Electrical and Polarimetric Radar Observations of a Multicell Storm in TELEX

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Abstract

On 28-29 June 2004 a multicellular thunderstorm west of Oklahoma City was probed as part of the Thunderstorm Electrification and Lightning Experiment (TELEX) field program. This study makes use of radar observations from the KOUN polarimetric WSR-88D, three-dimensional lightning mapping data from the Oklahoma Lightning Mapping Array (LMA), and balloon-borne vector electric field meter (EFM) measurements. The storm had a low flash rate (30 flashes in 40 min). Four charge regions were inferred from a combination of LMA and EFM data. Lower positive charge near 4 km and mid-level negative charge from 4.5–6 km MSL (0 to -6.5 °C) were generated in and adjacent to a vigorous updraft pulse. Further mid-level negative charge from 4.5–6 km MSL and upper positive charge from 6–8 km (-6.5 to -19 °C) were generated later in quantity sufficient to initiate lightning as the updraft decayed. A negative screening layer was present near storm top (8.5 km MSL, -25 °C). Initial lightning flashes were between lower positive and mid-level negative charge and started occurring shortly after a cell began lofting hydrometeors into the mixed phase region, where graupel was formed. A leader from the storm’s first flash avoided a region where polarimetric radar suggested wet growth and the resultant absence of non-inductive charging of those hydrometeors. Initiation locations of later flashes that propagated into upper positive charge tracked the descending location of a polarimetric signature of graupel. As the storm decayed, electric fields greater than 160 kV m\(^{-1}\) exceeded the minimum threshold for lightning initiation suggested by the hypothesized runaway breakdown process at 5.5 km MSL, but lightning did not occur. The small spatial extent (≈100 m) of the large electric field may not have been sufficient to allow runaway breakdown to fully develop and initiate lightning.
1. Introduction

The noninductive charging mechanism involving rebounding collisions between cloud ice particles and riming graupel is, at present, the leading candidate for the source of charge involved in lightning flashes within thunderstorms (Saunders 1993). Using parameterizations of this noninductive mechanism based on laboratory studies (Takahashi 1978; Saunders and Peck 1998), numerical cloud models have been able to reproduce realistic charge and lightning distributions (e.g., Mansell et al. 2005). Studies have also cited this charging mechanism as an explanation for inferred or measured distributions of storm charge (e.g., Stolzenburg et al. 1998; Takahashi and Keenan 2004).

Electrification by the non-inductive graupel-ice mechanism requires a cloud region where supercooled cloud water is present. Polarimetric radar can distinguish between liquid and ice phase precipitation, and the presence of liquid phase hydrometeors above the melting level is also a proxy for updraft strong enough to support supercooled cloud water. Observational studies employing polarimetric radar have looked at total graupel mass and updraft strength as predictors of initial electrification and lightning flash rate (Jameson et al. 1996; Bringi et al. 1997; Carey and Rutledge 2000), but have only recently begun to examine the details of the initiation and structure of lightning relative to the distribution of graupel (Wiens et al. 2005).

Also under investigation is the magnitude of electric field necessary to initiate lightning. Conventional dielectric breakdown requires larger electric field magnitudes than typically have been observed in storms (e.g., Marshall et al. 1995, 2005). Several studies have shown that energetic cosmic rays and runaway breakdown may serve to reduce the requisite field strength (e.g., Eack et al. 1996; Dwyer 2003, 2005; Gurevich and Zybin 2005), though this hypothesis is still somewhat
controversial.

Extensive data sets collected during the Thunderstorm Electrification and Lightning Experiment (TELEX) enable further exploration of these issues. TELEX took place during May and June of 2003 and 2004 in central Oklahoma, with the broad goal of relating storm electrification and lightning activity to storm structure and microphysics. Observational systems employed in TELEX included the KOUN polarimetric radar, a lightning mapping array, and mobile ballooning facilities. Fig. [I] shows a map of the fixed facilities used in TELEX.

This study presents polarimetric radar data, lightning mapping data, and thermodynamic and electric field soundings that were obtained from a small multicellular storm that occurred in central Oklahoma on 28-29 June 2004.

2. Instrumentation and Data Analysis

a. Instrumentation

The polarimetric KOUN radar in Norman, OK is identical to those deployed in the nationwide WSR-88D network ([Crum and Alberty](1993); [Klazura and Imy](1993)), except that it has been upgraded for polarimetry ([Doviak et al.](2000); [Ryzhkov et al.](2005b); [Scharfenberg et al.](2005)). During TELEX, the radar collected volume scans of conventional and polarimetric data every 5-6 minutes. Measured polarimetric variables were horizontal reflectivity factor $Z_h$, differential reflectivity $Z_{DR}$, correlation coefficient $\rho_{hv}$, and differential phase shift $\Phi_{DP}$. [Straka et al.](2000), [Bringi and Chandrasekar](2001), and [Zrnic and Ryzhkov](1999) provide thorough overviews of the theory and practice of radar polarimetry.

The Oklahoma Lightning Mapping Array (LMA) maps total lightning activity in three dimen-
sions within a storm. Thomas et al. (2004) describes in detail the operation and accuracy of the system. Briefly, the LMA detects the development of a lightning flash by receiving impulsive noise sources created in the VHF band by a leader as it propagates through the air. Complementing the LMA was the National Lightning Detection Network (NLDN) (Cummins et al. 1998), which detected cloud-to-ground lightning strikes. Field operations were restricted to within 75 km of the center of the LMA, where 3D mapping is most accurate; 2D mapping is available out to a nominal range of 200 km.

