed using a mixture of thresholds on polarimetric variables (e.g., correlation coefficient) and manual editing. Then the data were interpolated to a common Cartesian grid using the National Center for Atmospheric Research (NCAR) Sorted Position Radar Interpolator (SPRINT) software [Mohr and Vaughn, 1979; Miller et al., 1986]. Resolution of the grid was 1 km in all dimensions, and the vertical coordinate stretched from 2.5 to 18.5 km above mean sea level (msl; CSU-CHILL is at 1.43 km msl). Wind analyses were performed on the merged data set using the Custom Editing and Display of Reduced Information in Cartesian Space (CEDRIC) software [Mohr and Miller, 1983; Mohr et al., 1986]. In this study we present only the off-baseline horizontal wind results in and near the stratiform precipitation, as the primary convective core was along the CHILL-Pawnee baseline during our focus period for wind analysis (04–05 UTC). Radars were used in an analysis if their scans started within 3 min of one another, and analyses were available roughly every 6 min during 0200–0440 UTC.

Reflectivity data from all available National Weather Service (NWS) radars are also regularly included in 3-D national mosaics [Zhang et al., 2011]. During 2012, these mosaics (now called Multi-Radar/Multi-Sensor or MRMS) were available every 5 min, at 0.01° latitude/longitude resolution, and with a vertical coordinate that stretched from 0.25 km to 18.0 km [Lang et al., 2014]. Vertical resolution varied from 0.25 km near the surface to 2 km near the top. For the domain of this study, KFTG and KCYS were the primary radars that contributed to these mosaics in the study domain. These mosaics provided volumetric information on all storms in the region, since the research radars often were narrowly focused in their scanning.

2.3. Colorado Lightning Mapping Array

The Colorado Lightning Mapping Array (COLMA) is a very high frequency (VHF)-based network that maps flashes in three dimensions plus time [Rison et al., 1999; Lang et al., 2014; Fuchs et al., 2015]. During June 2012 the COLMA consisted of 15 separate stations, spread throughout northeastern Colorado. During most of the analysis periods, the storms in this study were well within the 100 km distance of network center that is generally considered optimal for 3-D analysis [MacGorman et al., 2008]. As is standard practice, a minimum of seven station detections and a chi-square error of 1 or less were required for a location solution to be considered valid [e.g., Lang and Rutledge, 2011]. Individual VHF sources were clustered into flashes via the algorithm developed by Fuchs et al. [2015], which required no more than 3 km distance and 0.15 s time between individual sources, for them to be considered part of the same flash. Flash energy was computed following the methodology of Bruning and MacGorman [2013]. For more information about the performance of COLMA during June 2012, see Lang et al. [2014], Basarab et al. [2015], and Fuchs et al. [2015].
2.4. National Lightning Detection Network

The National Lightning Detection Network (NLDN) is a well-known lightning network that in 2012 primarily detected cloud-to-ground (CG) lightning with a detection efficiency at the flash level of 95% or better and a location accuracy better than 0.5 km [Cummins and Murphy, 2009]. Flash-level data were used in this study. In order to filter out possible ICs, a 15 kA threshold was already applied to the positive CG flashes before receipt of the data [Biagi et al., 2007]. Prior to analysis we also applied the same magnitude threshold to the negative CG data. This was because in anomalously charged storms, the negative CG population is disproportionately filled with low-peak current flashes, and these are likely misclassified ICs [Montanyà et al., 2009; Lang et al., 2015]. This was also observed in the present study.

2.5. Charge Moment Change Network

In June 2012, the Charge Moment Change Network (CMCN) consisted of two stations, one near Fort Collins, Colorado, and one near Duke University Forest in North Carolina [Cummer et al., 2013]. The stations consist of orthogonal induction magnetic field sensors that are sensitive to signals from 2 Hz to 25 kHz or from extremely low frequency (ELF) to very low frequency (VLF). Via specialized processing detailed in Cummer et al. [2013] and Beavis et al. [2014], the impulse charge moment change (iCMC) from individual CG strokes can be determined, with a factor of 1.5 accuracy. The iCMC is the charge moment change within the first 2 ms of the return stroke. For this study we focused only on the strokes of either polarity with iCMC > 75 C km. Geolocation of these large-iCMC strokes was done via comparison with the NLDN, following the methodology described in Cummer et al. [2013].

2.6. Soundings

Regular National Weather Service (NWS) soundings from the Denver station (No. 72469) and the North Platte, Nebraska, station (No. 72562) were analyzed to determine environmental characteristics near the storms studied in this paper. Parameters such as convective available potential energy (CAPE), convective inhibition (CIN), lifting condensation level (LCL), level of free convection (LFC), equilibrium level (EL), shear, and storm-relative helicity (SRH) were analyzed by considering the most unstable parcel (for the variables where that is relevant; e.g., CAPE). The analysis package used for these calculations was the Sounding/Hodograph Analysis and Research Program in Python (SHARPpy) [Halbert et al., 2015]. Mobile soundings were available during the DC3 project, but none were launched on 8 or 25 June.

