Numerically Simulated Electrification and Lightning of the 29 June 2000 STEPS Supercell Storm

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(Manuscript received 30 June 2005, in final form 6 January 2006)

ABSTRACT

A three-dimensional dynamic cloud model incorporating airflow dynamics, microphysics, and thunderstorm electrification mechanisms is used to simulate the first 3 h of the 29 June 2000 supercell from the Severe Thunderstorm Electrification and Precipitation Study (STEPS). The 29 June storm produced large flash rates, predominately positive cloud-to-ground lightning, large hail, and an F1 tornado. Four different simulations of the storm are made, each one using a different noninductive (NI) charging parameterization. The charge structure, and thus lightning polarity, of the simulated storm is sensitive to the treatment of cloud water dependence in the different NI charging schemes. The results from the simulations are compared with observations from STEPS, including balloon-borne electric field meter soundings and flash locations from the Lightning Mapping Array. For two of the parameterizations, the observed “inverted” tripolar charge structure is well approximated by the model. The polarity of the ground flashes is opposite that of the lowest charge region of the inverted tripole in both the observed storm and the simulations. Total flash rate is well correlated with graupel volume, updraft volume, and updraft mass flux. However, there is little correlation between total flash rate and maximum updraft speed. Based on the correlations found in both the observed and simulated storm, the total flash rate appears to be most representative of overall storm intensity.

1. Introduction

The Severe Thunderstorm Electrification and Precipitation Study (STEPS) took place during 2000 to study severe storms in the high plains of the United States. One of the main goals of the project was to achieve a better understanding of the interactions among storm kinematics, microphysics, and electrification, especially in storms that produce predominately positive cloud-to-ground (CG) lightning (Lang et al. 2004). Numerous storms—including supercells, short-lived multicell storms, and large mesoscale convective complexes—were observed and documented during STEPS.

An in-depth study of storm processes requires a combination of observations and numerical simulations.
STEPPS provided a comprehensive observational dataset for detailed comparison with numerical simulations of storm evolution. The present study focuses on numerical simulations of the 29 June STEPS supercell that produced an F1 tornado, large hail, and predominately positive ground flashes. The objectives are to evaluate the simulated charge structure, lightning flash rate, and polarity in comparison with the observed storm and to determine the sensitivity of the modeled storm to different electrification parameterizations. The origins of positive (+) CG flashes and the relationships between the modeled total flash rate and storm characteristics are of particular interest.

a. Studies of inverted-polarity severe storms and positive CG flashes

The traditional conceptual model of the gross electrical structure of thunderstorms is that it can be described as either dipolar or tripoles (Figs. 1a and 1b), with the main charges being a middle-level negative charge and an upper-level positive charge. A small, sporadic positive charge was sometimes found beneath the negative charge to form the tripolar structure (e.g., Williams et al. 1989), and later it was found that a negative screening-layer charge is often near the upper cloud boundary. Several studies have suggested that dipole or tripoles are not sufficient to describe how charge is distributed in all thunderstorms. Rust and Marshall (1996) argued that the current tripoles are too simplistic to apply to all mature thunderstorms and mesoscale convective systems. A more complex charge structure consisting of four main charge regions near the updraft and six charge regions outside the updraft in the convective precipitation region was suggested by Stolzenburg et al. (1998). Other researchers have suggested that the tripoles is adequate to describe the basic thunderstorm charge structure, even arguing that abandonment of the tripoles model would be “ill-advised” (Williams 2001).

The existence of an inverted-polarity storm tripoles (Fig. 1d) was first noted by Marshall et al. (1995a) based on an electric field sounding in a strong storm near Dalhart, Texas. Marshall et al. suggested that +CG flashes in the Dalhart storm immediately following the sounding may have been caused by the inverted charge structure of the storm. Additional storms having inverted-polarity electrical structure were observed in the STEPS field program and were studied by Rust and MacGorman (2002), Rust et al. (2005), MacGorman et al. (2005), and Wiens et al. (2005). These last two studies include observations of the 29 June 2000 supercell storm being considered in the present paper.

Various studies have examined correlations between +CG flashes and severe weather. The first documentation of severe storms that commonly produce +CG flashes was provided by Rust et al. (1981), who concluded that the occurrence of +CG flashes may indicate a storm is severe. In a statistical study of Oklahoma storms, Reap and MacGorman (1989) found that storms that produced a larger number of +CG flashes had a higher probability of producing severe weather. Subsequently, Seimon (1993) noted +CG flashes preceding an F5 tornado, while MacGorman and Burgess (1994) showed that damaging tornadoes occurred after peaks in +CG flash rates in several storms. However, +CG flashes may prove ineffective as a severe weather indicator unless correlations can be shown to be reliable (Branick and Doswell 1992; Perez et al. 1997). Knapp (1994) and Carey et al. (2003b) have shown that +CG flashes are usually associated with severe storms in the central United States, not in the eastern states or coastal regions. Further study is needed to determine the relation of +CG flashes (if any) to the severity of the parent storm.

Several studies [including the above studies of +CG flashes, as well as MacGorman et al. (2001), Gilmore and Wicker (2002), MacGorman et al. (2005), and Williams (2001)] have examined the conditions needed to produce +CG flashes. Charge configurations hypothesized to be favorable for producing +CG flashes include tilting of the charge layers due to wind shear, precipitation unshielding of positive charge (i.e., fallout of a lower-level negative charge region revealing the upper positive charge), feedback formation of a lower
charge region based on the convective charging mechanism of Vonnegut (1963), the formation of a small lower negative charge below positive charge, and inverted-polarity storm electrical structure. In cloud model simulations, Mansell et al. (2002) found that +CG flashes were always initiated by the existence of negative charge just below a larger amount of positive charge.

Trying to understand why the kinematics and microphysics of a storm produce conditions conducive to +CG flashes is more challenging. Several investigators (e.g., Gilmore and Wicker 2002) have invoked the behavior of the noninductive mechanism that laboratory experiments, observations, and modeling have suggested is responsible for initial thunderstorm electrification (see discussion on pp. 65–68, 223–225, 329–331 of MacGorman and Rust 1998). Laboratory experiments have shown that the polarity of charge gained by graupel and cloud ice during rebounding collisions can depend on temperature, liquid water content, particle size, and the graupel riming rate (e.g., Takahashi 1978; Jayaratne et al. 1983; Keith and Saunders 1990; Jayaratne 1993; Pereyra et al. 2000). In regions in which graupel gains the majority of its charge in most thunderstorms, graupel usually appears to gain negative charge, with ice crystals gaining positive charge. After differential sedimentation, the resulting charge distribution is a dipole having the usually observed vertical polarity. The above laboratory studies generally agree that graupel gains positive charge at relatively high temperatures (the reversal to negative charging occurs at some temperature colder than $-10^\circ$C) or with large liquid water contents and graupel riming rates.