Balloon-borne instruments were flown into storms to provide in-situ measurements of storms’ electrical and thermodynamic properties. Pressure, temperature, relative humidity, and GPS position and winds were provided by a dropsonde system from the National Center for Atmospheric Research (Hock and Franklin 1999). The dropsonde was modified for upsonde operation through the addition of a simple baffle to redirect airflow past the temperature and relative humidity sensor elements.

Vector electric field \((E)\) measurements were provided by a balloon-borne electric field meter (EFM). MacGorman and Rust (1998, sec. 6.2.3v) discuss the history and operation of this instrument, first developed by Winn and Byerly (1975). For TELEX, the EFM was upgraded with more durable construction and on-board data recording to complement telemetered data. The EFM modulates \(E\) by spinning about its vertical and horizontal axes at nominal rates of 1 Hz and 3 Hz, respectively, thereby producing a sinusoidal raw electric field signal. A two-axis accelerometer and three-axis magnetic field sensor (each new for TELEX) provided improved reference signals for use in demodulation of the raw sinusoid to retrieve \(E\).
b. **Methodology**

The analysis presented here is focused on a 25-by-25 km horizontal domain centered 35 km west and 25 km north of KOUN (Fig. [I]). Data were visualized in virtual 3D space with custom display software developed in-house that allowed for efficient exploration of the data.

1) **Radar Data Processing and Interpolation**

The polarimetric KOUN data were processed to edit and unfold $\Phi_{DP}$, compute $K_{DP}$, correct $Z_{DR}$ and $\rho_{hv}$ for noise, and correct Z and $Z_{DR}$ for attenuation following the procedures described by Schuur et al. (2003) and Ryzhkov et al. (2005a). The data were then interpolated to a cartesian grid using a single-pass, isotropic Barnes-type inverse exponential weighting scheme (Koch et al. 1983; Trapp and Doswell 2000). Filtering was tuned to suppress wavelengths shorter than twice the maximum data spacing in the analysis domain. Maximum spacing was calculated for 50 km range and 1° azimuth and elevation angle spacing. A non-dimensional smoothing parameter of $\kappa^* = 0.1$ (Koch et al. 1983; Trapp and Doswell 2000) was chosen to preserve detail near the spatial Nyquist limit. Overall, these parameters preserve detail at low altitudes. Artifacts are present at higher altitudes, due to elevation angle spacing greater than 1°. Since most of the analysis in this study concerns altitudes below 7 km, these artifacts at upper levels were deemed an acceptable tradeoff relative to the detail at lower altitudes.

2) **Interpretation of Polarimetric Radar Signatures**

The interpretation of polarimetric radar data to infer microphysics is challenging. For a review, see Straka et al. (2000), whose definitions for hydrometeor classes and diameters ($D$) are followed
here. For this study, polarimetric signatures due to graupel \((0.5 < D < 5 \text{ mm})\) and small hail \((5 < D < 20 \text{ mm})\) are especially important due to their hypothesized participation in non-inductive charging process. From a polarimetric perspective, they are essentially indistinguishable (Straka et al. 2000), though graupel will be characterized by generally lower values of \(Z_h\) than small hail since \(Z_h \propto D^6\). \(Z_{DR}\) for graupel and small hail has a wide range of values, -0.5 – 3.0 dB. Negative values are associated with prolate orientation (Aydin and Seliga 1984). Positive values indicate oblate orientation, with smaller values indicating lower density, greater sphericity, canting, or tumbling (Bringi et al. 1986).

Graupel and small hail embryos may be produced at relatively low altitudes as an updraft pulse lofts raindrops in a column to temperatures as cold as \(-10^\circ\text{C}\) (Carey and Rutledge 2000; Takahashi and Keenan 2004), where they begin to freeze (Smith et al. 1999). This column of lofted drops is associated with elevated \(Z_{DR} > 1.0\) dB (Tuttle et al. 1989; Herzegh and Jameson 1992; Conway and Zrnic 1993; Hubbert et al. 1998) due to oblateness of the drops and \(Z_h > 30\) dBZ due to their moderate to large size. Freezing drops remain water coated (i.e., wet growth) due to the abundance of supercooled water within a vigorous updraft. As frozen drops fall below the altitude of 0 \(^\circ\text{C}\), melting maintains a coat of water and large values of \(Z_{DR}\) because the ice core within the melting hydrometeor supports a larger drop with greater oblateness than would be possible with completely liquid precipitation. Coexistence of water and ice phases in the same hydrometeor de-correlates the horizontally and vertically polarized radar channels (Tuttle et al. 1989; Herzegh and Jameson 1992; Conway and Zrnic 1993; Hubbert et al. 1998) and may reduce \(\rho_{hv}\) to as low as 0.9.

Graupel may also be present at colder temperatures where supercooled cloud water concentrations are low enough to support only dry growth of ice hydrometeors. It is expected that graupel might be mixed with crystals, snow, and aggregates at these cold temperatures. The dominant
hydrometeor class can be distinguished by comparing its vertical profile of $Z_{DR}$ with that from
the other class. Crystals, snow, and aggregates will be characterized by decreasing $Z_{DR}$ above the
melting level and in regions where large updrafts are absent. At the highest altitudes in the cloud,
horizontally oriented pristine crystals with high ice density return $Z_{DR} > 0.5$ dB. Aggregation as
crystals fall reduces particle density and leads to a more spherical shape, decreasing $Z_{DR}$ to $< 0.5$
dB (Meischner et al. 1991a,b). Melting snow is characterized by a local maximum in $Z_{DR}$ as ag-
gregation and water coating produce giant oblate hydrometeors. These break up into to smaller
raindrops with lower $Z_{DR}$ once melting is complete. By contrast, spherical and/or tumbling con-
cical graupel has lower $0.0 < Z_{DR} < 0.5$ dB (Aydin and Seliga 1984; Bringi et al. 1986) with no
height dependence. Mixtures that are the most dominated by graupel will have the lowest $Z_{DR}$.
In the melting region, $Z_{DR}$ for graupel increases more monotonically to its final rainy value than
does snow.