2.7. Colorado State University (CSU) Lightning, Environment, Aerosol, and Radar Analysis Framework

The CSU Lightning, Environment, Aerosol, and Radar (CLEAR) analysis framework provides a useful means of integrating lightning and radar data for this study. While the details of the framework are covered in Lang and Rutledge [2011] and Fuchs et al. [2015], a brief synopsis of how CLEAR was used for this study is included here. We used the Fuchs et al. [2015] CLEAR-processed data set, which merged MRMS radar features with COLMA and NLDN. Radar features were identified using two reflectivity thresholds applied to a 2-D composite reflectivity field. As in Fuchs et al. [2015], a minimum area threshold of 20 km² was applied to the 30 dBZ contour, and a threshold of 13 km² was applied to the 40 dBZ contour, when identifying radar features. This preferentially selected more intense, mature convective cells. By using these thresholds, CLEAR did not include distinct stratiform echo regions when producing time series analyses. Since this study was focused on the convective behavior of multicellular sprite-producing storms, these thresholds are considered appropriate.

LMA flashes were linked to a radar feature if they initiated within it or if not, then to the closest radar feature within 10 km of the initiation point. If an NLDN flash was located within the horizontal area of a radar feature, it was matched to that feature. If it was located outside a feature, then it was assigned to the closest feature within 10 km distance. If it occurred beyond 10 km, it was not linked to any feature. CMCN-detected large-iCMC strokes were associated with radar features identically to how NLDN flashes were. Sprites were associated with the radar features that were determined by CLEAR to have produced their NLDN-detected parent +CGs.

Once linked to individual radar features, which in turn were grouped together in time to form a track that represented the evolution of the main storm producing sprite-parent flashes, all lightning and radar observations were analyzed to produce time series for these storms. CLEAR is capable of automated feature tracking.
to enable this time series development [Fuchs et al., 2015]. However, in order to ensure maximum accuracy, and because we were only focused on two cases, the storms in this study were tracked manually by noting the CLEAR identification numbers for each cell in a radar volume and creating a list of all the cells associated with the particular storm at a particular time. Examples of how this worked are shown in Figure 1 for the 8 June 2012 storm and in Figure 2 for the 25 June 2012 storm. For example, at 0300 UTC on 8 June 2012 (Figure 1a) cell number 2732 comprised the main storm. By 0600 UTC, the CLEAR-assigned cell number for the main storm had changed and was now 2956. The manual tracking had to take account of these changes when they occurred, and assigning cells to the sprite storms was done via visual inspection and animations using plots like these.

At times, the storm of interest consisted of more than one CLEAR-identified radar feature in a particular volume. When this occurred, metrics for each cell were included in time series analyses. Subjective decisions about which cells to include took into account whether cells eventually merged into the main storm or split from it. If cells merged into the main storm, they were kept in the analysis prior to the merger. However, when splits occurred, the smaller cell(s) that split away were not kept if they moved away from the main convective envelope associated with the storm. Generally, it was relatively straightforward to identify the cell(s) responsible for the main storm of interest, and sensitivity studies found that including or excluding nearby smaller and shorter-lived cells did not make a significant difference on the time series results.
3. Results

3.1. The 8 June 2012 Storm

The 0000 UTC Denver sounding on 8 June is shown in Figure 3a (several derived sounding parameters are listed in Table 1). The atmosphere was particularly unstable on this day, with the most unstable parcel method yielding 2662 J kg$^{-1}$ of CAPE. In addition, significant shear and SRH were reported. This led to the development of a severe tornadic, multicellular hailstorm. Prior to the sprite analysis period (which was roughly 0300–0700 UTC; Figure 1), cells in Weld County, Colorado (the primary focus area of the DC3 campaign in Colorado), had produced multiple large hail reports (up to ~6 cm in diameter) as well as a tornado report. There were also wind gust reports up to ~40 m s$^{-1}$ [Storm Prediction Center, 2012a]. Even during the sprite analysis time, the storm was still strong, with multiple hail reports up to ~4.5 cm before 0500 UTC as well as severe-scale winds. After passing near the CSU-CHILL radar between 0430 and 0500 UTC, one of the authors (B. Fuchs) noted significant hail accumulations leading to impassable roads in the storm’s wake.

Around 0300 UTC (Figure 1a), the storm consisted of multiple discrete cells and was northwest of the Pawnee radar. These cells gradually coalesced until they formed a contiguous convective region by 0500 UTC (Figure 1c). As this occurred, the storm became larger and more intense in terms of reflectivity until just before 0430 UTC, when it began a long-term decline in size and intensity through 0600 UTC (Figure 4a). This weakening was notably interrupted by a brief convective burst around 0450–0500 UTC, which was observed in total flash rate and total flash energy (Figure 4b), and coincided with an upward shift in LMA

![Figure 2. Composite reflectivity from the MRMS mosaic valid at several times on 25 June 2012. Individual CLEAR-tracked cells are indicated by the numbers, with the dashed lines showing the convex hulls associated with each cell and the grey curves showing the fitted ellipses to each hull. Cells 3670, 3680, 3696, and 3707 are the primary cells associated with the sprite-producing storm. (a) 0300 UTC. (b) 0400 UTC. (c) 0500 UTC. (d) 0600 UTC.]
source density altitude (Figure 4d). Immediately after this burst subsided, the storm began producing most of its observed sprites, roughly between 0500 and 0610 UTC (Figure 4b). Spikes in mean flash area occurred within this time frame as well.