Positive charging of graupel at higher temperatures has been suggested as a source of the small, sporadic lower positive charge found below negative charge in most storms, and Gilmore and Wicker (2002) and Carey et al. (2003b) speculated that an unusually large liquid water content in a storm’s updraft could increase the amount of lower positive charge enough to produce +CG flashes. MacGorman et al. (2005) argued that, if negative charge is required below positive charge to produce +CG flashes, simply enhancing the lower positive charge would not be sufficient to produce +CG flashes. MacGorman et al. (2005) also suggested that the generation of an inverted-polarity electrical structure, as observed in some storms that produce +CG flashes, would require more than enhancing the lower positive charge. Positive charging of graupel needs to dominate charge generation in the storm, so that the main negative charge replaced by positive charge, with cloud ice then carrying negative charge into the upper part of the storm. Saunders and Peck (1998) found at sufficiently large riming rates that graupel tends to gain positive charge, regardless of temperature.

b. Goals of this research

The present study examines the consequences of different charge separation mechanisms on simulated electrification and compares the simulated charge structures to the observed charge structure of the well-documented 29 June 2000 STEPS storm (Tessendorf et al. 2005; Wiens et al. 2005; MacGorman et al. 2005). An analogous study has been performed using a two-dimensional model of the 19 July 1981 Cooperative Convective Precipitation Experiment (CCOPE) storm (Helsdon et al. 2001), in which three different noninductive charging schemes were examined. Results from the Helsdon et al. (2001) study suggested the laboratory results from Takahashi (1978) might be the best representation of NI charging. Mansell et al. (2005) tested different numerical parameterizations of noninductive and inductive charging in a small multicellular storm. The present study also examines multiple +CG flashes that occur at various times during both the simulations and the observed storm and relates these +CG flashes to the evolving kinematic, microphysical, and electrical structure of the storm.

Another interest is the relationships between the flash rate and the storm characteristics in the simulations. Several investigators have found relationships in observed storms between reflectivity or updraft strength and total flash rate (Goodman et al. 1988; MacGorman et al. 1989). Baker et al. (1995) and Solomon and Baker (1998) examined the total flash rate in conjunction with updraft speed, reflectivity, precipitation rate, ice concentration, and cloud radius in a one-dimensional model. The present study employs a three-dimensional model to further examine the relationships between flash rate and other storm properties such as updraft speed and mass flux, precipitation rate, graupel volume, and rain mass.

2. Model description

a. Dynamics and microphysics

The dynamic cloud model is described in detail by Straka and Mansell (2005). The model is three-dimensional, nonhydrostatic, and fully compressible, and is based on the set of equations from Klemp and Wilhelmson (1978). The model includes prognostic equations for velocity components (momentum), perturbation pressure, potential temperature, turbulent kinetic energy, water vapor and hydrometeor mixing ra-
tios, rime history, and charge variables. The model employs a microphysics package that includes 2 liquid hydrometeor categories and 10 ice categories distinguished by particle density, habit, and size (Straka and Mansell 2005). The ice habits include small and large hail; low-, medium-, and high-density graupel; snow aggregates; rimed ice; plate ice; and column ice.

b. Electrification and lightning

The model includes a choice of parameterizations for hydrometeor charging (Mansell et al. 2005). This study uses both inductive and noninductive (NI) charging for electrification (see the appendix for details). The results of laboratory, modeling, and observational studies strongly suggest that noninductive charging plays the primary role in producing electrification rates and magnitudes close to those of observed storms (MacGorman and Rust 1998; Helsdon et al. 2002). However, there is laboratory support for inductive charging to also play a role (e.g., Brooks and Saunders 1994). For this study, the model employs four different parameterizations of the NI charging process, each varying in the determination of the sign and level of charging. All of the NI schemes used in this study are described in detail in Mansell et al. (2005), but will be briefly described below for convenience.

The noninductive charging rate in the Saunders and Peck (1998, hereafter SP98) scheme is based on a critical rime accretion rate (RAR) from their measurements. The temperature-dependent critical RAR value ($RAR_{crit}$) defines positive and negative charging regions (Fig. 2a). The sign of the charge transferred to the graupel during a rebounding collision in the SP98 scheme is strongly influenced by the amount of water accreted on the graupel (i.e., the rimer). The riming rate (RR) scheme is developed in a similar fashion to...
that of SP98. It is also based on a critical rime accretion rate, but with a slightly different temperature and liquid water dependence (Fig. 2b).

The Takahashi charging scheme is based on the laboratory work of Takahashi (1978, hereafter TAKA). The polarity of charge gained by graupel is determined by the cloud water content and temperature (Fig. 2c). For the model parameterization, the results are taken from a look-up table used by Randell et al. (1994), with additions from Takahashi (1984) for variation in the charge separation per collision, which is dependent on impact velocity and crystal size. The charge separated per collision at temperatures between 0°C and −30°C and liquid water content from 0.01 to 30 g m⁻³ are included in the table. At temperatures below −30°C the charge separated per collision is the value at −30°C.

The final noninductive parameterization used is the Gardiner–Ziegler (GZ) scheme. This scheme is based on the laboratory results of Jayaratne et al. (1983), as adapted from Gardiner et al. (1985) by Ziegler et al. (1986, 1991). The dependence on liquid water content is given by an adjustable reversal temperature $T_r$ and the cloud water mixing ratio. At temperatures below $T_r$, graupel (ice) charges negatively (positively) and at higher temperatures the sign of charging is reversed. In the present study, the reversal temperature is set to −15°C (Fig. 2d).

Lightning flashes are parameterized by a stochastic dielectric breakdown model (Mansell et al. 2002). The lightning develops bidirectionally across a uniform grid with each step chosen randomly from among the surrounding points at which the electric field meets or exceeds a threshold value for propagation. After each step, the electric field is recalculated to determine the contribution by the lightning channel. The resulting flash has a branched or fractal-like leader structure in three dimensions. Positive leaders carry positive charge and travel preferentially through negative charge regions, while negative leaders carry negative charge and tend to travel through regions of net positive charge (Mansell et al. 2002). Therefore, the simulated flashes tend to reflect the simulated charge structure.

3. Observed and simulated 29 June storm evolution

a. The 29 June 2000 supercell storm

The 29 June 2000 supercell storm formed just ahead of a dryline with an approaching mesoscale cold front to the north (Fig. 3a). The storm’s first radar echo appeared around 2130 UTC near the borders of Colorado, Nebraska, and Kansas (Fig. 3b). The storm lasted approximately 4 h, moving southeastward through northwest Kansas (Fig. 3b) before being overtaken by part of a mesoscale convective system in central Kansas later that evening. During the first 3 h, the storm produced large hail, an F1 tornado, and a profuse amount of lightning. As the storm moved through the STEPS domain, it was observed by a network of Doppler radars, including two polarimetric radars, the T-28 armored research airplane, balloon soundings of the electric field, and the Lightning Mapping Array (LMA). The LMA is a GPS-based system that locates sources of VHF radiation from lightning discharges in three spatial dimensions and time (Rison et al. 1999; Kreibiel et al. 2000). The balloon-borne electric field meter (EFM) measures the vector electric field, $\mathbf{E}$ (as described by Winn et al. 1978; Marshall et al. 1995b; Coleman et al. 2003).

