7.4

LIGHTNING AND POLARIMETRIC SIGNATURES OF TWO ELECTRIFIED WINTER STORMS IN CENTRAL OKLAHOMA

Kristin M. Kuhlman* and Kevin L. Manross

Cooperative Institute for Mesoscale Meteorological Studies, Norman, Oklahoma
National Severe Storms Laboratory, Norman, Oklahoma

ABSTRACT

This study examines two winter storms and associated lightning activity in central Oklahoma in 2007 and 2009. These events produced significant ice and snow totals as well as blizzard conditions across the state. Total VHF lightning data from the Oklahoma Lightning Mapping Array (LMA) and cloud-to-ground (CG) data from the National Lightning Detection Network (NLDN) are examined for each of the events. The flash rate, charge structure and CG lightning polarity are compared to precipitation amounts and signatures from the KOUN WSR-88D polarimetric radar located in Norman, OK. Individual lightning flashes, though infrequent during most of the events examined, quite often extended more than 15-20 km in length. This was possibly due to an increased build-up of charge over an extended period of time before lightning initiation. The differential reflectivity, correlation coefficient, and specific differential phase signatures of regions containing VHF sources are examined to determine if the lightning occurred only in mixed-phase regions and if there are differences in the ice crystal alignment before and after a lightning flash.

1. INTRODUCTION

In the coming years, forecasters will have the opportunity to routinely view polarimetric data from the WSR-88D radar network as well as view total lightning data from satellite. These two platforms may be able to provide additional information about winter storms, including determination of mixed phase regions as well as focusing attention on regions of enhanced lift and heavier precipitation.

Little study has been done concerning winter thunderstorms in the US, more detailed work comes primarily from Japan and Russia. In study of winter storms on the coast of the Sea of Japan, Fukao et al. (1991) determined that lightning was associated with regions of graupel mixing with smaller ice and snow particles. Similarly, Maekawa et al. (1992) found that lightning was produced in winter storms only after growing graupel was detected at the -10C level near a region small crystals. Winter storms that were strongly electrified and produced lightning had graupel or other large ice hydrometeors present, but storms that had no or very small electric fields typically did not show the presence of graupel (Kitagawa and Michimoto 1994). Through what is currently understood about the process of storm electrification (e.g., noninductive or relative growth rate), we expect this to be true (Saunders et al. 2006). Unfortunately, most studies of US winter storms have been limited to cloud-to-ground (CG) lightning detection networks for verification of lightning activity in storms with surface temperatures below freezing.

Due to the shallow nature of winter storms, lightning in these storms typically propogates much closer to the ground than during spring and summer thunderstorms. For this reason, it is believed a higher percentage of the lightning activity involves a CG connection, with a higher percentage of positive polarity flashes during the winter months (Orville and Huffines 2001). The presence of +CG lighting may be linked to a main positive, lower negative charge structure of the storm with flashes having a higher peak currents than summer, possibly due to stronger electric fields and longer periods between individual flashes in these storms (Brook et al. 1982; Brook 1992; Orville and Huffines 2001).

Anecdotal reports commonly discuss heavy snowfall rates or intensity and size of flakes associated with times of lightning or "thundersnow," though CG lightning rates may not necessarily be directly related to measured accumulation of snowfall (Crowe et al. 2006). Strong synoptic-scale forcing as well as a warm layer at low to mid-levels likely play vital roles in the electrification of winter storms, due to the necessity of mixed phase microphysics. Like all convective storms, electrified winter storms obviously need a combination of moisture, lift, and instabilty with

**Corresponding author address:* Kristin M. Kuhlman, National Weather Center, 120 David L. Boren Blvd, Norman, OK 73019; email: kkuhlman@ou.edu

the addition below freezing temperture at the surface. Polarimetric data combined with total lightning data can determine where the active charge layers lie in respect to precipitation particle size, shape, and ice density.

2. DATA

The main tools used to investigate lightning in this study are the Oklahoma Lightning Mapping Array (LMA) and the National Lightning Detection Network (NLDN). As described in Krehbiel et al. (2000) and MacGorman et al. (2008), the LMA maps total lightning of thunderstorm in three dimensions. As a lightning flash propagates it emits very high frequency (VHF) radiation; the LMA uses a time of arrival technique synchronized by the Global Positioning System to locate these sources. The Oklahoma LMA consists of 11 antennas spaced 10-22 km apart and centered approximately 30 km west of Norman, OK. The time and three dimensional location of each source is determined by the difference in the time-of-arrival between pairs of stations. Hundreds to thousands of points may be mapped by the LMA for any given lightning flash. Over the period of a few minutes the LMA can give detailed maps of the total lightning activity for the storm.

Radar data is from the Norman, Oklahoma (KOUN) polarimetric WSR-88D. KOUN is located approximately 40 miles southeast of the LMA network center and is a S-band radar operating with simultaneous transmission of the horizontal and vertical pulses. Data for the 2007 case was collected by the National Severe Storms Laboratory in research mode. The 2009 event was collected during a system test by the Radar Operations Center under control of contractor L3/Baron. Quality control measures were performed manually after data collection in post-processing.