Starting during 0550–0600 UTC, the storm intensified again, in terms of reflectivity structure (especially 40 and 50 dBZ volumes; Figure 4a), total flash rate, and flash energy (Figure 4b). With one exception (near 0700 UTC), the sprite production ended when the peak of the reintensification was achieved. The reintensification also featured enhanced source activity at lower altitudes (Figure 4d). Even at its lowest value, however, the total flash rate in this storm was ~60 min⁻¹. Visually, from a position west of the storm in Fort Collins, Colorado, the first author (T. Lang) observed essentially continuous lightning as the storm moved north to south through Weld County.

Throughout this period the storm was producing very low CG flash rates, rarely above 10 per 5 min period, marking it as a high IC:CG ratio storm common to the Eastern Plains of Colorado [Lang et al., 2000; Boccippio et al., 2001; Lang and Rutledge, 2002]. The majority of these ground flashes consisted of +CGs, and the relative fraction of +CGs was highest during the most intense periods of the storm (roughly 0330–0430 and 0600–0700 UTC; Figure 4c). If no 15 kA filter was applied, then −CGs were 5–10 times more frequent than +CGs, and

Table 1. Derived Parameters From the Denver, CO, Sounding Most Relevant for the Two Colorado Sprite Cases

<table>
<thead>
<tr>
<th>Variable</th>
<th>6/8/2012 0000 UTC</th>
<th>6/25/2012 0000 UTC</th>
</tr>
</thead>
<tbody>
<tr>
<td>LCL (m msl)</td>
<td>3,502</td>
<td>6,239</td>
</tr>
<tr>
<td>LFC (m msl)</td>
<td>3,785</td>
<td>6,401</td>
</tr>
<tr>
<td>Melting Level (m msl)</td>
<td>4,278</td>
<td>4,267</td>
</tr>
<tr>
<td>EL (m msl)</td>
<td>14,021</td>
<td>11,417</td>
</tr>
<tr>
<td>CAPE (J kg⁻¹)</td>
<td>2,662</td>
<td>190</td>
</tr>
<tr>
<td>CIN (J kg⁻¹)</td>
<td>−27</td>
<td>−0.3</td>
</tr>
<tr>
<td>Lifted Index</td>
<td>−6.7</td>
<td>0.7</td>
</tr>
<tr>
<td>0–6 km shear (m s⁻¹)</td>
<td>15.4</td>
<td>13.2</td>
</tr>
<tr>
<td>0–3 km SRH (m² s⁻²)</td>
<td>64.5</td>
<td>33.4</td>
</tr>
</tbody>
</table>

Station elevation is 1625 m msl.
Figure 4. (a) Time series of echo volumes associated with various MRMS reflectivity thresholds, for the 8 June 2012 storm. (b) Time series of LMA-derived flash rate, mean flash area, and total flash energy for this storm. Also indicated are the times of the observed sprites. (c) Time series of NLDN-derived CG flash rates in this storm, separated by polarity, as well as the times of strokes that produced a positive iCMC above 75 C km. A 15 kA peak current threshold has been applied to the NLDN data. (d) Time-height plot of LMA source density in the 8 June 2012 storm.
largely tracked LMA total flash rates (not shown), similar to the El Reno storm studied by Lang et al. [2015]. This behavior suggested that these low-peak current +CGs were likely misclassified ICs.

Large iCMCs (all positive) were not recorded until after 0600 UTC. This is because power and Internet outages during the storm prevented the reporting of the CMCN sensor at Yucca Ridge Field Station (YRFS; near Fort Collins, Colorado) to the national iCMC data set during 03–06 UTC. Manual analysis of the CMCN data for select sprite-parent strokes during 05–06 UTC indicated that large-CMC strokes were occurring at this time, however. Based on that analysis, at least four large-iCMC strokes occurred during 0513–0525 UTC [Lyons et al., 2012]. This is likely a very significant underestimate of the pre-0600 UTC large-iCMC rate, however. Values for large-iCMC strokes after 0600 UTC ranged between 75 C km and 273 C km.

Overall, the lightning behavior was indicative of an anomalously electrified storm [Bruning et al., 2014]. This inference was based on a number of observations. First was the high percentage of positive CGs (77.6% over the analysis period), as well as the fact that only positive large-iCMC strokes were observed. Second, there was a midlevel (6–8 km or approximately −10 to −20°C based on the Denver sounding) LMA source density maximum during much of the analysis period (Figure 4d), which has been linked to the presence of a midlevel positive charge [Lang and MacGorman, 2011; Fuchs et al., 2015]. Finally, manual analysis (following the methodology of Wiens et al., 2005) of select periods of COLMA data (not shown) indicated that small and inverted IC discharges (between midlevel positive and upper level negative charge) were frequently occurring near the strongest convection.