3) Charge identification using LMA data

Rust et al. (2005) and Wiens et al. (2005) infer storm charge from LMA data by assuming that
a lightning leader moves through charge of opposite polarity (MacGorman et al. 1981; Williams
et al. 1985; Coleman et al. 2003), thereby serving to neutralize space charge. Assignment of
leader polarity to detected LMA sources was accomplished by assuming that initial breakdown
in the storm is a bi-directional process (Kasemir 1960; Mazur and Ruhnke 1993; Thomas et al.
2001) with the negative leader preferentially detected (Shao and Krehbiel 1996), and that sources
associated with negative breakdown radiate greater power (Thomas et al. 2001). The path of a
negative ground strike is often mapped well to within 1 km of the surface and can be used in
combination with NLDN data to confirm inferences about polarity.

4) **Electric field data and charge identification**

Gauss’s law may be used with electric field measurements to infer charge at the measurement location:

\[ \rho = \varepsilon \nabla \cdot \mathbf{E}, \]  

where \( \mathbf{E} \) is the vector electric field, the permittivity of air \( \varepsilon = 8.86 \times 10^{-12} \text{ F m}^{-1} \), and \( \rho \) is space charge density. However, the EFM measures the vector electric field only along the flight path of the instrument, making it impossible to calculate the full divergence of \( \mathbf{E} \). Interpretation is further complicated by the fact that the \( \mathbf{E} \) profile is not an instantaneous measurement.

In spite of these difficulties, it is still useful to infer charge location and polarity from a subjective interpretation of the vector \( \mathbf{E} \) measurements (Rust et al. 2005). The vectors, plotted along the balloon’s path, are examined for convergent and divergent patterns that indicate localized positive and negative charge, respectively, to the side of the balloon. Regions where the vectors are parallel give information about the orientation of more-extensive regions of charge, and complement interpretation of \( \mathbf{E} \) with the one-dimensional approximation to Gauss’s law (e.g., Schuur et al. 1991).

3. **Storm lifecycle and lightning data**

Multicellular storms (Byers and Braham 1949; Marwitz 1972; Weisman and Klemp 1982) grew and decayed throughout the local evening of 28-29 June 2004 during TELEX. This paper
focuses on one cluster of cells that initiated at approximately 2300 UTC west of Oklahoma City and approximately 50 km northwest of KOUN. The 00 UTC sounding from Norman, OK (Fig. 2) on 29 June indicated that the storm developed in an environment of weak shear, with winds less than 5 m s$^{-1}$ up to 400 hPa (about 8 km MSL). The troposphere was rather moist throughout its depth. The CAPE was about 900 J kg$^{-1}$. These environmental parameters are consistent with the multicellular convective mode. This storm had a low flash rate. All thirty flashes produced by the storm took place between 0000 UTC and 0040 UTC on 29 June.

A few cells with reflectivity greater than 50 dBZ at 2.5 km MSL were arranged in a linear fashion at 2335 UTC (Fig. 3a). Another cell was present in the northeast part of the analysis domain. At 2340 UTC (Fig. 3b), a new cell developed in the gap between cells near the center of the analysis domain. By 2351 UTC (Fig. 3c) the reflectivity pattern had consolidated into a larger cell with maximum reflectivity near 60 dBZ, flanked by cells with reflectivity less than 50 dBZ; these cells maintained a linear alignment. At 0013 UTC (Fig. 3d), the cell from the northeastern part of the domain had migrated toward the main line of convective cells, and the maximum in reflectivity within the line had shifted westward by 5–10 km. By 0029 UTC (Fig. 3e) maximum reflectivity was less than 50 dBZ across the entire domain. Reflectivity values continued decreasing through subsequent volume scans (e.g., Fig. 3f).

a. 2330 UTC - 0000 UTC: Convective pulse

Radar reflectivity factor exceeded 40 dBZ at low levels in individual cells before 2351 UTC, but decreased rapidly above the melting level (4.5 km, 0 °C). The relatively shallow vertical extent of large reflectivities suggests that warm rain collision-coalescence processes and relatively weak
updrafts were dominant in the production of precipitation (Zipser and Lutz 1994).

The strongest convective cell at 2351 UTC presents a unique radar signature in comparison to the cells preceding it. This signature was located 30 km west and 24 km north of KOUN at 5.1 km MSL (Fig. 4). At this location each polarimetric variable exhibited an extremum relative to the surrounding storm. A maximum in $Z_h = 60–65$ dBZ, a minimum $\rho_{hv} = 0.9$, and $K_{DP} > 2.5^\circ$ km$^{-1}$ were associated with a column of enhanced $Z_{DR}$ that extended above the melting level and was 3–5 km in horizontal extent. The polarimetric data indicate water-coated, lofted frozen drops and small hail (Sec. 2.b.2). Lofting of drops above the freezing level is a process known to produce graupel embryos efficiently (Carey and Rutledge 2000; Takahashi and Keenan 2004). The vertical extent of large values of $Z_h$, $Z_{DR}$, and $K_{DP}$, above the melting level distinguished this cell from earlier cells, and indicated that the updraft was stronger in this cell.