This study uses the following polarimetric variables: Reflectivity, Differential reflectivity, specific differential phase, and correlation coefficient. A detailed explanation of these and other polarimetric variables can be found in Straka et al. (2000). Those used in this study will also be discussed briefly below.

Reflectivity: This study uses the common value for reflectivity from the horizontally polarized wave (Z_h). The reflectivity factor is proportional to the hydrometeors cross-section over a volume.

Differential reflectivity: Differential reflectivity (Z_{dr}) is the ratio of the reflectivity factor of the horizontally polarized wave (Z_h) and the vertically polarized wave (Z_v).

Specific differential phase: The specific differential phase (K_{dp}) is the measurement of propagation effects of the signal in precipitation. Anisotropic hydrometeors produce different phase shifts for horizontally and vertically polarized waves; K_{dp} typically increases as both the oblateness and dielectric constant increase. K_{dp} values in pristine ice may be used to determine if a cloud is electrified, due to icecrystal orientation by a strong electric field (Weinheimer and Few 1987; Krehbiel et al. 1996). However, especially for S-band radars, K_{dp}-based ice-crystal orientation signatures can be easily masked by larger ice hydrometeors (Carey et al. 2011)

Correlation coefficient, ρ_{hv}: The correlation of the horizontally and vertically power returns. ρ_{hv} is often used as an indicator of regions of mixed precipitation and non-meteorological targets.

3. EVENTS

a. 14 January 2007 Freezing Rain/Sleet

This event actually began on 12 Jan 2007 as freezing rain for southeast Oklahoma as a cold front pushed into Texas and the first upper-level short-wave moved across the area. Significant icing occurred early in this event before transitioning to largely sleet for central Oklahoma by 14 Jan 2007 with a deep cold layer at the surface and additional lift provided by a second low-amplitude short-wave trough moving across the region. Little lightning was observed by the LMA during this event until this second period of precipitation on 14 Jan 2007. The lightning was located within in convective cells, coincident with regions higher reflectivity, and heavy sleet accumulation (exceeding 0.2 in hr^{-1}) at the ground. The sounding from Norman, OK at 12 UTC on 14 January 2007 depicts a warm layer of above freezing temperatures between 900 and 600 mb (Fig. 1). The cold layer at the surface continued to remain in place throughout the day and all ground reports during this event were primarily of sleet, needle crystals, and freezing rain.

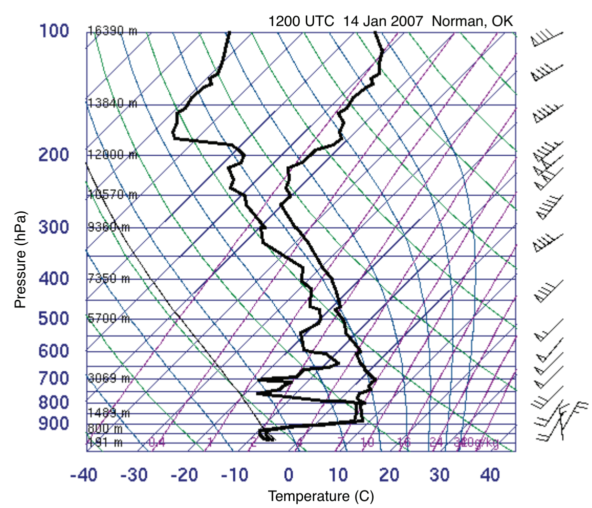

FIG. 1. Skew-T–logp plot of temperature and dewpoint from the 1200 UTC sounding by the Norman, OK National Weather Service Forecast office on 14 January 2007. Plot provided by the University of Wyoming online archive.