However, the charge structure in this storm featured substantial temporal and spatial heterogeneity. For example, after 0430 UTC the LMA activity shifted upward and a midlevel maximum did not return until the storm reintensified after 06 UTC (Figure 4d). In addition, between 0430 and 0530 UTC the relative fraction of +CG lightning fell (Figure 4c). Moreover, lightning in the region where sprite-parent flashes occurred did not typically feature inverted characteristics.

Sprite-parent flashes, without exception, occurred on the downshear (eastern) side of the convection, which after 0500 UTC started to exhibit a quasi-linear north-south structure. These larger flashes tended to initiate more than 10 km away from the most intense portion of the convection to the southwest, which featured frequent, small, and often inverted IC flashes. This pattern is reminiscent of the flash-size spectra results of Bruning and MacGorman [2013]. Figure 5 shows a very typical example of a sprite-parent flash in this storm.

Near the convection, the flash had the classic “I” structure common in normal bilevel IC flashes (Figure 5b) [Rison et al., 1999], with inferred positive charge near 9 km msl (roughly −40°C in Figure 3a) and inferred negative charge near 6 km msl (roughly −10°C). The flash then propagated eastward along the downward sloping upper positive charge layer, behavior that is very typical of normal sprite parents [Lang et al., 2010]. During the sprite there was LMA-mapped activity within this positive charge layer. The parent +CG came to ground just downshear of the convective line, while the majority of the flash activity occurred well east, within the stratiform precipitation. As the storm was traveling from the north-northwest to the south-southeast, this was technically the forward flank of the storm. This is a departure from common sprite-parent behavior in leading-line/trailing-stratiform MCSs, where stratiform lightning typically travels rearward and thus opposite of the direction of motion [Lang et al., 2010]. The closest analog to the 8 June case would be a cross between a parallel-stratiform and a leading-stratiform MCS [Parker et al., 2001].

Interestingly, late in the flash’s lifetime there is evidence for a negative charge region above the positive region, with recoil activity near 12 km msl in the stratiform region (Figures 5b and 5d). Inferred high-altitude negative charge also was commonly observed closer to convection in some (nonsprite-parent) flashes (not shown), so this flash may have revealed that this layer also advected eastward. Figure 6a shows the time-range development of this flash, following the analysis style of van der Velde et al. [2014]. Here the bidirectional leader development during much of the flash is readily apparent. Multiple negative leaders (the steep linear features in Figure 6a, with speeds −10^5 m s^-1) occur in succession, both before and after the sprite-parent +CG. This negative development often coincides with slower-moving positive leaders (−10^4 m s^-1), which are revealed by the less sloped recoil activity that is located closer to the flash origin. The activity associated with the upper level negative charge layer is readily apparent after 1 s and largely consists of two high-altitude positive leaders separated in time by a lower altitude negative leader. Additional sprite-parent flash animations can be found in the supporting information, and some of these flashes also indicated eastward advection of the lower negative charge region into the stratiform region.
The CSU-CHILL was operational until about 0440 UTC, prior to the first sprite observations. In addition, the Pawnee radar shut down around the same time. However, prior to this CSU-CHILL was able to contribute to multi-Doppler syntheses, via combination with the Pawnee, KFTG, and KCYS radars. Figure 7 shows the last synthesis (valid 0423 UTC) before CSU-CHILL terminated operations. At this time much of the storm was along the radar baselines and was very nearly on top of CSU-CHILL. Thus, we do not attempt to consider vertical wind. However, the horizontal winds were still useful away from the baselines. They indicated strong westerly to northwesterly flow (Figure 7a), with increasing vertical shear in the zonal component of the wind (Figure 7b), particularly on the eastern flank of the storm. This suggests that the sprite-parent flashes were responding to the eastward motion of the charged hydrometeors detraining from near the top of the storm, enabled by the strong vertical shear relative to storm motion (Figure 3a and Table 1).

3.2. The 25 June 2012 Storm

The 0000 UTC Denver sounding from 25 June 2012 is shown in Figure 3b, with derived parameters listed in Table 1. The atmosphere on this day was significantly different than it was on 8 June. It was much drier, with the lapse rate approximating a dry adiabat up to the LCL, which was nearly 2800 m higher than it was on 8 June. This led to the melting level being below the LCL and thus zero expected warm-cloud depth. CAPE was greatly reduced, only 190 J kg⁻¹, and the equilibrium level was also lower than 8 June. While shear

