A Multiscale Overview of the El Reno, Oklahoma, Tornadic Supercell of 31 May 2013

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(Manuscript received 12 November 2014, in final form 4 March 2015)

ABSTRACT

On 31 May 2013 a broad, intense, cyclonic tornado and a narrower, weaker companion anticyclonic tornado formed in a supercell in central Oklahoma. This paper discusses the synoptic- and mesoscale environment in which the parent storm formed, based on data from the operational network of surface stations, rawinsondes, and WSR-88D radars, and from the Oklahoma Mesonet, a Doppler radar wind profiler, Rapid Refresh (RAP) analyses, and photographs. It also documents the overall behavior of the tornadoes and their relationships to features in their parent supercell based on data from a nearby, rapid-scan, polarimetric, mobile Doppler radar. The supercell formed near the intersection of a cold front and a dryline in an environment of moderately strong vertical shear and high CAPE, at the southern end of a line of multicell convective storms. The tornado damage path was as wide as 4.2 km according to the NWS damage assessment and ground-relative Doppler velocities of at least 135 m s$^{-1}$ were found at the theoretical beam height of 20 m AGL. The tornado debris signature in the copolar cross-correlation coefficient $r_{hv}$ was as wide as 4–5 km. After the strong tornado formed, at least one additional cyclonic tornado formed and rotated cyclonically around the main tornado; it was then absorbed by it and the main tornado broadened. Smaller subvortices, which rotated cyclonically around a common axis of rotation, were subsequently observed. The tornado then weakened but remained broad, while the anticyclonic tornado formed to the southeast along the rear-flank gust front.

1. Introduction

On 31 May 2013 an unusually broad and very intense tornado formed in a supercell in central Oklahoma, resulting in eight deaths (http://www.srh.noaa.gov/oun/?n=tornadodata), including those of several storm chasers (Wurman et al. 2014). The storm subsequently produced flash flooding in the Oklahoma City metropolitan area, resulting in 13 additional deaths. A mobile, rapid-scan, polarimetric Doppler radar from the University of Oklahoma collected volumetric data while ordinary-cell convective storms first formed and then evolved into the supercell, and also during the genesis-to-mature stages of the main, cyclonic tornado. In addition, for a short time period, the radar documented the multivortex structure of the tornado. In addition, a subsequent, rare, companion anticyclonic tornado was documented. The cyclonic tornado during its mature stage and the anticyclonic tornado were probed at a relatively close range of ~4–6 km (from the centers of the vortices; the range to the outer portion of the tornadoes was even closer). The cyclonic tornado is referred to as the El Reno tornado, owing to its proximity to El Reno, Oklahoma.

This severe weather event had a high public profile, in part because it was the first time veteran storm chasers were killed by a tornado and because members of a well-known crew from the Weather Channel covering the tornado live were injured when their vehicle was overturned. In addition, the National Weather Service...
Forecast Office in Norman, Oklahoma, measured the damage path to be as wide as 4.2 km or more, arguably tagging it as the widest tornado damage path ever documented. The extremely large and violent tornado that affected primarily rural areas of El Reno in central Oklahoma followed just 11 days after the devastating, Moore, Oklahoma, tornado of 20 May 2013 (e.g., Kudrzo et al. 2014) and other significant tornadoes in central Oklahoma on 19 May 2013. Estimates of the intensity of the tornado were controversial because the greatest damage found by the National Weather Service was rated only at as category 3 on the enhanced Fujita scale (EF3), while data from the mobile rapid-scan X-band (3-cm wavelength) polarimetric radar (RaXPol; Pazmany et al. 2013) indicated wind speeds as high as 135 m s$^{-1}$ or greater where the bore-sighted beam height was <20 m AGL (Snyder and Bluestein 2014), well above the category 5 on the enhanced Fujita scale (EF5) wind threshold of 90 m s$^{-1}$. Although the issues of how to make use of mobile Doppler radar measurements in rating tornadoes on the enhanced Fujita scale are beyond the scope of this paper (see Snyder and Bluestein 2014), the measurements of wind speeds by RaXPol and other mobile radars (Wurman et al. 2014) have prompted considerable informal discussion. Finally, it could be useful to analyze the event from the perspective of hindsight to see if anything can be learned that would increase the forecasting accuracy of future events.

For all of the aforementioned reasons, we believe that an examination of the event from synoptic-, meso-, and storm-scale perspectives is warranted. In addition, this overview paper will serve as a reference for more detailed storm- and tornado-scale analyses of the mobile radar data collected and for numerical studies involving the data assimilation of radar and other data. The purpose of this paper is therefore to provide a synoptic- and mesoscale context for forthcoming finer-scale analyses of the event. A description of the data types is given in section 2. The synoptic- and mesoscale environments along with a description of storm formation and evolution are found in section 3. The relationship between storm organization and tornado behavior is examined in section 4. Section 5 contains a summary of the findings and a guide for future studies.