Atmospheric conditions in the storm environment supported supercell storm development (Johns and Doswell 1992; Weisman and Klemp 1982). Environmental winds near the storm were from the south at ground level and veered to the west with height as shown in the 2022 UTC National Center for Atmospheric Research (NCAR) GPS Dropsonde Loran Atmospheric Sounding System (GLASS) sounding from Goodland, Kansas (Fig. 4). The sounding was released approximately 65 km to the southeast of where the storm initiated, and 1 h before it was first detected by radar. Although the environment was unstable as depicted by the 1319 J kg⁻¹ of convective available potential energy (CAPE), the sounding was capped as indicated by a convective inhibition (CIN) of about 100 J kg⁻¹. Strong 0–3-km storm relative helicity (SRH) indicated the support for the development of rotating updrafts during the supercell phase of the storm.

The observed storm was multicellular during its early stage, developing reflectivity factor values up to 54 dBZ and several dominant updrafts by 2305 UTC (Fig. 5a). At approximately 2330 UTC the storm made a right turn and slowed as it moved southeastward (Figs. 3b and 6). The graupel volume of the storm increased in height and amount at a greater rate during this period than at any earlier time (Fig. 7a), while the updraft volume (for $w$ greater than 10 m s⁻¹) increased from 70 to 110 km³ in about 0.5 h (Fig. 7c). During this period, the Doppler-derived maximum updraft speed was near 50 m s⁻¹ (Tessendorf et al. 2005) as the storm intensified, assumed a supercell structure, and increased in size. A bounded weak echo region (BWER) also became apparent following the right turn (Figs. 5c and 8a–c).

The LMA indicated that the storm began producing lightning at 2150 UTC, 20 min after the first radar echo (Fig. 9a). The lightning rate in the developing stages
was 20 flashes min\(^{-1}\), but by approximately 0015 UTC
the storm reached a maximum of 300 flashes min\(^{-1}\) (Fig. 9a). During the first 3 h, the storm produced approximate 10 000 total flashes (Table 1). The flash counting algorithm determined an actual “flash” as that consisting of at least 10 source points detected from the LMA’s VHF detectors in close proximity and time (Wiens et al. 2005). The first CG flash was detected at 2239 UTC by the National Lightning Detection Network (NLDN). The first ground flash and approximately 90% of the CG flashes thereafter were positive. A total of 140 +CG and 19 −CG flashes were counted by the NLDN during the first 3 h of the storm. Only 1.5% of the total lightning were CG flashes. Both the total flash rate and CG flash rate rapidly increased as the storm displayed extensive development and made its right turn at approximately 2330 (Fig. 9a).

The charge structure of the 29 June storm was inferred using LMA activity and EFM soundings. Observations from the LMA can be used to infer charge structure since detected leaders move through regions of opposite polarity charge. The LMA preferentially detects negative polarity breakdown, and inferred charge regions therefore tend to be positive (Hamlin et al. 2003). By the time +CG flashes occur in the storm, a persistent midlevel positive charge region had developed. Hence, an inverted charge structure (similar to Fig. 1c) was inferred from the LMA observations of +CG flashes in the present case (Wiens et al. 2005). The +CG flashes typically initiated with upward negative leader development into the positive charge region just above 5 km simultaneously with downward positive leader development through a negative charge region to ground. A balloon-borne EFM sent through the updraft region of the storm depicted a main positive charge region centered near 9 km and an upper negative charge region centered at 11 km using the 1D Gauss method of interpretation of Marshall and Rust (1991) (Fig. 15f).

b. Model initialization
The 29 June supercell storm was simulated on an 80 km by 80 km by 20 km domain. The horizontal grid spacing was 1 km, while the grid was stretched vertically from 200 m at the surface to 500 m above 4 km. The horizontally homogeneous model environment was initialized using a modified version of the NCAR mobile GLASS sounding from Goodland, Kansas (Fig. 4). The temperature and moisture in the convective boundary layer were increased to better depict surface observa-
tions of the environment into which the storm moved (Fig. 3). In particular, a mobile mesonet observation near the storm (Fig. 3b) recorded higher temperatures and dewpoints than the Goodland sounding at 2022 UTC (E. Rasmussen 2004, personal communication). The base of the elevated residual layer capping the moist convective boundary layer was warmed adiabatically to maintain a minimum concentrated cap strength, thus controlling the spurious growth of instabilities and preserving the mixed layer. The instability of the environment in the modified sounding was thus greatly increased, raising the CAPE from 1370 to 2875 J kg$^{-1}$ and lowering the CIN from 100 to 22 J kg$^{-1}$. The bulk Richardson number (BRN), defined as the ratio of the CAPE to the lower-tropospheric vertical wind shear, increased from 10 to 23. The CAPE and BRN of the modified sounding supported possible supercell development (Weisman and Klemp 1982). A warm bubble (Δθ = 3 K) with randomized thermal perturbations and a horizontal radius of 9 km was used to initialize the convection.

c. Dynamical and microphysical evolution

The present model configuration allows no feedback from the electrification to the microphysics or dynamics. Therefore, each of the four simulations has exactly the same dynamical and microphysical evolution despite differing charging schemes. The simulated storm initially develops an elongated multicell structure (Fig. 5b) with successive main updraft cores along the upshear side of the outflow. By 76 min, the storm has developed a solid core of reflectivity extending to ground with a deep updraft and forward anvil region. During the first 60 min, the storm moves toward the east-northeast. It is hypothesized that storm rotation and the cold pool have intensified sufficiently to force the storm to turn right toward the southeast and decelerate by 90 min. At 104 min, the precipitation core intensifies while a low-precipitation, cloud-filled mesocyclone develops on the southwest flank (Fig. 5d). By 116 min, the simulated storm has developed a pronounced BWER coincident with the intense main updraft on the southwest flank of the storm (Figs. 8d–f). The storm continues along a southeasterly track for the remainder of the simulation (Fig. 6).

The timing of the right turn is used as a basis for comparison between the simulated and observed storms. The early development of the observed storm was much slower than in the simulations due to the distinctly different initiation processes, the latter being initialized by a thermal bubble and the former forced from boundary layer convergence just east of the dryline. The developments of the observed and modeled storms are in rather close agreement from the time of the right turn onward, as supported by comparison of storm morphologies after 90 min of the simulation and at 2330 UTC in the observed storm (e.g., Figs. 7 and 8).

The maximum updraft speed of the simulated storm reaches 30 m s$^{-1}$ at 16 min and becomes stronger throughout the simulation, with a peak of 61 m s$^{-1}$ at 147 min (Fig. 10). The simulated supercell exhibits evidence of convective surges during its life cycle. The first growth phase occurs at approximately 20 min with increases in updraft mass flux (Fig. 10), graupel volume (Fig. 7b), and updraft volume (Fig. 7d). Another convective surge is centered at 80 min (Figs. 7b, 7d, and 10) as the storm veers toward its southerly track. The maximum strength of the storm occurs between 140 and 160 min, when updraft mass flux and graupel volume peak values and a reflectivity maximum of 69 dBZ is attained. The overall simulated storm evolution is similar to that of the observed storm, especially after 90 min. This agreement is significant as most of the total lightning and virtually all the CG flashes occur after the right turn in the simulations and the observed storm.