Figure 5. Typical example of a sprite-parent +CG during the 8 June 2012 storm. The flash occurred at the indicated time, with radar data coming from 0505 UTC MRMS mosaic. (a) Plan view with composite radar reflectivity (filled contours) and lightning data. The pink dots show the LMA sources from 50 ms prior to the sprite-parent +CG through the end of the sprite, and the black dots are all other LMA sources during the flash. The red dashed crosshairs indicate the vertical cross sections in Figures 5b and 5c. (b) Vertical cross section through the indicated constant latitude. (c) Vertical cross section through the indicated constant longitude. (d) Time-height plot of the flash.
was comparable to 8 June, the SRH was reduced by about 50%. Since the main storm on this day occurred near the Nebraska border, the North Platte sounding at 0000 UTC also was checked. Though this station was much farther from the storm than Denver (~370 km versus ~120 km), its sounding was moister and significantly more unstable, with the LCL reduced to 2890 m msl and CAPE increased to 3970 J kg$^{-1}$. This gradient suggested that the storm likely had access to a more favorable convective environment than would be expected based on the Denver sounding alone. In fact, an NWS analysis at 2336 UTC on 24 June indicated a surface boundary and CAPE gradient in northeastern CO, near where this storm developed [Storm Prediction Center, 2012b]. This storm occurred during the High Park fire, which flared up and was reported on 9 June and was not fully contained until July [Lang et al., 2014]. Thus, the region was affected by significant amounts of smoke aerosol [Barth et al., 2014]. Photographs by one of this study’s authors (W. Lyons) indicated that several clouds were ingesting smoke aerosol on this day.

An intense storm developed during the local afternoon hours, northeast of CSU-CHILL near where Wyoming, Nebraska, and Colorado meet. A westward moving outflow boundary from this convection merged with diurnally forced storms moving off the Rocky Mountain foothills and formed a storm (Figure 2) that persisted into the evening as it moved north-northeastward from Colorado into Wyoming. The period for which this storm was analyzed was 0300–0700 UTC. As it evolved, between 0400 and 0500 UTC it transitioned from a small, quasi-linear collection of cells to a more unicellular system (Figure 2).

Interestingly, this consolidation process was coincident with a significant overall decrease in the convective strength, as indicated by reflectivity volumes (Figure 8a). This decrease in strength was also indicated by a long-lived secular decline in both flash rate and flash energy (Figure 8b) from their relative maxima around 0315 UTC. However, after 0400 UTC, mean flash area began to increase even as flash rate declined toward a short-lived relative minimum at 0415 UTC. The first two sprites were observed at 04:18:24 UTC and then

Figure 6. Horizontal distance versus time for COLMA sources observed in two sprite-parent flashes. Coloring is by source altitude. Reference location for horizontal distance is the median location of the first 10 sources in each flash. Each sprite-parent +CG is indicated by a black cross. The dashed lines indicate 104 and 105 m s$^{-1}$ leader speeds. (a) Sprite parent occurring around 05:05:07 UTC on 8 June 2012. (b) Sprite parent occurring around 04:18:23 UTC on 25 June 2012.
again at 04:21:02 UTC. Total flash rate continued to decrease after 0425 UTC to an absolute minimum of 46 flashes during 0505–0510 UTC (9.2 min⁻¹). Between 0437 and 0511 UTC, during this total flash rate lull, there was a burst of 24 observed sprites. The rapid production of sprites was coincident with a large spike in mean flash area, which peaked above 1500 km² at 0455 UTC. The occurrence of large flashes helped drive an increase in flash energy around this time as well. Total flash rate recovered as sprite production terminated, although the storm never regained its pre-0400 UTC intensity.

This storm also was anomalously charged. During the analysis period 92.1% of the filtered CG lightning was positive (if no 15 kA filter were applied, the positive and negative rates were comparable to one another). In contrast to 8 June, the greatest production of +CGs occurred during its lull period, between roughly 0400 and 0600 UTC (Figure 8c). Peak +CG rates were about twice the number on 8 June as well (compare Figures 4c and 8c). Large positive-iCMC strokes were most frequent during this lull as well, especially the main sprite period (0437–0511 UTC). The values for large-iCMC strokes in this storm ranged between 75 and 290 C km. Throughout the analysis, LMA source densities were maximized near 7 km msl (approximately −15°C; Figure 8d). Manual analysis of the LMA data [Wiens et al., 2005] indicated frequent in-cloud activity between midlevel positive charge and a negative layer above it. This is a classic anomalous storm signature [Lang et al., 2004; Wiens et al., 2005; Lang and Rutledge, 2011; Fuchs et al., 2015]. The burst of sprites occurred after a sharp decline in source activity.
Figure 8. (a) Time series of echo volumes associated with various MRMS reflectivity thresholds, for the 25 June 2012 storm. (b) Time series of LMA-derived flash rate, mean flash area, and total flash energy for this storm. Also indicated are the times of the observed sprites. (c) Time series of NLDN-derived CG flash rates in this storm, separated by polarity, as well as the times of strokes that produced a positive iCMC above 75 C km. A 15 kA peak current threshold has been applied to the NLDN data. (d) Time-height plot of LMA source density in the 8 June 2012 storm.
Analysis of individual sprite-parent flashes indicated that they were inverted in structure, and occurred in or near convection, with leaders often tapping regions within the surrounding storm anvil. Figure 9 shows an XLMA-style plot [Rison et al., 1999] of a typical sprite parent in this storm. The flash in question was the first sprite parent, which initiated at 04:18:23 UTC and lasted about 2 s. The flash struck ground very near its initiation point, during a tightly clustered burst of LMA source activity. The subsequent activity indicated that negative leaders spread in all directions after the flash struck ground, though they traveled farthest to the southeast, toward the CSU-CHILL radar. Figure 6b shows the time-range behavior of this flash. It indicates that the sprite-parent +CG marked a transition toward a single negative leader that neutralized positive charge.
within the cloud during the TLE. This simpler behavior contrasts with the more complex bidirectional development that was common on 8 June (Figure 6a).