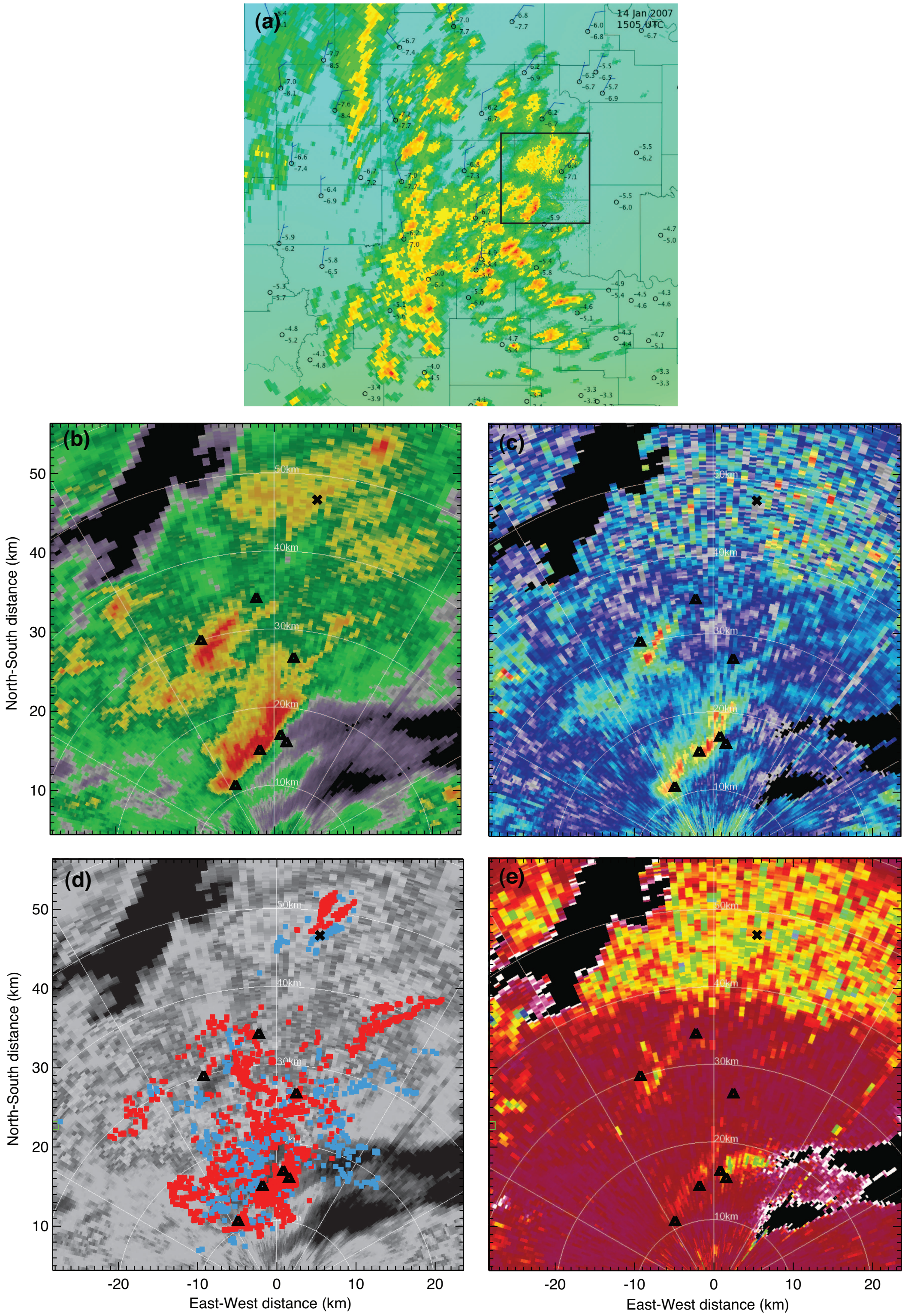

FIG. 2. (a) Horizontal reflectivity from KTLX at 0.5 degree elevation angle and Oklahoma mesonet station plots at 1505 UTC. Black box indicates zoomed region in plots b-e. (b) Horizontal reflectivity from KOUN at 3.0 degree elevation angle (KOUN is located at [0, 0]) and NLDN strike locations, -CG flashes are indicated by triangles and +CG flashes by crosses. (c) same as (b) but for Zdr. (d) LMA VHF source points, by charge. Red squares are associated with regions of positive charge, blue negative. (e) Same as (b) but for correlation coefficient.

The first lightning evident in the OKLMA with this event was associated with small storm cells in Comanche and Cotton counties in southwest Oklahoma between 1300-1400 UTC on 14 Jan 2007. At first, flashes occurred at a rate of 1 to 2 every 5 min, but by 1340-1350 the cells in Comanche county produced over 20 flashes in a ten minute period, with 12 containing a negative connection to ground. Surface temperatures in this region remained between -6 and -7 degrees Celcius, with a mixture of freezing rain and sleet reported at the surface. Due to the distance from KOUN, no radar evaluation was done for this time period.

By 1430 UTC, the precipitation moved into central Oklahoma including Grady, McClain, Cleveland and Oklahoma counties. The focus for this study during this time period is on the storm moving from the southwest across these counties beginning at 1445 UTC (see Fig. 2). This main cell had impressive growth as it moved into Oklahoma county from Cleveland County around 1510 UTC, both in height and size as well as flash rate. As it moved through McClain county at 1450 UTC, the peak reflectivity was 50 dbZ reaching 2-2.5 km, by 1510 as it moved into Oklahoma county, the 50 dBZ reached over 4 km with a peak of near 65 dBZ at 3.5 km. A Z_{dr} column also became evident at this time (Figs. 2c and 3c).

The flash rate with this storm also increased over the same time period. The total flash rate increased from 5-7 flashes per 10 min from 1450-1510 to 25-30 flashes over 10 min after 1510 UTC, averaging 4-5 flashes per min after 1515 UTC. This peak activity was contained primarily within the highest values of horizontal (above 40 dBZ) and differential reflectivity (greater than 3.5 dB). However, individual leaders did travel from one cell to other nearby cells often extending over 30 km in horizontal distance. These larger flashes often occurred during time periods of less frequent lighting activity or initiated in a region further away from the highest reflectivities of an individual cell, however, individual leaders typically remained in regions of at least 18-20 dBZ.