Values of liquid water content (LWC) during the simulations are comparable to aircraft observations of
supercell storms. Musil et al. (1986) found LWC in the updraft regions of 1–3 g m⁻³ with a peak near 6 g m⁻³, where the updraft was greater than 55 m s⁻¹. In the simulation, values in the updraft regions are approximately 1–2 g m⁻³ on the edge of the updrafts with peaks of 3–5 g m⁻³ in the interior, where \( w \) is near 40 m s⁻¹. Mixing ratios are shown for cloud ice, aggregates, and cloud droplets in Fig. 11 as well as low-, medium-, and high-density graupel and rainwater [see Fig. 5 of Straka and Mansell (2005) to compare with multicell storm simulation]. Cloud ice, and low- and medium-density graupel are located at heights up to 14 km due to the smaller fall velocities and orientation of the updraft. High mixing ratios of cloud water, medium-density graupel, and cloud ice (at higher elevations) make up the majority of the content of the main updraft region and are vital to the electrification of the storm.

Fig. 5. (left) Triple-Doppler analysis at 5 km AGL based on S-band dual-polarization Doppler radar (S-Pol), CHILL, and Goodland Weather Surveillance Radar-1988 Doppler (WSR-88D) observations on 29 Jun 2000 (Tessendorf et al. 2005). Reflectivity with ground-relative vectors and 10 m s⁻¹ updraft (black contour) at (a) 2305 and (c) 2343 UTC. [Analyses provided courtesy K. Wiens (2005, personal communication).] (right) Parameterized reflectivity from model simulations at (b) 66 and (d) 104 min with ground-relative wind vectors, 10 m s⁻¹ updraft (black contour), and cloud outline (gray contour) at 5.3 km AGL. Corresponding times are relative to the right turn taken by the observed storm at 2330 UTC and the simulated storm at 90 min (Fig. 6).
d. Evolution of electrical properties

1) Rime accretion rate (RAR)-based noninductive charging

As previously discussed in section 2b, the sign of the SP98 noninductive charging is dependent on the RAR and the degree of supercooling (see Fig. 2a). The high liquid water content in the updraft caused a transfer of positive charge to graupel to dominate the charge produced by the SP98 scheme during the early electrification phase. The resulting charge morphology features a midlevel positive charge with an upper negative charge at 28 min (Fig. 12a).

The inverted dipolar charge structure is replaced by an inverted tripoal structure at about 35 min (not shown), as precipitation recycling and fallout and inductive charging quickly develop a lower negative charge region for the third layer. By 76 min, the storm exhibits an inverted tripolar structure with a main midlevel positive charge region between two negative charge regions (Fig. 12b). All three charge regions extend horizontally through much of the storm. A positive surface corona charge layer is also noted below 0.5 km AGL at this time. The mature stage of the storm at 116 min depicts a very complex structure with opposite charges occurring at the same altitude (Fig. 12c). The reflectivity core regions maintain a tripoal structure, but outside this region there are five or more vertically layered charge regions. The overall charge structure is similar to that of an inverted storm as proposed by Marshall et al. (1995b) with complexities as described in Stolzenburg et al. (1998).

Simulated intracloud (IC) flashes begin at 28 min, with a flash rate of approximately 30 flashes min$^{-1}$ during the first hour (Fig. 9b). The IC flash rate briefly reaches a maximum of 264 flashes min$^{-1}$ at 120 min, then decreases slowly while maintaining a flash rate above 150 flashes min$^{-1}$ during the remainder of the simulation. Lightning leaders travel preferentially through layers of opposite charge, with positive leaders concentrated in negative charge near 5 and 13 km (Fig. 13a). Conversely, midlevels of the storm are dominated by negative leaders and positive charge (Fig. 13b).

The SP98 scheme produced a total of 98 +CG flashes, the first occurring at 67 min, with no negative ground flashes produced (Table 1). The CG flashes typically initiate between 5 and 7 km (Fig. 13a), between the main positive charge region above and a negative charge region below (Fig. 14). In the simulation, +CG flashes are composed of a negative leader traveling upward through positive charge and a positive leader traveling downward through negative charge to ground (Fig. 14). The majority of the CG strikes are located just downshear of the main convective core, though some occur directly under the main updraft.

The riming rate (RR) noninductive charging scheme is also dependent on the rime accretion rate, though with a different critical RAR than the SP98 scheme (Fig. 2b). This produces a slightly different local charge distribution (Figs. 12d–f), but lightning rates (forced by macroscopic charge distributions) are similar to those of SP98 (Figs. 9c and 13c).

2) Takahashi (TAKA) and Gardiner–Ziegler (GZ) noninductive charging

The charge structure resulting from the TAKA noninductive charging scheme is similar to the conceptual model of a normal polarity dipole or tripoal structure. During early electrification, a normal dipole charge structure develops in the updraft region (Fig. 12g). The storm quickly develops a normal tripoal structure as a lower positive charge region is formed. At 76 min, the charge structure consists of a midlevel negative charge, an upper positive charge region, and a lower positive charge region situated in the updraft area (Fig. 12h). By 116 min, the storm is reaching its mature stage and the...
Normal tripole structures are evident in the reflectivity cores, though outside of these convective cores the charge structure contains up to six vertically layered charge regions.

The IC flashes begin at 27 min and the flash rate averages about 20 flashes min$^{-1}$ for the first hour (Fig. 9d). Steady growth of the flash rate continues for the second hour, and a maximum of 208 flashes min$^{-1}$ occurs at 122 min. A dropoff in flash rate follows, and during the third hour the flash rate stays near 120 flashes min$^{-1}$. Positive leaders are concentrated in the midlevels of the storm between 4 and 10 km (Fig. 13e). Although the majority of negative leaders are in the upper levels of the storm between 8 and 16 km, negative leaders are also noted in lower portions of the storm (Fig. 13f).

A total of 63 ground flashes, all negative, are produced during the TAKA simulation (Table 1), the first CG occurring at 71 min (Fig. 9d). The initiation points for ground flashes are limited to between 4 and 6 km (Fig. 13f), between a lower positive charge region and middle negative region.