Up until 0450 UTC, the CSU-CHILL radar performed a series of RHI volumes, providing detailed vertical structure information as the storm was producing sprites. Figure 10 shows a relevant sweep near the time of the flash, through the center of the storm. The storm was comparable to 8 June in terms of vertical structure, with echo tops near 14 km above ground level (~15.5 km msl). The core contained a classic melting hail signature [Ryzhkov et al., 2013], with hydrometeor identification indicating a transition from hail to heavy rain with large drops (Figure 10f) as the precipitation fell from the storm core. This was associated with an increase in differential reflectivity below the melting level (Figure 10b), as well as strong propagation and specific differential phase shifts near the ground (Figures 10d and 10e). There were also alternating positive

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**Figure 10.** RHI sweep at 328° azimuth, from the CSU-CHILL radar volume starting at 0413 UTC. In all subplots, the black dots show the LMA observations of the sprite-parent flash shown in Figure 9, the white diamond is the median position of the first 10 LMA sources in the flash, and the black cross is the sprite-parent +CG. (a) Reflectivity. (b) Differential reflectivity. (c) Dealiased radial velocity. (d) Specific differential phase. (e) Differential phase shift. (f) Hydrometeor identification. Note that the vertical axis is above radar altitude (1432 m msl), not msl.
and negative differential phase shifts above the melting level, likely associated with the presence of dendritic crystals [Thompson et al., 2014] as well as crystals vertically aligned by strong ambient electric fields [Krehbiel et al., 1996; Carey and Rutledge, 1998].

The flash initiated to the southwest of this melting hail core (Figure 9), in a region where the RHIs indicated limited hail but considerable volume of low-density graupel (not shown). Despite the less intense precipitation structure, the upper level diffuence and midlevel confluence pattern in Doppler velocity (Figure 10c) were more pronounced toward the southwest near the center of the flash (particularly near 320° azimuth from CSU-CHILL), and an overshooting top was observed between the flash initiation point and the hail core shown in Figure 10. This indicates that the sprite-parent nonetheless initiated near a strong updraft.

Animations of all the sprite-parent flashes from this storm (available in the supporting information) show a very similar pattern to the flash in Figures 9 and 10. The flashes initiated near, but not in, the strongest precipitation cores. They showed no strongly preferred region of initiation, however, with slightly more than half initiating on the northern side of the storm and the rest the southern side. In addition, their horizontally propagating negative leaders (likely helping drive continuing current in the sprite-parent +CG) showed no preferred direction of travel. This is consistent with the weaker upper level flow compared to 8 June (Figure 3) and the more omnidirectional anvil structure on 25 June (e.g., Figures 2c and 10a).

4. Discussion

Lyons et al. [2008] demonstrated that sprites could occur over anomalously charged warm-season storms, but the present study is the first to robustly document this phenomenon. Both the 8 and 25 June 2012 storms were anomalous, with LMA-inferred positive charge near –10 to –20°C, in contrast to the negative charge that is commonly observed [Rust et al., 2005]. It is, of course, apt that such sprite-producing storms would be observed in Colorado, as the presence of anomalous charge structures is very common in this region [Lang and Rutledge, 2011; Fuchs et al., 2015]. The reasons for this are not totally clear, but the preference for higher cloud bases in and near Colorado (and hence reduced warm-cloud depths) likely plays a major role [Williams et al., 2005; Fuchs et al., 2015], especially when combined with strong instability and shear [Lang and Rutledge, 2011].

The hypothesis is that such a combination would produce large and vigorous updrafts, leading to increased liquid water contents (LWCs) [Williams et al., 2005]. This would tend to favor net positive charge transfer to riming particles (e.g., graupel) undergoing collisions with ice crystals [Saunders and Peck, 1998], as opposed to negative charge that would be expected under more typical LWCs [Takahashi, 1978]. Since the riming particles are heavier, they would fall relative to the ice crystals, leading to negative charge layer overlaying positive charge [Bruning et al., 2014] and thus an inverted electrical structure.