A charge analysis of the LMA VHF activity leads to a normal tripole charge structure for the core of the cell, with the area above 4 km dominated by positive charge and a negative charge layer between 2.5 to 3.5 km coincident with the melting layer as evident in ρ_{hv} (Fig. 3). Flash initiations were most common in the region between these top two charge layers (3-4 km). However, flashes containing a -CG component were typically initiated lower in the cell, around 1.5 to 2.5 km and included breakdown through a lower positive charge region. For the few +CG flashes that occurred in this region, they remained away from the storm cores, initiating at the height of the melting layer (Fig. 3). As the sleet storm began to weaken in northeast Oklahoma county after 1530, all charge seem to remain in the lowest two regions of the storm and all flashes after 1530 contained a -CG component.

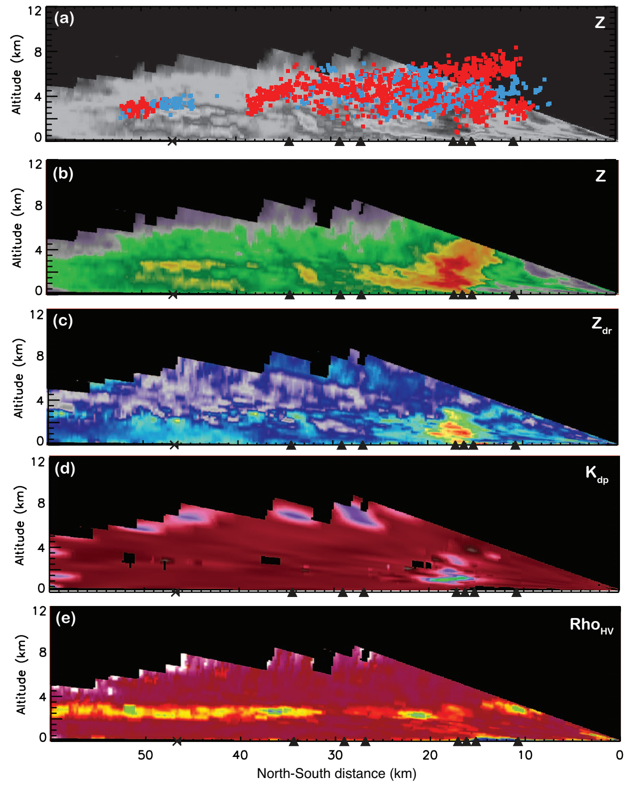

FIG. 3. North-south cross-section of KOUN polarimetric variables along region shown in Fig. 2b. (a) Horizontal reflectivity from KOUN (grayscale) and LMA VHF source points, by charge. Red squares are VHF source points associated with a region of positive charge, blue negative. (b) Horizontal reflectivity (c) differential reflectivity (d) specific differential phase (e) correlation coefficient.

By 1800 UTC, frozen precipitation was still occurring in central Oklahoma, it remained a mix of both sleet and freezing rain, though none of the cells were nearly as intense as earlier in the day. Lightning activity continued with these clusters of precipitation, though by this point at a much lower frequency than in storms previously discussed. Flash rates with the clusters at this time had reduced to one or two flashes every 5-10 min, focused in regions with at least 30 dbZ horizontal reflectivity. The focus on the analysis during this time period is examining evidence of crystal alignment in south central Oklahoma with these clusters.

Multiple flashes showed evidence of altering the crystal orientation. Flashes occurring between to consecutive degree sweeps (e.g., 3.5 and 4 degrees) often depicted altered Kdp values. As mentioned previously, in regions of pristine ice, a strong vertical electric field can vertically align crystals such that Kdp values would become negative (assuming no masking by larger ice hydrometeors). As seen