2. Data

a. Operational network

Data used in this study came in part from the operational, national network of rawinsonde sites and from the Weather Surveillance Radar-1988 Doppler (WSR-88D) network (Crum and Alberty 1993). The former data were accessible in real time from the University Corporation for Atmospheric Research/National Center for Atmospheric Research/Research Applications Program (UCAR/NCAR/RAP) website and some of each was obtained later from the NOAA Storm Prediction Center (SPC) archive and the National Climatic Data Center (NCDC). Surface data were obtained from the Oklahoma Mesonet (Brock et al. 1995; Fiebrich and Crawford 2001) archive and images were prepared using WeatherScope software. Doppler wind profiler data were obtained online from the NOAA archive of the National Profiler Network (Weber et al. 1990; http://www.profiler.noaa.gov/npp/index.jsp). Satellite imagery was obtained in real time from the College of DuPage “NeXt Generation Weather Lab.”

Data were also accessed online from NCDC for Rapid Refresh (RAP) model analyses (Pan et al. 2014; http://rapid-refresh-rap). The polarimetric variables differential reflectivity $Z_{DR}$ and copolar cross-correlation coefficient $\rho_{hv}$ are computed. These variables can be useful in identifying the type of scatterers in the radar volume; in particular, they can be used to distinguish between debris (typically, $Z_{DR} < 1$ dB and $\rho_{hv} < 0.90$) and precipitation (typically, $Z_{DR} > 1$ dB and $\rho_{hv} > 0.90$) in tornadoes (e.g., Ryzhkov et al. 2005; Bluestein et al. 2007a; Snyder et al. 2010; Bodine et al. 2013; Snyder and Bluestein 2014; Wurman et al. 2014).

For this case, most of the time when the tornado was close to the radar, the radial sampling resolution was 30 m, although several minutes of data were collected using 15-m radial sampling resolution. For the latter, only eight pulses were used to calculate radar moments, which is considered marginal for obtaining low-variance estimates (e.g., Doviak and Zrnic 1993; Bringi and Chandrasekar 2001). There is a trade-off between the
number of quasi-independent samples obtained and the radial resolution as a consequence of the relationship between available system bandwidth, decorrelation time of the signal, and the dwell time for each radial. Snyder and Bluestein (2014) have shown how in tornadic debris clouds, where the $p_{nv}$ is often very low, data from the horizontal and vertical channels have only a low degree of correlation, so that the effective number of independent samples for pulse-pair covariance processing is increased when data from each channel are averaged before the radial velocity is estimated.

During the time that Doppler velocity data are described in subsequent sections, the Nyquist interval was $\pm 30.8\text{ m s}^{-1}$. Dealing with the velocities was relatively easy early on, but became much more difficult when multiple vortices were apparent. Based on temporal and spatial continuity considerations, we believe the dealt with the velocities are reliable.

Although the antenna half-power beamwidth is $\sim 1^\circ$, some smearing in azimuth occurs as a result of the rapid rotation rate of the antenna. RaXPol is often configured with a dwell time that yields radials every $1^\circ$. For rotation of the antenna at $180^\circ\text{ s}^{-1}$, the effective azimuthal resolution is $\sim 1.4^\circ$–$1.5^\circ$ [see (7.34) in Doviak and Zrnic (1993) and see Pazmany et al. (2013)].

Other special radar data came from KCRI, NOAA’s Radar Operation Center’s legacy radar, a test bed WSR-88D, located near University of Oklahoma Westheimer Airport in Norman (Saxion et al. 2011).

c. Data processing and display

Radar data were edited and displayed using Solo II software from the Earth Observing Laboratory (EOL) at NCAR or Warning Decision Support System–Integrated Information (WDSS-II) (Lakshmanan et al. 2007) software. RAP data were displayed using WDSS-II.

3. Synoptic- and mesoscale environment, and storm formation

a. Mesoscale features

Prior to the formation of convective storms in Oklahoma on 31 May 2013, at midday, a weak front was situated over northwestern Oklahoma (Fig. 1a). This front was marked by a wind shift from south, south-southwest, or south-southeast ahead of the front to northeast behind the front, and temperatures were $\sim 5^\circ\text{C}$ cooler northwest of the wind-shift line marking the front. By 2100 UTC, the winds ahead of the front had backed to south-southeasterly (Fig. 1d) and then southeasterly (Figs. 1e,f).