The GZ noninductive scheme (Figs. 12j–l) develops a normal dipolar or tripolar charge structure that is broadly similar to the TAKA-simulated charge distribution. However, the lower positive charge region in the GZ simulation does not develop until late in the third hour of the simulation and remains weaker than that of the TAKA simulation. The GZ simulation produces only five CG flashes: three positive and two negative. The initiation heights of the +CG flashes for the GZ simulation (Fig. 13g) are anomalous for any model simulation, possibly implying an unrealistic result of the CG lightning scheme since the length of the positive and negative leaders in the three +CG flashes are unbalanced. The two negative (−) CG flashes (Fig. 13h) seem more realistic and broadly comparable to the −CG flashes produced using the TAKA scheme.

e. Observed and simulated electric field meter soundings

A balloon carrying an electric field meter (EFM) was released from Brewster, Kansas, at 0004 UTC into the simulated storm (km$^3$). (c) Volume of updrafts >10 m s$^{-1}$ (km$^3$) as inferred from triple-Doppler analysis. (d) Volume of updrafts >10 m s$^{-1}$ (km$^3$) in the simulated storm (km$^3$). Same scales are used in (a), (b) and (c), (d), and the time scale is aligned according to when each storm took its right turn. Plots of observed variables are adapted from Tessendorf et al. (2005).

Fig. 7. Time–height sections of selected radar-observed and simulated storm variables. (a) Graupel echo volume (km$^3$) as inferred from S-Pol radar from 29 Jun. (b) Graupel volume in the
updraft region of the storm (Figs. 8a and 8c). Using a 1D Gauss model (Marshall and Rust 1991; Stolzenburg and Marshall 1994), it is assumed that vertical gradients in $E_z$ are caused by the EFM rising through regions of charge. Interpreting the observed vertical electric field profile using the 1D Gauss technique, the 29 June storm’s charge profile is inferred to consist of a main positive charge region from 8 to 10 km and a main negative charge region from 10 to 11 km (Fig. 15f). A detailed analysis of the three-dimensional electric field vector profile reveals additional smaller charge regions along the balloon track through the reflectivity core of the storm (MacGorman et al. 2005).

A simulated EFM balloon was also “released” into the main updraft of each of the model simulations at 113 min (Figs. 8d, 8f, and 15b). The ascent rate was assumed to be 5 m s$^{-1}$ greater than the vertical air motion. The simulated EFM followed roughly the same track at approximately the same time during the storm’s life cycle as the observed EFM sounding. Both the simulated and observed soundings are contained within the intense main updraft region, rising through the BWER into the overlying precipitation core. Consequently, almost all charge is above 8 km at the level where the EFM penetrates the top of the BWER. The soundings maintain similar tracks below 12 km. The simulated EFM sounding detrains from storm top above 12 km, while the observed sounding moves horizontally and subsequently descends through the storm without reaching the top (possibly due to the balloon being punctured by hail). The profiles of observed and simulated relative humidity with respect to water saturation are remarkably similar profiles below 12 km, implying comparable altitudes of the transition from supercooled drops to cloud ice (Fig. 11).

The simulated balloon soundings have an inverted tripolar charge structure for the SP98 and RR simulations (Figs. 15a and 15c). The height and magnitude of the upper negative and main positive charge regions are similar to those of the corresponding regions from the observed storm. Both simulations also include a small lower negative charge region near 8 km, a feature that is not revealed in the observed sounding. The simulated sounding could be closer to the lower negative charge region than the observed sounding, the latter beginning farther away from the storm (e.g., Figs. 8c, 8f, and 15b).

Fig. 8. (top) Radar reflectivity from Goodland, KS, WSR 88D radar at 0004 UTC. (a) The 0.5° elevation scan, cross sections for (b) and (c) denoted by solid black lines. The path of the EFM balloon launched from Brewster, KS, at approximately 0004 UTC into the main updraft (see Fig. 15) is denoted by a black dashed line. (b) Cross section along line AB. (c) Cross section along line CD, path of balloon shown by black dashed line. (bottom) Reflectivity, ground-relative vectors, and cloud outline (gray) from simulations at 116 min. (d) X–Y planar view at 6.8 km. Path of EFM balloon along short dashed line from 0 to 12 km (see Fig. 15). Line AB denotes the cross section used in Fig. 11 while the long dashed line denotes cross section used in the right column of Fig. 12. (e) Cross section along line AB. (f) Cross section along line CD, maximum vector of 54.5 m s$^{-1}$ in updraft. The path of the simulated EFM balloon along dashed line begins at 113 min (at 116 min, EFM is just below 2 km). Compare with Fig. 2 of MacGorman et al. (2005).
The EFM soundings from the TAKA and GZ simulations depict a normal tripolar charge structure (Figs. 15d and 15e), in direct opposition to the inverted tripole of the SP98 and RR simulations and the observed sounding. The TAKA and GZ soundings reveal a small lower positive charge region below 8 km, a main negative charge region centered near 10 km, and an upper positive charge region from 10 to 12 km.

The analysis of the observed downward EFM profile from 29 June depicts a much more complex profile in the precipitation core downwind from the updraft region (MacGorman et al. 2005). Interpretation of charges via the 1D Gauss model depicts four or five levels of charge in the precipitation core, in contrast with two or three charge layers in the main updraft region. A similarly complex profile is also evident in the SP98 simulation outside the main updraft regions (e.g., Fig. 12c), where pockets of charge account for added complexity in the profile.

4. Discussion and conclusions

a. Role of noninductive charging in charge structure development

The simulated storm charge structure depends strongly on the choice of noninductive charging parameterization. As will be discussed in more detail, the two schemes based on rime accretion rate (SP98 and RR) develop very similar inverted-polarity charge distributions that can be described roughly as inverted-polarity

<table>
<thead>
<tr>
<th>Charging scheme</th>
<th>No. of flashes</th>
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<tr>
<td></td>
<td>Tot (IC + CG)</td>
</tr>
<tr>
<td>OBS (LMA/NLDN)</td>
<td>10 000</td>
</tr>
<tr>
<td>SP98</td>
<td>17 274</td>
</tr>
<tr>
<td>RR</td>
<td>13 243</td>
</tr>
<tr>
<td>TAKA</td>
<td>12 678</td>
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<td>GZ</td>
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Fig. 9. Lightning time series for the 29 Jun supercell and simulations. (a) Flashes per min counted from LMA detection (black) and total CG per min from NLDN (gray). Adapted from Wiens et al. (2005). (b)–(e) In-cloud flashes per min (black) and cloud-to-ground flashes per min (gray). (b) SP98 noninductive charging, (c) RR noninductive charging, (d) TAKA noninductive charging, and (e) GZ noninductive charging.
tripoles. The liquid water content scheme (TAKA) and the simpler temperature reversal scheme (GZ) develop normal polarity charge distributions. The rime accretion rate parameterizations appear to provide the best overall agreement with the observed electrical structure on 29 June, which can be approximated as an inverted-polarity tripole with an upper positive screening layer.