b. 0000 UTC - 0008 UTC: Initial flashes

The first six lightning flashes in the storm occurred from 0001 UTC to 0008 UTC (Fig. 5). For each of these flashes, LMA and NLDN data indicate that a negative leader descended toward and struck ground. Within the cloud, lightning activity propagated into positive charge mostly below 5 km MSL (-1 °C) and negative charge mostly above that level. The lightning-inferred charges were arranged in a layered manner. The altitudes of inferred charge were slightly higher near the horizontal center of lightning activity, with a gradual downward slope of the layered charge with horizontal distance away from the region most dense with LMA sources. These two charge regions did not necessarily represent the complete electrical structure of the storm at that time, as the LMA only reveals charges that participate in lightning (Sec. 2.b.3).
Between 2351 UTC and the initiation of the first flash at 0001 UTC, the KOUN data indicated a microphysical transition above the melting level near the lightning initiation locations. The locations at which these six flashes initiated were clustered between 5 and 6 km MSL (-1 to -6.5 °C) and within 1–2 km of one another in the horizontal. The cluster of initiation points was located on what had been the northwest edge of the enhanced $Z_{DR}$ column and $\rho_{hv}$ minimum in the 2351 UTC volume (Fig. 4).

The column of enhanced $Z_{DR}$ collapsed by 2357 UTC (Fig. 4e–g), with $Z_{DR} < 1.0$ dB above the melting level. A local minimum in $Z_{DR}$ of 3-5 km horizontal extent developed between 5.5 - 6.5 km MSL (-4 to -9.5 °C), and descended and weakened somewhat by the 0002 UTC volume. The minimum value of $Z_{DR}$ was 0.2 dB. $\rho_{hv}$ was greater than 0.99 above the melting level, and showed no pronounced local minima. Overall, the microphysical state above the melting level had transitioned from one dominated by lofted, freezing drops to one indicative of dry graupel mixed with snow and ice (Sec. 2b2).

By three minutes before the first flash the frozen drop polarimetric signature seen in the 2351 UTC volume scan had descended to 2.8 km, at about 10 m s$^{-1}$ (Fig. 4e–g). The large fall speed is consistent with a signature dominated by melting frozen drops and small hail (Knight and Heymsfield 1983), as are the depressed values of $\rho_{hv}$ and $Z_h > 50$ dBZ (see Sec. 2b2).

Fig. 4e–g shows LMA sources from the first flash, plotted along with $Z_h$, $Z_{DR}$, and $\rho_{hv}$ from the 2357 UTC volume. The flash initiated above the melting level. A negative leader propagated downward toward the $\rho_{hv}$ minimum. As the leader neared the region of depressed $\rho_{hv}$ values, it turned to the southeast, propagating horizontally above the minimum. The leader eventually came to ground on the southeast side of the $\rho_{hv}$ minimum. Abrupt changes in the leader propagation direction suggest that local variability in the charge distribution was correlated with the water-
coated hydrometeors indicated by polarimetric data in the 2351 and 2357 UTC volume scans.

c. 0009 UTC - 0013:30 UTC: Addition of lightning in upper positive charge

Four lightning flashes from 0009–0013 UTC (Fig. 6) continued to propagate through negative charge above and positive charge near and below the melting level. The NLDN indicated ground strikes with two of these flashes. However, beginning at 0009 UTC, some flashes began to propagate through upper level positive charge instead. During this time interval and that discussed in the following section (ending at 0021 UTC), there were a total of six flashes involving upper level positive charge.

The NLDN reported a positive CG strike with the first flash to propagate into upper positive charge (at 0009:56 UTC, Fig. 6). However, analysis (not shown) of the LMA data indicates that the ground strike was coincident with the initial breakdown process, indicated by an upward propagating negative leader. Peak current reported by the NLDN was 5.2 kA. Cummins et al. (1998) state that flashes identified by the NLDN as low peak current positive CG flashes are usually misidentified cloud flashes, as this one appeared to be.

The mean altitude of LMA sources associated with activity in the upper positive layer decreased with each subsequent flash (Fig. 7). The apparent descent rate of these sources was calculated as follows. The vertical center of LMA sources inferred to have occurred in the upper positive charge was manually estimated at a series of times. A regression analysis of altitude vs. time of these locations produced a 3 m s\(^{-1}\) downward speed. The fit had a correlation coefficient of 0.99, indicating a good linear fit to the data, thus supporting the idea that upper level positive charge neutralized by the lightning flashes descended at 3 m s\(^{-1}\). This does not necessarily mean that the
upper positive charge was on graupel or other rapidly descending hydrometeors, a point that will be discussed further in Section 5.

Fig. 8 shows radar data from KOUN during the time of the six flashes that propagated into the upper positive charge layer. Two convective cells were present. The cell to the south was the mature to decaying stage of the cell formed by the convective pulse at 2351 UTC. Precipitation in this cell extended well above 0 °C.

In the 0008 UTC volume scan (Fig. 8), a local minimum in $Z_{DR} \approx 0.2$ dB was present at about 6.7 km MSL (-10 °C) and 26 km north of KOUN. By 0019 UTC this signature descended at 3 m s$^{-1}$ to about 5.3 km MSL (-4 °C), with its centroid moving roughly 3 km toward the south. $Z_{DR}$ below the minimum increased monotonically through the melting level to ground. $Z_{DR} > 0.5$ to the sides of this minimum at 7 km MSL (especially evident at 0008 UTC) suggests the presence of pristine crystals. $Z_{DR}$ decreased to as low as 0.3 dB below the crystals and above the melting level, suggesting the formation of aggregations. Within the melting zone, a local maximum in $Z_{DR}$ was consistent with melting aggregates. The measurements suggest precipitation sized ice throughout the region of the storm above the melting level, but the patterns in $Z_{DR}$ suggest that the minimum was associated with relatively high concentrations of spherical or tumbling conical graupel (Sec. 2b.2). The initiation locations of the flashes that propagated into upper positive charge track the location of the $Z_{DR}$ minimum in space and time.

d. 0013:30 UTC - 0025 UTC : New activity to west

Lightning flashes at 0011:40 UTC and 0013:10 UTC, which, like all previous flashes, initiated about 31 km west of KOUN, hinted that lightning activity might soon begin in a cell farther west.
Each of these flashes contained a descending, westward propagating negative leader at altitudes below 4 km. Both leaders followed nearly the same path. As each leader reached an altitude of 2.5 to 3 km, 37 km west and 24.5 km north of KOUN, it abruptly began propagating upward. The leaders terminated at an altitude of 4–5 km (near 0 °C), directly above where they turned upward, and can be seen protruding to the west of the most dense lightning activity in Fig. 6.