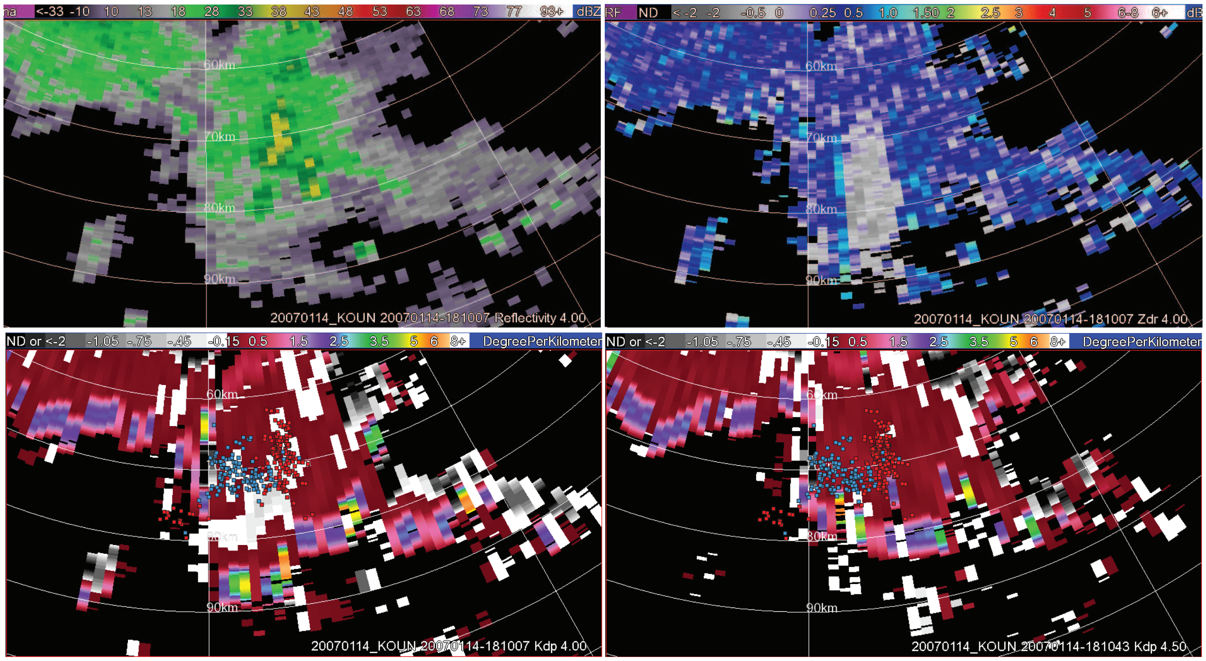

FIG. 4. Horizontal reflectivity (top), differential reflectivity (top-right), specific differential phase (bottom left) at 1810 UTC 14 Jan 2007, 4 degree elevation angle and specific differential phase at 4.5 degree elevation scan. Lightning flash occurs between the 4 and 4.5 degree elevation scans at 1810:31.9 UTC. LMA VHF source points from this flash are shown in both of the bottom panels.

in Fig. 4, a region of negative Kdp existed at 70 km from the KOUN radar at the 3.5 and 4.0 degree elevation angles prior to the flash occurring at 1810:31 UTC, but had become positive after the flash at the 4.5 degree elevation angle. Numerous flashes occurred during this time period and many exhibited the same behavior. Obviously, not all flashes matched times of consecutive KOUN elevation angles and we did see evidence of regions of negative Kdp during one volume scan that were not seen on next volume scan, without a coincident lightning flash occurring over the period.

b. 24 December 2009 Blizzard

Blizzard conditions affected the vast majority of northern, central, and southwest Oklahoma, and all of western north Texas, for at least 5 to 7 hours on 24 Dec 2009. Strong dynamic-lift occurred across Oklahoma as an intense upper-level system ejected northeast out of west Texas and moved across the region. A warm layer of above freezing temperatures remained in central Oklahoma throughout the day, seen in the Norman soundings from 0600 through 1800 UTC on 24 Dec (Fig. 5). However, strong vertical lift through the dendritic crystal growth zone allowed for heavy snowfall to still reach the surface, with periods of accumulation in central Oklahoma exceeding 1-2 in hr^{-1} during the event. With the strong surface low tracking just south and east of the region, surface winds remained strong and increased throughout the day with sustained winds of at least 15-20 m s^{-1} and gusts exceeding 30 m s^{-1}.

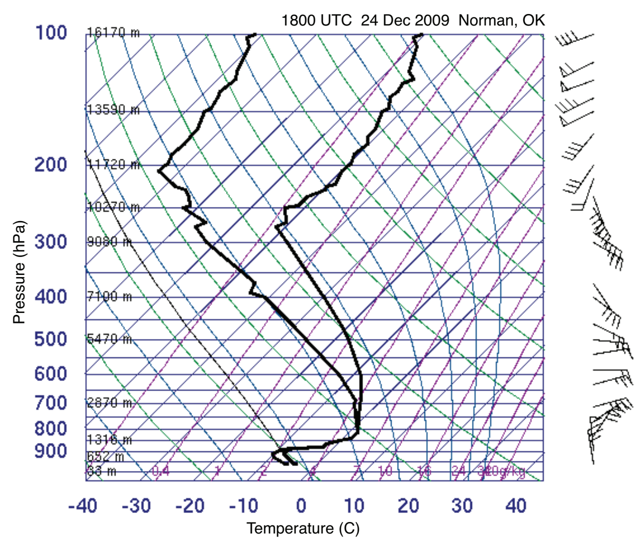

FIG. 5. Same as Fig. 1, but for 1800 UTC on 24 December 2009.