An examination of hydrometeor charging at 76 min illustrates the impact of the various laboratory charging constraints on the simulated electrification (Fig. 16). The majority of collisions between riming graupel and ice occur in the main updraft region, so charging in this region has a major influence on the electrical structure of the storm. Under the SP98 scheme, which deposits positive charge on graupel at rime accretion rates above 1 g m⁻² s⁻¹ at high temperatures and above 3 g m⁻² s⁻¹ at lower temperatures (Fig. 2a), the graupel in the cloudy updraft charges positively at all temperatures owing to the large rime accretion rates (Fig. 16a). Thus, the inverted-polarity charge structure is controlled by the positive charging of graupel and the negative charging of rebounding ice in the updraft region, which provides the main positive charge center at middle levels of the storm and the upper negative charge region. Immediately downshear from the updraft core, however, graupel charges negatively, due to the very low rime accretion rates there. These negatively charged graupel are subsequently transported via sedimentation and advection to lower regions of the storm (Fig. 16a), where a negative charge layer forms, as illustrated previously. Additional contributions to the lower negative charge region are provided by inductive charging of recycling graupel outside the updraft core. The major features of this charge distribution are similar to those of the observed distribution.

Although the RR and SP98 charging schemes have slightly different functional dependencies on rime accretion rate, they produce very similar charge distributions. At 76 min, the main difference in graupel charging between RR and SP98 occurs adjacent to the updraft from −30°C to −40°C (Fig. 16b), where the SP98 RARₘₜᵣᵢₚ drops toward zero (Fig. 2b), but the RR...
RAR$_{\text{crit}}$ does not. Thus, in this region graupel gains positive charge in the SP98 simulation and gains negative charge in the RR simulation. This has a relatively minor effect, so the charge structure resulting from the RR simulation is very similar to that from the SP98 simulation. However, the lower negative charge region in the RR simulation is stronger and slightly more dominant than in the SP98 simulation, probably influenced

Fig. 12. Charge structure for (a)–(c) SP98, (d)–(f) RR, (g)–(i) TAKA, and (j)–(l) GZ simulations at (left) 28, (middle) 76, and (right) 116 min. Red and blue shading denotes positive and negative charge regions, respectively. Lighter shading indicates areas of at least ±0.1 nC m$^{-3}$, darker shading indicates areas of at least ±0.25 nC m$^{-3}$. Cloud outline is a gray contour. Black contour is (left) 25 and (middle), (right) 45 dBZ. Compare (c), (f), (i), and (l) with descent sounding illustrated in Figs. 2a and 2f of MacGorman et al. (2005).
by the additional negatively charged graupel recycling through lower levels of the storm.

The TAKA noninductive scheme produces a charge structure that has normal vertical polarity, the opposite of the structure produced by the SP98 and RR charging schemes and observed in the storm. Instead of the polarity of graupel charge depending on rime accretion rate, as in the previous two schemes, it depends on...
Fig. 14. Positive cloud-to-ground flashes during the SP98 simulation. (left) Vertical cross section through storm. Positive and negative charge regions are contoured in solid red and blue, respectively; vectors are of the electric field; and black and gray contours indicate equipotential lines. Lightning leaders in white fill contour, positive with red outline, negative with blue outline. (right) A 3D view of flash, initiation shown in green, positive leaders in red, and negative leaders in blue; signs of net charge are indicated. Location of x–z cross section shown in left panel denoted by gray-dashed line: (a) 80, (b) 137, and (c) 170 min.
Fig. 15. (a), (c)–(e) Simulated EFM at 113–140 min in the four different simulations. Graphs show $E_z$ (black), temperature (red), relative humidity (blue), and rise rate of the balloon (brown). Red bars and blue bars represent height of positive and negative charge levels using 1D approximation to Gauss’s law, smaller bars represent charge region of smaller magnitude. (Height is converted to MSL to compare with observations.) Flight path of balloon superimposed on reflectivity in Figs. 8d and 8f. (b) Path of balloon through SP98 simulation denoted by dashed black line; positive (red) and negative (blue) charge regions shown (height AGL). (f) EFM balloon sounding at 0004:58–0034:09 UTC 30 Jun 2000, displaying same variables as in (a). See Figs. 8a and 8c for path of balloon on 29 Jun.
cloud liquid water content as in the Takahashi scheme. At 76 min, the graupel charges negatively in the updraft region, which contains liquid water contents from 0.5 to 2.5 g m\(^{-3}\) at temperatures less than \(-10^\circ\)C (Fig. 16c). The predominately negative charging of graupel and positive charging of snow and ice within the updraft region (Fig. 2c) produce a main midlevel negative charge region (Fig. 12h) and an upper positive charge region, due to the differential fall speeds of the hydrometeors. A small lower positive charge region develops from weak positive charging of graupel by the noninductive mechanism outside the main updraft core in low liquid water content. This lower positive charge region is also enhanced by inductive charging of roughly the same magnitude as the noninductive charging.

At 76 min, the GZ noninductive charging scheme, like the TAKA scheme, develops a normal dipolar
The main negative charge region and upper positive charge region are produced by the advection and differential sedimentation of negative graupel and positive snow and ice charged by the noninductive mechanism at temperatures less than \(-15°C\) (Figs. 2d and 16d). Small pockets of weak positive charge densities form at lower levels due to positive noninductive charge transfer to graupel at temperatures greater than \(-15°C\) (Fig. 16d). The GZ scheme possibly has difficulty producing the lowest charge region due to the specific characteristics of its simple dependence on temperature. While the other schemes produce opposite-sign charging in the downshear updraft flank, the GZ scheme produces opposite charging only at the base of the updraft, and this is then countered by the charging at lower temperatures as graupel is lifted higher within the updraft. Inductive charging acts to enhance the preexisting charge outside of the updraft core.

The overall charge structure of the mature storm in all the simulations is much more complex than the basic dipole or tripole conceptual model, consistent with the observations of Stolzenburg et al. (1998). The early storm development is multicellular, and each successive updraft and precipitation core produces another region of charging, each with its own evolving vertical charge structure and each advecting charge into adjacent regions. Thus, pockets of concentrated charge develop throughout the storm. The charge structure is further complicated by the parameterized lightning activity.

When a simulated leader travels through a region of opposite polarity, a localized reversal in the net charge can occur within the larger charge region, thereby leading to a more complex charge structure (Helsdon et al. 1992; Mansell et al. 2002; Coleman et al. 2003). Furthermore, even if the net polarity of charge is not reversed locally by lightning, subsequent differential sedimentation and advection of cloud and precipitation particles that capture the charge from frequent lightning flashes create additional layers of charge through the anvil region (Ziegler and MacGorman 1994).

b. Positive CG development relative to storm polarity

Over a period of 3 h, the observed storm produced roughly 160 CG flashes, most of which lowered positive charge to ground, and the simulations employing the SP98 and RR noninductive charging schemes produced similar +CG activity (Fig. 17). In both the observed and simulated storms (i.e., SP98, RR, TAKA), CG flashes account for roughly 1% of the total lightning activity.

As mentioned previously, Williams (2001), Lang and Rutledge (2002), and Wiens et al. (2005) discussed several hypotheses concerning the charge structure required for storms to produce frequent +CG flashes. These charge structures included a tilted dipole (i.e., the updraft is sheared by strong mid- and upper-level winds, exposing the upper positive charge region to ground), precipitation unshielding (i.e., the positive re-
region is exposed to ground), and an inverted dipole. Though some of these configurations may apply to other storm situations, none is completely consistent with the observations and simulations of the STEPS 29 June storm (Fig. 17). The inverted dipole is close, but it oversimplifies the distribution and would not by itself be expected to produce +CG flashes.