At 0013:46 UTC, a flash initiated in the western cell where the negative leaders from the eastern cell terminated. This flash initiated well west of flashes preceding it and was the first of eight flashes in the western extent of convection before 0025 UTC (Fig. 9). These western flashes, like the initial activity in the convection to the east, neutralized positive charge primarily below the melting level and negative charge above. There was no evidence of lightning that propagated into upper-level positive charge in the western activity through 0025 UTC. The NLDN indicated that five of the western flashes produced a negative CG; the LMA data indicate leaders propagating toward ground in these and two additional flashes.

Two more negative CG strikes (indicated by NLDN and LMA) were associated with flashes originating in the eastern cell. The negative leader from one of these flashes began in positive charge at 7 km MSL (-12.5 °C) and traveled westward 12 km, coming to ground west and well south of all other ground strikes (Figs. 7 and 9). The second negative CG was produced by a complex flash that appeared to propagate through the lower and upper positive and mid-level negative charges.
e. 0025 UTC - 0040 UTC: Diffuse flashes

The last seven flashes in the storm took place from 0025–0040 UTC (Fig. [10]). All flashes had leaders that moved through positive charge at upper levels and negative charge just above the melting level. Three of these flashes also exhibited leaders that appeared to move through positive charge below the melting level.

The NLDN reported that the last flash in the storm (at 0038:48 UTC) was a positive CG flash, with peak current of 8.5 kA, below the cautionary threshold set by Cummins et al. (1998). Unlike the flash at 0009:56 UTC, the reported CG strike was not associated with initial breakdown. Since the LMA does not map positive leaders well, it was not possible to confirm the ground strike. There is some possibility that a positive CG strike was associated with this flash.

During previous time intervals it was possible to define eastern and western groupings of flashes. Flashes during this time interval were characterized by more extensive propagation across the entire convective region. This was especially true of later flashes during this time interval. The leaders were indicated by fewer LMA sources. In comparison to flashes before 0025 UTC, the larger aerial extent of each flash and sparse tracing of the leaders lent a more diffuse look to the plot of LMA sources associated with these flashes.

4. In-situ measurements

At 0020:39 UTC, an electric field meter and radiosonde were launched into the eastern side of the high reflectivity region present at that time (see Fig. [3]). The electric field and radiosonde profile are shown in Fig. [11]. Figure [2] shows a skew-\(T\) plot of temperature and dewpoint temperature for this sounding.
a. Radiosonde data

Maximum vertical velocity of the balloon was 7 m s$^{-1}$ in a 1500 m deep layer above 5 km (Fig. 11). This ascent rate is somewhat larger than the nominal 5 m s$^{-1}$ rise rate of the balloon in calm air, and indicates that the balloon was in weak, and perhaps residual, updraft above the melting layer.

The radiosonde (Fig. 2), which was launched in rain, reported nearly saturated conditions from near the surface to the top of the sounding (near 10 km). Some details of the thermodynamic data offer clues about the altitude of cloud top. A downward jog was seen in the temperature profile at 8.5 km MSL (∼-25 °C), the altitude of radar echo top. Near 8.5 km (about 370 hPa), a slight drop in RH and increased variance in the RH profile were observed. Enhanced mixing with environmental air (which was still quite moist) may have been responsible for increased variance in RH; this contrasts with the smoother RH profile at lower altitudes associated with conditions more in equilibrium within the cloud.

A quasi-isothermal layer was observed from 4.8–5.3 km with a mean temperature $T = -1^\circ$C. It was warmer than a continuation of the near-moist-adiabatic profiles above and below the layer would suggest. A large-lapse-rate (adiabatic) layer was present above the quasi-isothermal layer (Fig. 2). Quasi-isothermal layers reported by Willis and Heymsfield (1989) and Shepherd et al. (1996) have $T \geq 0^\circ$C, a large adiabatic lapse rate below the quasi-isothermal layer, and are cooler than the surrounding moist-adiabatic profiles. The characteristics of the quasi-isothermal layer in this study do not appear compatible with the findings of these previous studies. Perhaps this layer was being warmed by release of latent heat (through freezing of supercooled water or deposition of vapor), though testing of this idea is beyond the scope of this study.
b. Electric field data

The gross electric field structure, interpreted by using the one-dimensional form of Gauss’s law, indicated up to six charge regions (Fig. 11b). Charge regions were examined without regard to cloud base because precipitation outside the cloud may carry significant charge. Positive charge was present just above the ground, followed by alternating positive and negative regions of space charge up to 8.3 km MSL. The magnitude of the electric field decreased to small values near and above 8.5 km MSL, in agreement with 8.5 km MSL being the upper cloud boundary as inferred from radiosonde and radar measurements.

Positive charge density from 0.5–1.8 km and 2.1–4.0 km MSL was low compared to the rest of the profile, consistent with the idea that positive charge was being carried on rain that was precipitating out of the storm. Only a few lightning leaders at this time appeared to propagate into the lower positive charge (2.1–4.0 km MSL, Fig. 10), unlike lightning before the balloon launch. Whether or not the hydrometeors carrying the charge were within or beneath the cloud, the charge history of the storm inferred from lightning activity and the EFM support the presence of a lower positive charge carried on precipitation.