As strong as this system and forcing was, very little lightning occurred over the course of this event. Early in

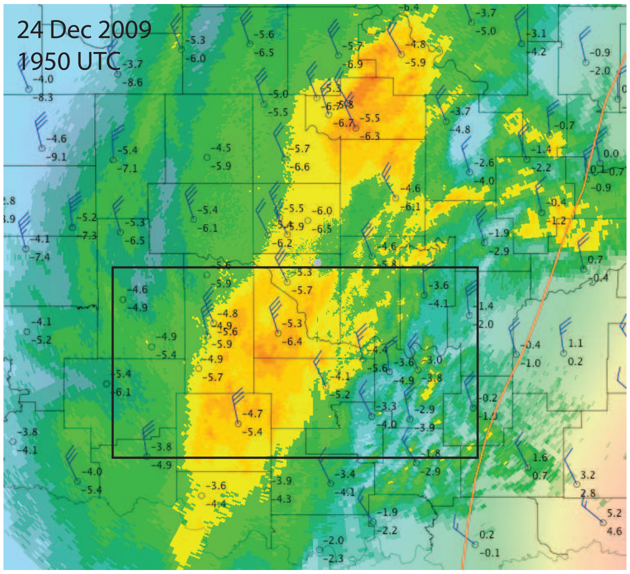

FIG. 6. Horizontal reflectivity from KTLX at 0.5 degree elevation angle and Oklahoma mesonet station plots at 1950 UTC. Black box indicates zoomed region in Figs. 7 and 8.

the day, during a transition period from heavy rain to snow, individual lightning were evident in OKLMA VHF data from Garvin, Carter and Murray counties between 1224-1244. The next lightning occurred, at 1400 UTC in Carter and Murray counties in south-central Oklahoma as snowfall rates began to intensify. Unfortunately during the early part of this event the KOUN radar was not operational. The next lightning activity that occurred during this event are three individual flashes all traversing the same region in southwest McClain county during the 10 min period between 1949 and 1958 UTC. Through the remainder of this event only 3 more flashes were observed by the Oklahoma LMA: in Oklahoma county at 2012 UTC; in Cleveland county at 2214 UTC, and in Tulsa and Okmulgee counties at 2341 UTC. Our analysis will focus on the 3 flashes occurring in a 10 min period in McClain county in central Oklahoma.

Each of the flashes that occurred during this 10 min period were contained within 0-5 km layer, which included a brief warm inversion to above freezing from 1.4 through 1.9 km as seen in the 1800 UTC sounding from Norman, OK (Fig. 5). The first and longest of these three flashes initiated at 1950 UTC at the southwest end of McClain county (lat: 34.88, lon:-97.65) at the height of this melting region (roughly, 1.4 km), with breakdown continuing primarily northeast towards the radar through the region of highest reflectivity (Fig. 7). The second flash occurred four minutes later at 1954:16.5 UTC. This flash initiated near the same point of the first, very slightly northeast (lat: 34.9, lon: -97.55), also between 1 and 1.5 km in height. Though a clustering algorithm considered this a single flash, it may have been two separate flashes the second overlapping in time, but initiating slightly to the southeast approx 2 km away. The third and smallest flash of this cluster occurred at 1958 UTC. This flash also began in southwest McClain county with continued breakdown towards the northeast. Similar to the other flashes, it initiated around 1km (lat: 34.94, lon: -97.53). Though all of these flashes were contained within the 0-5 km region, with evidence of VHF source points on or near the surface, no NLDN CG flashes were reported in conjunction with the LMA breakdown.

These flashes occurred during a time of blizzard conditions at the surface with 15-20 m s^{-1} sustained winds and gusts over 25 m s^{-1}. The surface temperature was less than -5 C and snowfall rates were greater than 1 in per hour with visibility of less than 0.10 miles at the surface (Fig. 6). The lighting seems to have occurred in what appears to be a ”transition zone” situated between the mixed-phase and sleet to the immediate east (as seen in Fig. 7b) and snow to the immediate west. The region of snow is characterized by a swath of high (upwards of 48 dBZ) horizontal reflectivity, Z_{dr} near 0.0-0.1 dB, and relatively high (near 0.98) ρ_{hv} (Fig. 7). Near where the flashes initiated, there appears to be a pocket of liquid precip present. A thin layer of ice crystals between approximately 4.0-4.5 km MSL is seen in the northwest half of the radar domain and the southern end of this crystal growth zone overlays the ”transition zone” swath where the lightning occurred seen in the enhanced values of Zdr and correlation coefficient above this region (Fig 8). There appears to be a band of graupel that moved from west to east, on the southern end of this transition zone in between about 2.5 - 4.5 km MSL. However, there is a strong northerly wind through much of the layer and the majority of the upstream hydrometeors still appear to be snow, albeit larger aggregates.

4. CONCLUSIONS

As seen in previous studies, a mixed phase region and strong dynamic forcing was present for both of these events. For the storms that occurred on 14 Jan 2007, liquid water was evident (ρ_{HV}, K_{dp}, Z_{dr} column) for those cells containing lightning. The peak area of lightning activity and initiations typically co-located (4-5 km) with the highest horizontal and differential reflectivity. Charge analysis depicted a normal tripole structure, with a main negative charge region right at and above the melting layer. CG flashes were predominately negative polarity and initiated between this main negative charge and a smaller lower positive charge region. Positive CG flashes did also occur, but were less numerous and typically initiated away from storm cores at the height of the melting layer.