In the northern Texas Panhandle, the air was much cooler than to the south and the winds were relatively light (not shown). A dryline was located in far southwestern Oklahoma, where surface dewpoints were $\sim 5^\circ\text{C}$ drier than ahead of it. The boundary between the dryline and the front extended westward through the Texas Panhandle (not shown). The front weakened (as measured qualitatively by the horizontal temperature gradient seen by the change in colors) with time where it abutted against the warmer air mass to the southeast (Figs. 1b–e; the temperature difference across the front in north-central Oklahoma was $\sim 4^\circ$–$5^\circ\text{C}$ at 1800 UTC and only $\sim 1^\circ$–$2^\circ\text{C}$ or less at 2200 UTC), while it strengthened with time in western Oklahoma, where surface temperatures rose into the upper $30^\circ\text{C}$ behind the dryline (the temperature difference across the front–dryline border was $\sim 7^\circ\text{C}$ or less at 1800 UTC and $\sim 10^\circ\text{C}$ at 2200 UTC). The warmest air at the surface lay in a tongue oriented in the northeast–southwest direction from west of the dryline in western Texas to east of the dryline in northwestern Texas and far southwestern Oklahoma (not shown). The frontolysis
along the eastern portion of the front must have been due mainly to differential diabatic heating, as the air was heated by insolation more rapidly behind the front than ahead of it, because deformation alone should have increased the temperature gradient; the frontogenesis where the dryline abutted against the front must also have been due mainly to differential diabatic heating, where the air was heated more on the warm, dry side, than to the rear of the front; changes in surface temperature on either side of the front could not be accounted for by horizontal advection alone. The meridionally oriented dryline became sharper with time during the afternoon, as surface dewpoints dropped to below 10°C in the hot air, while surface dewpoints remained above 20°C in the moist air mass. The wind shifted to a more westerly direction in far western and southwestern Oklahoma, to the west of the dryline boundary and also just ahead of the front, consistent with the observed (not shown) and inferred formation of a meridionally oriented trough in low pressure where the air was being heated most rapidly. At 2100 UTC (Fig. 1d), prior to storm formation, from the wind field it is seen that qualitatively there was some cyclonic shear along the front, and weak convergence along the front ahead of the dryline.
b. Storm formation and evolution

It is seen in satellite imagery that convective storms first began to form along the front in extreme southeastern Kansas and later in north-central Oklahoma, as evidenced by small anvils (Fig. 2a) at 2115 UTC. A line segment of convective clouds appeared along and south of the front along the surface thermal ridge (Figs. 2a and 1d) and produced a first echo northwest of El Reno (Fig. 3b). This line segment extended southwestward along the axis of the warmest air behind the dryline as a zone of more sparse convective clouds. The most concentrated line segment, in western Oklahoma, consisted mostly of highly sheared cumulus congestus (Fig. 4). Another zone of convective clouds extended to the south-southwest behind the dryline in the deeply mixed air along the axis of warmest air (Fig. 1d). Cloud bands of stratus oriented north-northwest to south-southeast were seen south of the front in the moist air mass, indicative of very stable conditions. Bands of clear-air

FIG. 2. GOES visible satellite imagery for Oklahoma and small portions of surrounding states, every ~10–15 min, depicting storm initiation, organization, and development into mature convective storms. Shown are results from (a) 2115, (b) 2132, (c) 2145, (d) 2155, (e) 2215, and (f) 2240 UTC 31 May 2013. Images were obtained in real time from the College of DuPage website. The approximate location of the front (solid white line) and dryline (dashed white line) are indicated in (a).
FIG. 3. Storm initiation, organization, and evolution (a)–(f) prior to the development of supercells (every ~15 min), and (g)–(i) as supercells evolved (every ~30 min), as depicted by images of radar reflectivity factor (dBZ; color scale is at the bottom), at a 0.5°-elevation angle, from KTLX. Shown are images from (a) 2114, (b) 2128, (c), 2144, (d) 2201, (e) 2215, (f) 2229, (g) 2301, (h) 2328 UTC 31 May, and (i) 0001 UTC 1 Jun 2013. Range rings are plotted every 50 km. Zoomed-in view of the northern and southern fine lines as depicted by radar images of (j) radar reflectivity factor (dBZ) and (k) Doppler velocity (m s⁻¹) at a 0.5°-elevation angle, from KCRI at 2232 UTC 31 May 2013. Color scales for reflectivity and Doppler velocity are shown.
echoes oriented in the north-northwest to south-southeast direction were also noted on radar (Fig. 3).