Detailed analysis of the full vector data provided some added insight about possible origins of excursions in $E$ from 1.6–3.0 km MSL. $E_y \simeq E_z$ near 2.2 km (Fig. 11a), indicating that the application of 1-D Gauss may not have been appropriate. Below 2.0 km, $E$ pointed upward. As the balloon neared 2.0 km MSL, $E$ began to point downward and toward the south. $E$ continued to point southward after $E_z$ became positive again at 2.5 km MSL. During these $E$ excursions, the balloon flew through a region of large horizontal gradient in reflectivity, with a precipitation core to its north. The vector pattern above 2 km is consistent with positive charge carried on precipitation.
north of the balloon.

The signal near 2 km, as $E_z$ became negative and $E_y$ began to increase in magnitude, does not appear to be attributable to positive charge. The charge could have been deposited by lightning. A negative leader from a lightning flash at 0021:37 passed directly above the balloon at 3 km MSL, 6 min before the balloon reached 2 km. This gives an approximate charge descent rate of about 3 m s$^{-1}$, a fall speed typical of $< 1$ mm diameter raindrops (Beard and Pruppacher 1969; Brandes et al. 2002). Drop size distributions across a variety of precipitation regimes are dominated by these small drop sizes (Schuur et al. 2001; Bringi et al. 2003). Most of the ions generated by the lightning channel would attach to these small drops (Moore et al. 1964; Helsdon et al. 1992; Ziegler and MacGorman 1994). A charge region, generated by a transient lightning flash, that precipitated out of the storm would arguably not be representative of the charge structure of the storm on the time scale of the storm’s life-cycle. Ignoring this charge as unrepresentative of the overall storm charge distribution reduces the number of vertically separated charge regions to four.

The negative and positive charges inferred from the electric field between 4.0–7.3 km MSL are consistent with charge regions indicated by lightning activity near the time of the balloon flight. Negative charge indicated between 7.3–8.3 km MSL in the EFM data (-14 to -22 °C, Fig. 11) overlaps with positive charge inferred from the LMA data from 6–8 km (-6.5 to -19 °C, Fig. 10), but the difference in time of these two measurements was sufficient for the net positive charge to descend to below 7.3 km by the time of the EFM measurement. The uppermost negative charge was not penetrated by any mapped lightning channels and was probably a screening layer (Grenet 1948; Vonnegut et al. 1966; Hoppel and Phillips 1971; Klett 1972). At the time of the sounding, deep convective motions had subsided, leaving a relatively quiescent cloud top conducive to screening layer formation.
The EFM measured large horizontal components in E near 4 km. The largest values of \( E_h \) were measured between the second to last flash and the last flash in the storm. Field changes associated with the last three lightning flashes in the storm are indicated in the E profile in Fig. 11.

The field change due to the last flash is quite apparent in \( E_y \). LMA sources from this flash exist almost entirely south of the balloon. The balloon was at 4.5 km MSL (near 0 °C), the altitude where leaders moved through negative charge. E pointed toward this charge before the flash, and pointed in the same direction, but was reduced in magnitude, after the flash, consistent with reduction of net negative charge by lightning south of the balloon.

After the last lightning flash in the storm, a very large peak in the electric field was measured at 5.5 km (-3.5 °C, just above the quasi-isothermal layer). Temporary clipping of the sinusoidal \( E \) signal produced a gap in the E profile (Fig. 11) during peak |E| of at least 160 kV m\(^{-1}\).

Conventional breakdown requires |E| a few times greater than has been observed by many in-situ observations over several decades. The runaway breakdown hypothesis for lightning initiation requires smaller fields than conventional breakdown. These smaller fields are comparable to the larger values observed in storms. To initiate lightning by runaway breakdown, |E| must exceed some critical level \( |E_{rb}| \) over a region large enough to support formation of a self-propagating leader (McCarthy and Parks 1992, Gurevich et al. 1992, 1994, Roussel-Dupre et al. 1992, Dwyer 2003, 2005).

Dwyer (2003) gave a threshold \( |E_{rb}| = 284 \text{ kV m}^{-1} \) at STP (1013 hPa, 273 K), where atmospheric density \( n_0 = 1.29 \text{ kg m}^{-3} \). \( |E_{rb}| \) is scaled to other altitudes and conditions using the ratio \( n/n_0 \), where \( n \) is the atmospheric density at the other altitude (Dwyer 2005). The 29 June in-situ radiosonde data give \( n = 0.68 \text{ kg m}^{-3} \) at the altitude of maximum |E|, so \( |E_{rb}| = 150 \text{ kV m}^{-1} \). This threshold was exceeded over a depth of about 100 m as the balloon traveled about 70 m in hor-
orizontal distance, although the horizontal extent of large $|E|$ may have been greater. Dwyer (2003) found that runaway breakdown is semi-stable when $|E| > |E_{rb}|$ over finite length $L$, with complete instability not occurring for $L = 100$ m until $|E| \gtrsim 480$ kV m$^{-1}$. The EFM sounding appears to support this idea, since a lightning flash did not occur during or after the time $|E| \geq 160$ kV m$^{-1}$ was measured.

5. Interpretation of convective and electrical activity

The data presented so far described aspects of the cellular, microphysical and electrical evolution of the storm. These data are now integrated to describe the storm’s electrification.

Initially, convection was dominated by collision-coalescence (warm rain) processes. Although large reflectivities were present, these cells did not extend into the mixed phase region, and so did not produce lightning.