Crystal alignment is suspected cause of decreased Z_{dr}

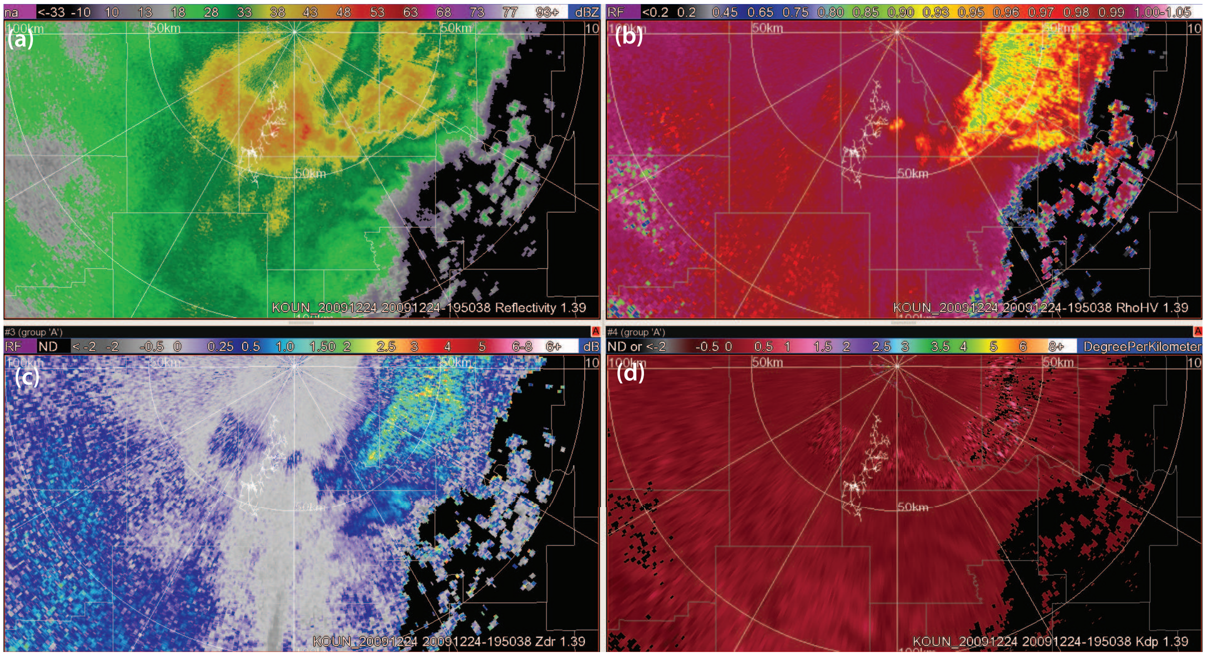

FIG. 7. (a) Horizontal reflectivity, (b) correlation coefficient (c) differential reflectivity, and (d) specific differential phase at 1950 UTC 24 Dec 2009, 1.39 degree elevation angle. Lightning flash occurring at 1950 UTC shown on all panels in white.

and negative K_{dp} values, coincident with lighting in Garvin Co on 1750-1840 UTC. Negative Kdp values were seen in PPI scans preceding a flash, with that region becoming positive in the PPI scan following a flash. More cases need to be examined to determine how frequently this phenomenon is seen with S-band radars, as it is likely to be more commonly observed by smaller wavelength radars.

Three flashes that occurred between 1950-2000 UTC on 24 Dec 2009 were located in what seemed to transition zone between snow to west, graupel to the southeast and sleet further east. The initiation location of these flashes seemed to coincide with a temporary region of increased lift and enhanced dendritic crystal growth above 5 km.

5. ACKNOWLEDGEMENTS

KOUN post-processing done with the help of Terry Schuur and John Krause. Data on 14 Jan 2007 collected in ”research mode” (operated by NSSL scientists) – QC performed manually after data collection. Data from 24 Dec 2009 collected during ROC ”System Test” under control of contractor L3/Barron. We acknowledge and appreciate L3/Barron and the Radar Operations Center for allowing us to use these data. We also appreciate Genna Langum and Brice Coffer for initial LMA data mining.

REFERENCES

Brook, M., 1992: Breakdown electric fields in winter storms. *Res. Lett. Atmos. Electr.*, **12**, 4752.

Brook, M., M. Nakano, and P. Krehbiel, 1982: The electrical structure of the Hokuriku winter thunderstorms. *J. Geophys. Res.*, **87**, 12071215.

Carey, L. D., W. A. Petersen, and E. V. Schultz, 2011: Dual-polarimetric signatures of ice orientation for lightning prediction: A radar modeling study of ice mixtures at X, C and S bands. *Fifth Conference on Meteorological Applications of Lightning Data*, Amer. Meteor. Soc., Seattle, WA, 9.6.

Crowe, C., P. Market, B. Pettegrew, C. Melick, and J. Podzimek, 2006: An investigation of thundersnow and deep snow accumulations. *Geophys. Res. Lett.*, **33**, doi:10.1029/2006GL028214.

Fukao, S., Y. Maekawa, Y. Sonoi, and F. Yoshino, 1991: Dual polarization radar observation of thunderclouds on the coast of the Sea of Japan in the winter season. *Geophys. Res. Lett.*, **18**, 179–182.