While the electrical structures of this study’s storms were complicated and featured substantial temporal and spatial heterogeneity (e.g., Figure 4d), the process described above is supported by the weight of the available evidence. Detailed updraft information was not available during the analyzed time periods, but both storms featured strong horizontal (Doppler-derived) flow structures (e.g., Figures 7 and 10), along with intense, vertically developed reflectivity structures. In addition, flash rates often were in excess of 60 min⁻¹. All this is indicative of peak updrafts likely exceeding 20 m s⁻¹ (e.g., Basarab et al., 2015). In combination with the high cloud bases (especially on 25 June), these collective observations are entirely consistent with anomalous electrical structures.

However, on 8 June the sprite-parent flashes displayed typical behavior, as documented by multiple past studies [Lang et al., 2010, 2011; Lu et al., 2013; van der Velde et al., 2014; Soula et al., 2015]. The sprite parents initiated as normal intracloud flashes (positive over negative), and then negative leaders traveled along downward sloping upper positive charge layers into the stratiform region, while the positive leaders struck ground. The charge structure in the 8 June storm was heterogeneous, with midlevel positive charge more commonly observed in the southwestern portion of the convection, while the northeastern portion featured a higher-altitude positive charge layer and was responsible for many of the sprite parents (e.g., Figure 5). This complexity is suggestive of the continuous variability in thunderstorm charge distributions described by
to the possibility that the lower ionosphere was in a state more favorable for the production of sprites

June were well in excess of the 200 C km threshold found by

Lyons et al. [2010] demonstrated that the full CMC estimates (including continuing current) from several sprite parents on 8 June were much higher than 8 June. This may have assisted with the anomalous charging, via the \([\text{flash} + \text{CG}]\) and the first video image with a sprite, has an uncertainty of \(\pm 0.017\) s due to video frame rate. Sprite duration also is affected by this uncertainty. LMA source altitude during the sprite time is set by LMA source altitudes from 50 ms before the +CG to 100 ms after.

By contrast, there was very little charge layer heterogeneity observed on 25 June; the storm was anomalous throughout the analysis period. While we lack enough data to determine the precise reason 25 June was more heterogeneous, it is interesting that the cloud base on 25 June was much higher than 8 June. This may have assisted with the anomalous charging, via the Williams et al. [2005] mechanism. Alternatively, but perhaps complementarily, the presence of wildfire smoke aerosol may have aided the anomalous electrification of this storm by helping suppress large precipitation growth and thereby invigorating updrafts and increasing their LWC [Lyons et al., 1998; Rosenfeld et al., 2008]. In addition, smoke has been implicated in increasing sprite production in storms [São Sabbas et al., 2010]. Detailed aerosol observations were not made for this case, however, so these hypotheses cannot be confirmed.

Sprite-parent flash characteristics for both storms are summarized and compared in Table 2. There are notable differences. In particular, 8 June sprite-parent flashes had larger areas and were longer lived compared to 25 June, although the flashes for both storms were much smaller than those reported for larger sprite-producing storms [Lang et al., 2010]. Peak currents for the sprite-parent +CGs were higher on 25 June, but ICBC magnitudes were similar for both storms. However, these ICBC magnitudes were smaller on average compared to past studies of sprite-producing flashes [Hu et al., 2002; Cummer and Lyons, 2005; Lang et al., 2011; Qin et al., 2012]. Indeed, ICBC values \(\sim 100\) C km are at the lower limits of those that have been associated with sprites [Lyons et al., 2009; Beavis et al., 2014]. Lyons et al. [2012] demonstrated that the full CMC estimates (including continuing current) from several sprite parents on 8 June were well in excess of the 200 C km threshold found by Qin et al. [2012], however. Thus, in addition to the possibility that the lower ionosphere was in a state more favorable for the production of sprites from weaker flashes, it is also possible that the continuing current made up for any low impulse CMC. However, full CMC retrieval has not been done for all sprites on these two days. Moreover, ICBC data were missing for part of 8 June. Thus, more research into the detailed coupling between individual flash characteristics and sprite production may be warranted for these cases, but this was beyond the scope of the present study.

Interestingly, while sprite parents in both storms initiated at a similar altitude (\(~6.6\) km), the 25 June sprite parents averaged \(~0.5\) km higher both around the sprite time and during the median lifetime of the flash.

<table>
<thead>
<tr>
<th>Variable</th>
<th>6/8/2012</th>
<th>6/25/2012</th>
</tr>
</thead>
<tbody>
<tr>
<td>NLDN peak current (kA)</td>
<td>54</td>
<td>86</td>
</tr>
<tr>
<td>Impulse CMC (C km)</td>
<td>105</td>
<td>114</td>
</tr>
<tr>
<td>LMA flash area (km(^2))</td>
<td>2040</td>
<td>978</td>
</tr>
<tr>
<td>LMA flash altitude (m msl)</td>
<td>6840</td>
<td>7313</td>
</tr>
<tr>
<td>Initiation altitude (m msl)</td>
<td>6606</td>
<td>6657</td>
</tr>
<tr>
<td>Sprite delay (s)</td>
<td>0.014</td>
<td>0.059</td>
</tr>
<tr>
<td>Altitude at sprite time (m msl)</td>
<td>6694</td>
<td>7286</td>
</tr>
<tr>
<td>Sprite duration (s)</td>
<td>0.068</td>
<td>NA</td>
</tr>
<tr>
<td>Time to +CG (s)</td>
<td>0.385</td>
<td>0.111</td>
</tr>
<tr>
<td>Flash duration (s)</td>
<td>1.686</td>
<td>0.485</td>
</tr>
</tbody>
</table>