At 2351 UTC polarimetric radar data (Fig. 4) indicated that a stronger updraft pulse had lofted liquid precipitation above the melting level. Some hydrometeors froze and likely served as graupel embryos, and the stronger updraft also likely supported supercooled cloud water. Ten minutes later, six flashes initiated in a cluster near the location where the signature of lofted, freezing drops was observed in the polarimetric radar data. While some charging may have taken place before this convective pulse, it appears that it was the activation of the non-inductive graupel-ice mechanism by the stronger updraft that led to sufficient electrification to initiate lightning, consistent with other studies (e.g., Carey and Rutledge 2000, and references therein). Furthermore, the presence of graupel was inferred directly from the polarimetric data at the location and time of initiation of the first flashes, thereby lending credence to the idea that graupel carried one of the polarities of
charge neutralized by lightning.

Previous studies have also related initiation of lightning to updrafts strong enough to loft hydrometeors into the mixed phase region. This storm shares many similarities with storms in Alabama and Florida (Goodman et al. 1988; Bringi et al. 1997; French et al. 1996; Jameson et al. 1996), and with “Hector” convection in Australia (Carey and Rutledge 2000; Takahashi and Keenan 2004).

Convective pulses in both the eastern and western regions of lightning activity appeared to be immediately effective in initiating lightning only at lower altitudes in the storm (Fig. 5), though charge may have been present at higher altitudes. Flashes neutralized lower positive charge at \( \sim 4 \) km MSL (\( \sim 2^\circ\)C) and mid-level negative charge from \( \sim 4.5–6 \) km MSL (0 to -6.5 \(^\circ\)C). Graupel has the faster terminal fall speed of the two hydrometeor types involved in the non-inductive charging process, so it is reasonable to infer that the lower positive charge was carried by graupel or rain from melted graupel. Ice crystals participating in rebounding collisions were probably produced by the Hallett-Mossop process (Hallet and Mossop 1974). Strong updrafts likely provided abundant water for large riming rates at temperatures greater than -10 \(^\circ\)C. The positive charging of graupel is consistent with laboratory studies of non-inductive charging in the presence of large liquid water content and \( T > -10^\circ\)C (Takahashi 1978; Saunders and Peck 1998).

The lofted, frozen drops seen in the polarimetric data at 2351 UTC were water-coated and remained so as they fell below the melting level by 2357 UTC, as indicated by a region of low values of \( \rho_{hv} \) (Fig. 4). Negative leaders from a lightning flash propagated around this region (Fig. 4g), suggesting that the region was of lower (or neutral) charge density than the surrounding precipitation. Hydrometeors in wet growth are less conducive to the rebounding collisions necessary for non-inductive charge separation, which likely explains the observation that hydrometeors with a
history of wet growth contributed to a region with smaller net charge density.

As each convective cell matured, lightning developed at higher altitudes (Fig. 6) between positive charge from 6–8 km MSL (-6.5 to -19 °C) and mid-level negative charge from 4.5–6 km MSL (0 to -6.5 °C). At approximately the time lighting became active at upper levels, radar signatures suggested the absence of vigorous updrafts. Each subsequent flash neutralized net charge, so some regeneration or rearrangement of charge was necessary in the absence of vigorous updraft in order to repeatedly attain large enough electric fields to initiate subsequent flashes.

Upper level lightning was seen to initiate at the location of a descending minimum in $Z_{DR}$ (Fig. 8), where dry, spherical or tumbling graupel probably dominated polarimetric returns in a mix of precipitation sized ice. The $Z_{DR}$ minimum developed near storm top at $\sim$ 7 km ($\sim$ -13°C) and did not appear traceable to the graupel inferred just above the melting level immediately following the convective pulse. This suggests that graupel accumulated with time near storm top, as will be discussed later. The local correlation of lightning initiation with higher graupel concentration suggests that graupel played an important role in pushing the electric field past the threshold for lightning initiation. The faster fall speed of graupel (compared with ice crystals or snow) suggests it was a negative charge carrier between 4.5–6 km MSL (0 to -6.5 °C). However, it was probably not the only one, because flashing at lower altitudes also suggested that separation of charge near the melting level left a negative charge on ice crystals from 4.5–6 km.

Neutralization of charge by lightning occurs as ions generated by the lightning channel attach to nearby hydrometeors (Moore et al. 1964). The attachment occurs in proportion to the surface area of a given hydrometeor type (Helsdon et al. 1992; Ziegler and MacGorman 1994). As noted by Ziegler and MacGorman (1994), the surface area of ice crystals and aggregates probably greatly exceeds that of graupel in the negative charge region, so negative charge carried on the graupel
was not substantially neutralized by any of the lightning flashes. Instead, most positive charge from lightning was likely deposited on ice crystals and aggregates in the vicinity of the negatively charged graupel, which neutralized the mixture of graupel, ice crystals, and aggregates.

Negative ion flux from lightning to positively charged ice crystals was favored at upper levels, which neutralized net charge and reduced the total charge on all positively charged crystals, so that this region would tend not to be involved in subsequent flashes. A subsequent flash involving the negative graupel masked by the previous flash took place after sufficient time passed to allow the mixture of positively charged ice crystals and aggregates and negatively charged graupel to separate by differential sedimentation. This unmasking process made it appear that lightning activity within upper positive charge descended with time because previous flashes had redeposited some of that positive charge on ice crystals that were mixed with the graupel, which was itself descending with time.

The process of masking and unmasking of graupel requires that graupel became negatively charged in sufficient concentration before the initiation of lightning and remained sufficiently negative while it was masked by positive charge. [Ziegler et al. (1991)] describe how graupel concentration may increase via the interplay between the terminal fall speed of the graupel and the negative vertical gradient in vertical velocity at the top of an updraft. This process might have been reflected in the intensification of the graupel signature from 6–8 km MSL in the 0002 and 0008 UTC volume scans.