MacGorman et al. (2005) and Wiens et al. (2005) suggested that the observed inverted-polarity charge structure of the 29 June storm may have been responsible for the tendency of the storm to produce +CG flashes. However, they noted that the inverted dipolar structure observed in the updraft core probably was not adequate for initiating +CG flashes, but that a lower negative charge that was observed in the rainy downdraft may have been needed, as suggested by Mansell et al. (2002). Similarly, in the simulations, the prevailing polarity of the simulated ground flashes is governed by the polarity of the main midlevel charge region in the storm, and a charge region of the opposite polarity is needed beneath this main charge to initiate the flash. Only the SP98 and RR simulations develop an inverted-polarity charge structure, and so produced +CG flashes. The TAKA and GZ simulations develop normal polarity charge structures dominated by a main midlevel negative charge region, and so produce few, if any, +CG flashes.

The development of a lower negative region in the SP98 and RR simulations and in the observed storm is crucial for +CG development (Fig. 17). In the model, flashes initiate between two charge regions of opposite polarity, where the electric field magnitude tends to be largest. It then propagates bidirectionally toward and through the two charge regions. In both the simulations and the observed storm, almost all of the +CG flashes initiated near 6 km, between the main positive charge region and the lower negative charge region. Once initiated, the negative end of the simulated channel travels upward and branches through the main positive charge region, while the positive end travels downward through the smaller lower negative charge region to ground (e.g., Fig. 14). Similar behavior was inferred from lightning mapping observations of the 29 June storm by Hamlin et al. (2003). Positive CG flashes occur only after the formation of a lower negative charge region in the SP98 and RR simulations, a behavior that further reinforces the importance of the lower charge. [This behavior is similar to the behavior hypothesized by Williams et al. (1989) that −CG flashes occur only after the formation of a lower positive charge, as is true of the TAKA and GZ simulations.]

In the RR and RAR simulations, the positive channel of +CG flashes follows regions of negative charge that have descended almost to ground. Though such a charge distribution certainly enhances the probability of a channel propagating to ground, the necessity for this charge configuration in the simulations may be a result of the limitations described by Mansell et al. (2002). The model grid resolution is orders of magnitude larger than the cross-sectional dimension of a real lightning channel, so the tip of a simulated channel produces a much smaller enhancement of the ambient electric field than is produced by the tip of an actual channel. This reduction makes it more difficult for the simulated lightning to propagate beyond charge regions.

c. Positive CG production relative to storm microphysics and kinematics

As discussed above, the production of +CG flashes by the 29 June storm was caused by the storm having inverted-polarity electrical structure, with a lower negative charge below a larger, midlevel positive charge. Thus, those microphysical and kinematic properties of the storm responsible for producing inverted-polarity charge structure were also responsible for producing +CG flashes. However, all the simulations in this study have the same microphysical characteristics (e.g., liquid water and ice particle contents), so the different polarities of CG flashes in different simulations are due to the specific sensitivities of each noninductive charging scheme. With the RR and SP98 noninductive charging schemes, the large riming rates present in the mixed-phase region of the updraft are responsible for producing an inverted-polarity dipole and charging in regions of smaller riming rates is responsible in part for producing the lower negative charge. However, with the TAKA and GZ schemes, the simulated maximum cloud water content, though approaching adiabatic values, remains too low to support positive charging in the updraft region. Instead, the TAKA and GZ simulations maintain a normal polarity charge structure throughout the simulation.

Lang and Rutledge (2002) suggested that one of the factors leading to predominately +CG flashes in several midlatitude storms (one of which is 29 June) was an unusually large updraft volume. A large updraft volume inhibits mixing into the core of the updraft and so helps preserve large updraft speed and moisture supply. Indeed, the observed and simulated storms contain a total updraft volume consistently greater than 50 km³ through the −10°C level (Figs. 7c and 7d).

The observations showed that +CG flashes tended to cluster downshear from the main updraft region (Wiens et al. 2005), and this is mirrored by the SP98 and RR simulations. The clustering of +CG flashes downshear of the main updraft is probably caused by the presence
of descending graupel and hail in that region. MacGorman and Burgess (1994) and Stolzenburg (1994) found that the storms in which +CG flashes composed a large percentage of CG activity usually also produced large hail. Carey et al. (2003a) found that the +CG activity tended to cluster in the area of highest reflectivity during the 1998 Spencer, South Dakota, supercell, though no large hail was reported with this storm. During the 29 June storm, almost all of the +CG flashes were collocated with graupel and hail cores, probably due to positive charge within the radar-inferred hail region (Wiens et al. 2005). In the simulations, hail carries much less charge than graupel. Carey and Rutledge (1998) likewise inferred from observations of another storm that hail was not a major charge carrier. The simulations suggest that the onset of +CG flashes is associated with descending positively charged graupel, along with the still lower descent of negative graupel charged by both the noninductive and inductive mechanisms, as discussed previously. This relationship is probably what causes a correlation between the timing of melting graupel and rain (i.e., graupel meltwater) and the onset of +CG flashes (Figs. 9 and 10b).

Carey et al. (2003a) found that the +CG flash rate increased during or just after pulses in storm growth during the Spencer storm. Similar relationships are suggested during the SP98 and RR simulations, though it is difficult to determine a direct relationship with storm growth. The reason is that ground flash rates reflect both the general level of storm electrification and the conditions needed specifically for ground flashes. The rate of noninductive charging is controlled in all parameterizations by the rebounding collision rate of riming graupel with ice and snow, which depends, in turn, on the speed and size of the updraft. As discussed above, however, ground flashes require the formation of a lower charge region from noninductive charging of graupel that descends to lower altitudes along with inductive charging of graupel colliding with droplets in the downdraft region. If the conditions for ground flashes are met, the overall charging rate can modulate ground flash rates, but otherwise, ground flash rates will be at or near zero, regardless of the storms charging rate.

Cloud-to-ground activity does not begin in the simulations until the onset of rainfall, as inferred from the rain mass time series (Fig. 10b), because rainfall is related to the descent of graupel and inductive charging. Because conditions for ground flashes are met during the period of intense electrification in the simulations, the peak +CG rates of 3–5 min$^{-1}$ occur shortly after the maximum graupel volume is reached at 150 min, roughly when there was a short-lived maximum in the updraft mass flux and updraft volume. No relationship exists with peak updraft speed, because peak updraft speed was largely unrelated to the overall charging rate.

d. Total flash rate relative to storm microphysics and kinematics

The simulations produce total flash counts similar to the approximately 10 000 flashes observed over 3 h in the 29 June 2000 storm (Table 1). Flash rates tended to increase with time, as the storm grew larger, but shorter cycles were also superimposed on this linear trend. As observed in other storms by Lhermitte and Krehbiel (1979), MacGorman et al. (1989), and Goodman et al. (1988), the evolution of flash rates on 29 June was similar to fluctuations in convective intensity and precipitation growth. The long-term increase in graupel volume was well correlated with total lightning activity in both the simulations and the observed storm, as were trends in updraft mass flux and updraft volume (Figs. 9 and 10a). As graupel volume, updraft mass flux, and updraft volume increase, the number of collisions occurring between graupel and ice particles also increases. Consequently, noninductive charging and electrification increase and force the flash rate to increase to control the maximum electric field.