\(^\text{a}\)Sprite durations were not determined for 25 June. The two storms’ distributions for the listed parameters are significantly different at \(^\text{b}\)99% confidence based on a two-tailed rank-sum test, except for flash initiation altitude and impulse CMC. Sprite delay, or the time between the +CG and the first video image with a sprite, has an uncertainty of \(\pm 0.017\) s due to video frame rate. Sprite duration also is affected by this uncertainty. LMA source altitude during the sprite time is set by LMA source altitudes from 50 ms before the +CG to 100 ms after.
We interpret this as an indication of the downward sloping of the stratiform components of flash channels, which was observed during 8 June but not in 25 June. Overall, the differences between the sprite parents in the two storms are consistent with the more convective nature of the sprite-parent lightning on 25 June (compare, e.g., Figure 5 to Figure 9).

On both case days, sprites did not begin occurring until the convection in the mature storms began weakening, especially as indicated by total flash rate and 40 dBZ echo volume. The sprites also often were associated with short-term spikes in mean flash area. In addition, the reduced convective strength was indicated by a general decrease in LMA-estimated flash energy during the sprite period, although the energy sometimes did undergo short-lived increases along with flash area, especially on 25 June. A transition to less frequent flashes with larger areas is a commonly observed behavior in weakening thunderstorms or away from updraft regions [Bruning and MacGorman, 2013]. Essentially, with fewer flashes to neutralize them, charge layers have more time to advect and spread out, so that when the flashes do occur, they cover a larger area. Many studies [Williams and Yair, 2006; Soula et al., 2009; Lang et al., 2010] have found that sprites were associated with weakening convective lines or intensifying stratiform regions, but these studies mostly looked at large MCSs, which often have many cells in various lifecycle stages. The June 2012 storms were smaller than typical MCSs [Houze et al., 1990]. In fact, during their sprite-producing stages both the 8 and 25 June storms were perhaps best described as unicells (e.g., Figures 1c and 2c). Thus, this study's results provide a much more direct link between the production of sprites and convective lifecycle. In these cases, sprite production clearly indicated weakening convection and a transition to less frequent flashes with larger areas. This is consistent with the reduction in flash rate during a transition toward sprite production that was observed in a sprite-producing MCS studied by Yang et al. [2013b].

However, the weakening should be understood in relative terms. During its sprite stage the 8 June storm was a mature, long-lived storm that remained dangerous and severe and was still producing more than a flash per second. And while its flash rate was much lower than that during its sprite stage, 25 June was nonetheless still producing significant amounts of hail, even if most of it appeared to melt before reaching the ground (e.g., Figure 10). The 25 June storm also was mature and long lived prior to sprite production. While total convective plus stratiform echo size was not calculated (CLEAR analysis focused on mainly the convective regions), based on Figures 1 and 2, 10 dBZ composite echo area for these storms was roughly ~10,000–20,000 km$^2$ during their sprite periods. This put them near or perhaps somewhat below the rough threshold of 20,000 km$^2$ for sprite production that was reported by Lyons [2006].

5. Conclusions

This study adds to the growing body of evidence [Lyons et al., 2003; Sáo Sabbas and Sentman, 2003; Lyons et al., 2008; Lang et al., 2010; Yang et al., 2013a; Soula et al., 2009, 2015] that sprite production is an indicator of weakening convection in mature long-lived storms and is also an indicator of a transition to less frequent flashes with larger areas. Moreover, due to the small and quasi-unicellular structure of the analyzed cases, this study provides the strongest evidence to date that sprite production can be directly related to convective trends in long-lived storms and does not necessarily require a robust stratiform precipitation region such as those typically observed in mature MCSs. Access to large reservoirs of stratified charge provides the best chance for a flash to produce a large-CMC ground stroke, which in turn maximizes electrical stress in the upper atmosphere. However, the stratified charge layers can be the result of detraining hydrometeors from vigorous and long-lived, but sub-MCS convection.

This study is also the first to fully document the production of sprites from anomalously charged storms, with LMA-inferred positive charge near midlevels. The results support the pioneering work of Lyons et al. [2008] on this issue. Such storms are capable of producing sprites from +CGs within their convective regions, and this implies that large and long-lived (but still submesoscale) anomalous storms are good candidates for the production of sprites from convective +CGs, due to their ability to produce and distribute widespread positive charge layers within their convective and adjoining anvils/stratiform regions. Such storms can produce sprites even when they are producing lightning at rates greater than 60 min$^{-1}$, but sprite production appears to be most favored during relative lulls in convective strength. This study thus provides additional support to the thesis of Lang et al. [2015] that the production of large-CMC and sprite-class lightning provides useful information about the meteorological state of thunderstorms.
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