Radar reflectivity evolution suggests that the updraft was not particularly strong at the time of appearance of the graupel signature in $Z_{DR}$ and weakened with time, suggesting that most of the collisional, non-inductive charging had probably been completed by the time of the first upper-level flash. Nevertheless, it is possible that non-inductive generation of charge continued in the upper
part of the updrafts, especially in the region of higher graupel concentration where collision rates with ice crystals were higher. The non-inductive graupel-ice charging mechanism requires active riming supported by an updraft, and radiosonde measurements indicated an ambient updraft of 2 m s\(^{-1}\), which is sufficient to produce supercooled water necessary for riming (Rauber and Tokay [1991], Stith et al. [2002]). Enhanced charging due to this process could have played a role in favoring lightning initiation near the graupel signature in \(Z_{DR}\).

As upward vertical motion subsided throughout the storm, lightning activity broadened to include the entire area that had been electrically active. The distinction between eastern and western cells in the lightning activity disappeared. A negative screening layer was inferred from EFM data near storm top, at altitudes where positive charge had previously been indicated by lightning. The LMA did not map any channels at this altitude during the sounding, so the screening layer indicated in the EFM data could not be verified by the LMA.

In summary, four charge regions were inferred with a large degree of confidence from a combination of EFM and LMA data. Lower level positive charge was found below the melting level, followed by negative charge just above the melting level. Positive charge resided above this. Negative charge just within the upper cloud boundary was attributed to a screening layer.

It seems that the convective pulse at 2351 UTC and subsequent storm growth for the next 10–15 min were responsible for the generation of most of the charge that participated in lightning. However, rapid separation of charge led to lightning at low levels, while later separation of charge at upper levels delayed the onset of lightning there until the storm began to decay.

Stolzenburg et al. (1998) presented a conceptual model of charge structure in the convective region of thunderstorms. Theirs is the most recent and complete synthesis of observations seeking a generalized model. They examined supercells and MCSs over the Southern Great Plains of the
United States and multicellular convection over the mountains of New Mexico. They found two charge profiles that were related to the relative strength of vertical motion within each storm type. For multicellular convection having transient updraft pulses of small maximum magnitude, they distinguished between soundings in or near the center or core of the convection and those outside the core. The convective core soundings were found to have four charges, beginning with positive charge and alternating in polarity with height. Outside the core there were at least six charge regions, again beginning with positive charge and alternating in polarity with height. The electric field sounding in this storm was taken near the center of convection and exhibited four charge regions in the vertical. By these criteria, the storm agrees with their model.

However, the detailed radar and lightning history examined for this storm raises a few questions about the model. The key idea in the conceptual model is the storm-relative strength of vertical motion. Based on the descent of the height of constant reflectivity contours across the domain with time, it appears that the storm analyzed in this study lacked a strong updraft and was decaying at the time of the sounding. Because the storm had relatively weak vertical motion, a crude application of the conceptual model might lead to the expectation of six charges instead of the observed four. These concerns may unfairly stretch the intended applicability of their conceptual model, which seems most valid on the time scale of the whole storm. Analysis of this storm has revealed details of the development of the charge structure within individual cells that are outside the scope of the Stolzenburg et al. (1998) conceptual model.
6. Concluding Remarks

Observations have been presented from a multicellular thunderstorm occurring on 28–29 June 2004 in central Oklahoma. Measurements from a lightning mapping array, an in-situ electric field and thermodynamic sounding, and a polarimetric radar were employed.

The data constrain the lightning initiation problem. After the last flash in the storm, a large electric field was measured which exceeded the minimum threshold for runaway breakdown. However, the field did not exceed the absolute maximum threshold proposed by Dwyer (2003). Apparently because of its small spatial extent, $|E|$ remained within the semi-stable range defined in that paper. This magnitude of the field was the largest measured during TELEX, even though convective strength and flash rates were low compared to most other TELEX cases.

Evidence was presented in support of the idea that hydrometeors with a history of wet growth carried less net space charge than neighboring hydrometeors with a drier growth history. Rain was lofted by an updraft pulse, began to freeze, but remained water coated, then fell below the melting level. A negative lightning leader propagated around the minimum in $\rho_{hv}$ that indicated the water-coated ice particles.

A principal contribution of this study is observational evidence of the presence of local maxima in graupel concentrations that are well correlated in space and time with the initiation locations of lightning flashes. At upper levels, it was possible to detect graupel mixed with other ice particles in the storm by looking for a local minimum in the $Z_{DR}$ field. In this storm, it took a while for the minimum to develop even after the partially liquid-phase hydrometeors evidenced by the enhanced $Z_{DR}$ column at 2351 UTC had frozen. The minimum appeared to be associated with an increase in graupel number density as updraft speeds decreased with height. Fall speed calculations further
support the presence of graupel. Three lightning and EFM-inferred charge regions were consistent with those expected to be gained by graupel and ice in a non-inductive charging process. A fourth charge region, an apparent screening layer near cloud top, was also inferred from the EFM data.

The inferred charge structure for this storm agreed in polarity and vertical order with the conceptual model of Stolzenburg et al. (1998). Charging was due to two different kinematic regimes. Lower level charge was generated in and adjacent to a vigorous updraft pulse, while upper level charge sufficient to initiate lightning was generated later as the updraft decayed. This study has shown that it is possible to relate charging and lightning activity to convective motions on the space and time scale of the development of individual cells. Future studies of this nature across the spectrum of thunderstorm intensity would be useful in clarifying the origin and location of charge within thunderstorms.

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Flight 41, SW Yukon, OK
29 June 2004
Launch: 0020:39 UTC

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