To examine whether the shorter-term cycles of these parameters also were correlated, the linear increasing trend was subtracted, and the correlation calculation was then repeated (e.g., MacGorman et al. 1989). The resulting unbiased correlation estimates indicate lightning flash rates are well correlated with the other storm parameters listed in Table 2. Graupel volume and total lightning are best correlated at approximately zero time lag. However, cycles in updraft mass flux and updraft volume tend to lead cycles in flash rates by 4–5 min. It is apparent in both the model simulations and the observed storm that the total flash rate closely tracks storm intensity. Thus, unlike ground flash rates and the maximum electric field magnitude, the total flash rate

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<td>Graupel vol ( (T &lt; 0^\circ \text{C}; \text{lag, 0 min}) )</td>
<td>SP98 0.770 RR 0.706 TAKA 0.721 GZ 0.808</td>
</tr>
<tr>
<td>Updraft vol ( (w &gt; 10 \text{ m s}^{-1}; \text{lag, 4 min}) )</td>
<td>0.732 0.653 0.705 0.787</td>
</tr>
<tr>
<td>Updraft mass flux ( (T = -20^\circ \text{C}; \text{lag, 5 min}) )</td>
<td>0.683 0.600 0.656 0.745</td>
</tr>
<tr>
<td>W-max (lag, 0 min)</td>
<td>0.091 0.090 0.098 0.075</td>
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appears to be an excellent measure of the evolving size and severity of the observed and simulated 29 June STEPS supercell storm.

Acknowledgments. The authors thank Kyle Wiens and Sarah Tessendorf at Colorado State University for providing LMA flash counts, multiple radar analysis figures, and results including graupel volume and updraft volume from the 29 June storm. Two anonymous reviewers provided comments that helped improve the manuscript. We also thank Erik Rasmussen for providing mobile mesonet observations for the 29 June case. Professor William Beasley at the University of Oklahoma offered many valuable discussions of an earlier version of this work. Support for this research was provided under National Science Foundation Grants ATM-0119398 and ATM-0340693. Additional funding for this research was provided under NOAA-OU Cooperative Agreement NA17RJ1227.

APPENDIX

Model Electrification

a. Inductive and noninductive charging

Inductive charging occurs in the presence of an electric field when a rebounding collision occurs between two polarized particles. In the model, inductive charging is only included during graupel–droplet collisions when graupel are in dry growth mode. Noninductive charging (i.e., independent of the electric field) occurs with rebounding collisions between riming graupel and ice particles in the presence of cloud droplets. The macroscopic spatial separation of opposite charges on cloud and precipitation particles from the combined effect of their differential fall speeds and wind shear subsequently generate fields strong enough to produce lightning.

Inductive charging in the model is calculated based on a formula from Ziegler et al. (1991). This equation is expressed by Mansell et al. (2005) in terms of characteristic diameter $D_G$ and mass-weighted mean fall speed $V_g$ of graupel as

$$\frac{\partial q_g}{\partial t} = \left(\frac{\pi^3}{8}\right) \left[ \frac{6.0V_g^2}{\Gamma(4.5)} \right] E_{\text{gc}} E_r n_{t,c} n_{0g} D_G^2 \times \left[ \pi \Gamma(3.5) \epsilon \cos(\theta) E \cdot dD \Gamma(1.5)n_{\text{gc}}(3n_{t,c}) \right].$$

(A1)

In Eq. (A1), $E_{\text{gc}}$ and $E_r$ are the collection and rebound efficiencies, $n_{t,c}$ and $n_{0g}$ are the total cloud water and graupel number densities, $n_{\text{gc}}$ is the number concentration intercept for graupel, $D_G$ is the cloud droplet diameter, $(\cos \theta)$ is the average cosine of the angle of the rebounding collision, $E_x$ is the vertical component of the electric field, $q_g$ is the charge on graupel, and $\epsilon$ is the permittivity of air. The inductive charging used in the present simulations approaches values described as “strong” by Mansell et al. (2005), with $E_x = 0.01$ and $(\cos \theta) = 0.45$.

Noninductive charging involving riming graupel and ice crystals has been the focus of several laboratory experiments. The charge gained by the graupel is dependent on the ambient temperature and the liquid water content as well as the size and growth state of the hydrometeors. The general formula for noninductive charge separation between colliding particles $x$ and $y$ is

$$\frac{\partial q_{xy}}{\partial t} = \int_0^\infty \int_0^\pi \frac{\delta q_{xy}(1 - E_x)}{4} |V_y - V_x|$$

$$\times (D_x + D_y)^3 n_x(D_x)n_y(D_y) dD_x dD_y, \quad (A2)$$

where $D_x$ and $D_y$ are the diameters of the colliding particles, $E_{xy}$ is the collection efficiency, $|V_y - V_x|$ is the relative fall speed, $n$ is the number concentration, and $\delta q_{xy}$ is the charge separated per collision. A representative weighted-average separated charge per collision $\delta q_{xy}$ replaces $\delta q_{xy}$. Thus, (A2) is simplified by permitting $\delta q_{xy}$ to be moved outside the integral (Mansell et al. 2005). The magnitude of $\delta q_{xy}$ is limited to a maximum of 50 FC for graupel–snow collisions and 20 FC for graupel–cloud ice collisions to prevent unrealistic charging and lightning rates. For this study, the model includes four different parameterizations of the noninductive charging process as each varies in determining the sign and level of charging.

b. Charge conservation, advection, and ions

A charge density is connected with every hydrometeor type. As mass shifts between categories in the microphysics, the charge also is transferred from one category to another (e.g., mass from ice to rain). Although charge is conserved within the model domain, charge is not globally conserved due to charge movement from ion currents entering or exiting the domain, advection through a lateral boundary, sedimentation to ground, or by cloud-to-ground lightning. The charge continuity equation from Mansell et al. (2005) resembles a typical conservation equation with treatment of advection, diffusion, and sedimentation. The model neglects the accelerations of charged hydrometeors in an electric field. The electric field is determined as the negative gradient of the potential: $E = -\nabla \phi$.

Conservation equations are defined for both positive and negative small ion concentrations (Mansell et al.
2005) proceeding from the method described by Helsdon and Farley (1987). The equations take into account advection and mixing, drift motion (ion motion induced by the electric field), cosmic ray generation, ion recombination, ion attachment to hydrometeors, corona discharge from the surface, and release of ions from evaporating hydrometeors.

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