Kitagawa, N. and K. Michimoto, 1994: Meteorological and electrical aspects of winter thunderclouds. *J. Geophys. Res.*, **99**, 713–721.

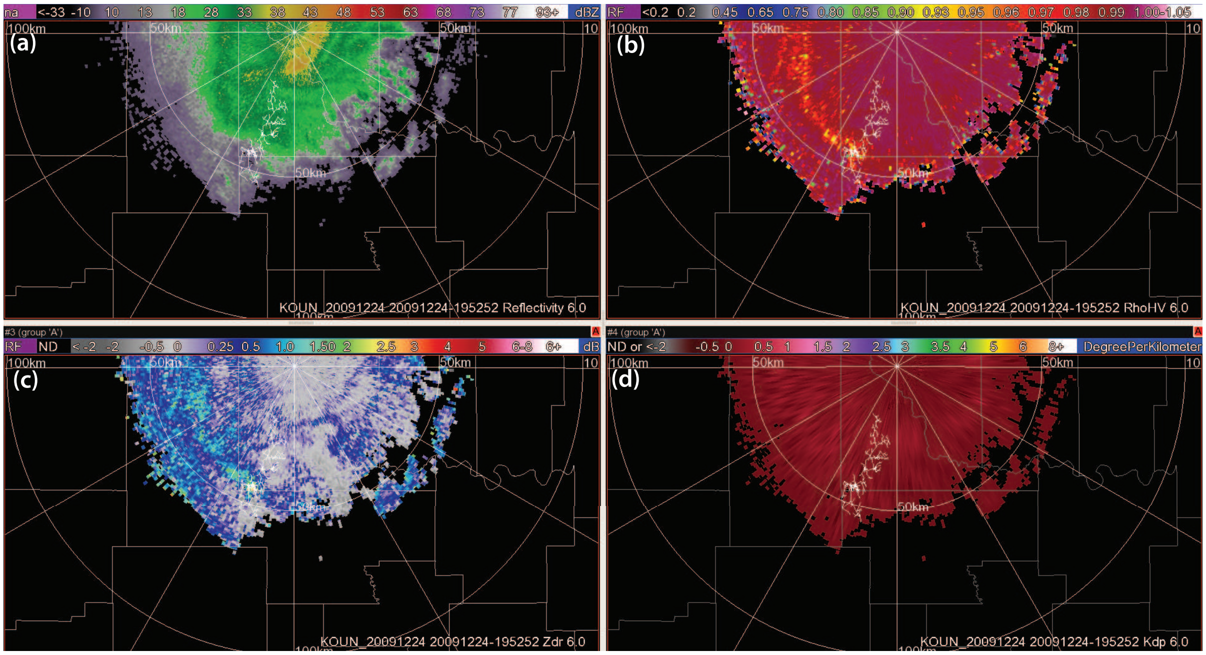

FIG. 8. Same as Fig. 7, but for 6 degree elevation angle at 1952 UTC.

Krehbiel, P. R., T. Chen, S. McCrary, W. Rison, G. Gray, and M. Brook, 1996: The use of dual channel circular-polarization radar observations for remotely sensing storm electrification. *Meteor. Atmos. Phys*, **59**, 65–82.

Krehbiel, P. R., R. J. Thomas, W. Rison, T. Hamlin, J. Harlin, and M. Davis, 2000: GPS-based mapping system reveals lightning inside storms. *Eos, Trans. Amer. Geophys. Union*, **81**, 21–32.

MacGorman, D. R., W. D. Rust, T. Schuur, M. E. Biggerstaff, J. Straka, C. L. Ziegler, E. R. Mansell, , E. C. Bruning, K. M. Kuhlman, N. Lund, J. Helsdon, L. Carrey, K. Eack, W. H. Beasley, P. R. Krehbiel, and W. Rison, 2008: TELEX: The thunderstorm electrification and lightning experiment. *Bull. Amer. Meteor. Soc.*, **89**, 9971013.

Maekawa, Y., S. Fukao, Y. Sonoi, and F. Yoshino, 1992: Dual polarization radar observations of anomalous wintertime thunderclouds in Japan. *IEEE Trans. Geosci. Rem. Sens.*, **30**, 838–844.

Orville, R. E. and G. R. Huffines, 2001: Cloud-to-ground lightning in the United States: NLDN results in the first decade, 1989 - 98. *Mon. Wea. Rev.*, **129**, 11791193.

Saunders, C. P. R., H. Bax-Norman, C. Emerisic, E. E. Avila, and N. E. Castellano, 2006: Laboratory studies of the effect of cloud conditions on graupel/crystal charge transfer in thunderstorm electrification. *Quart. J. Roy. Meteor. Soc.*, **132**, 2653–2673.

Straka, J. M., D. S. Zrnić, and A. V. Ryzhkov, 2000: Bulk hydrometeor classification and quantifications using polarimetric radar data: Synthesis of relations. *J. Appl. Meteor.*, **39**, 1341–1372.

Weinheimer, A. J. and A. A. Few, 1987: The electrical field alignment of ice particles in thunderstorms. *J. Geophys. Res.*, **92**, 14833–14844.