By ~2130 UTC, anvils began to emanate from the cumulus congestus, which had grown into narrow cumulonimbus clouds, in western Oklahoma (Figs. 2b and 4b). These incipient storms, which were marked by several radar echoes oriented west-northwest to east-southeast or east–west (Fig. 3c), were located near the axis of maximum equivalent potential temperature at the surface (Fig. 5). By 2145 UTC, a broken line of convective storms had extended up to the north-northeast, along the front (Fig. 3c), while new, smaller cells had begun to form to the south, near the dryline (Figs. 1e and 3c). The storms near the dryline were more widely scattered and not as concentrated along a line segment as the storms to the north (Fig. 2d). By ~2200 UTC, a line of convective towers appeared to the northeast, along the front, northeast of the previously existing convective storms (Figs. 2d and 3d). At this time, the storms to the north took on the appearance of a broken line of convection, while the weaker storms to the south took on the appearance of a more widely scattered set of quasi-equally spaced, narrow radar echoes oriented in a northwest-to-southeast direction (Fig. 3e). About this time, RaXPol was deployed just north of El Reno, where the broken line of convection along the front was clearly visible (Fig. 4c, left of center to right; Fig. 2e), as were the more widely scattered storms along the dryline to the south (Fig. 4c, far left; Fig. 2e).

Between 2200 and 2300 UTC, the quasi-stationary front began to move southward as a cold front (Figs. 1e,f). To the northeast of the dryline, the front strengthened, apparently in response to evaporative cooling of precipitation from the convective storms in unsaturated air on the cool side of the front. Strong outflow winds of 20 m s$^{-1}$ were reported at one Mesonet station (Fig. 1f). However, the portion of the front near and to the north of the hot, dry air mass behind the dryline behaved differently. The wind shift propagated to the south, while the zone of strong horizontal temperature gradient that marked the location of the front remained stationary, behind the wind-shift line. Thus, the wind-shift line in western Oklahoma took on the appearance of a prefrontal wind-shift line, not a front (Hutchinson and Bluestein 1998; Schultz 2004). It might be that the apparent movement of the eastern portion of the front and wind-shift line to the west was caused by outflow produced by the convective storms. At 2229 UTC, two fine lines in the radar imagery appeared to the west and southwest of the broken line of convective storms along the front (Fig. 3f). The fine lines were tracked by reflectivity imagery from the KCRI radar (Figs. 3j,k), which at the time was providing low elevation scans every ~2–2.5 min, which is twice as often as usual, owing to a test of the Supplemental Adaptive Intra-Volume Low-Level Scan (SAILS) technique that was being conducted (http://www.roc.noaa.gov/wsr88d/NewRadarTechnology/NewTechDefault.aspx). The northern fine line, which was associated with approaching Doppler velocities, was moving to the south, while the southern fine line was quasi-stationary. The northern fine line was therefore...
probably associated with the cold front, while the southern fine line was probably associated with the dryline.

c. Synoptic- and mesoscale environment

The location where storms were first initiated to the west and northwest of El Reno (Figs. 2b,c; 3c,d; and 4) was marked by a local maximum in equivalent potential temperature (Figs. 5a,b). The sounding nearest in time and space from the storms was at Norman, to the southeast of the front, ~2–3 h after storms had initially formed (Fig. 6a). The moist boundary layer was ~100 hPa deep and was surmounted by a capping inversion, above which it was relatively dry (the relative humidity was ~30%–50%) up to the level of anvils from the storms. The convective temperature for a well-mixed boundary layer was ~38°C, which was not attainable at Norman [convective inhibition (CIN) was ~100 J kg\(^{-1}\)]. With both this convective temperature and the observed temperature, convective available potential energy (CAPE) was greater than 3000 J kg\(^{-1}\), which is relatively high for tornadic-supercell environments [e.g., Rasmussen and Blanchard (1998), their Fig. 13]. Farther to the west, along the dryline, convective temperature had been reached, as evidenced by the storms growing there in an environment of high temperatures (near 38°C) and high dewpoints; convective temperature may not have been reached to the north along the front, where lift may have been necessary to bring air parcels to the level of free convection (LFC).

The RAP model (Pan et al. 2014; http://www.ncdc.noaa.gov/data-access/model-data/model-datasets/rapid-refresh-rap) analysis (not shown), however, indicated the 100-hPa mean-layer/mixed-layer convective temperature was only ~28–30°C northwest of Oklahoma City along the front and these temperatures were reached by mid-afternoon. The highest mixed-layer CAPE at the time of storm initiation (~2200 UTC) was in excess of 5000 J kg\(^{-1}\) in a localized area just to the south-southwest of El Reno (Fig. 5c) while the mixed-layer convective inhibition was less than 100 J kg\(^{-1}\) and near zero in a swath extending to the south-southwest of El Reno.
FIG. 6. Upper-air data in central Oklahoma, near the time that the storms were forming/in progress on 31 May 2013. (a) NWS sounding from the Norman WSR-88D (KOUN) at 0000 UTC 1 Jun 2013, released when the El Reno tornado was in progress to the northwest and displayed as a skew (temperature)–log (pressure) diagram. Temperature (°C) is given by the red line, dewpoint (°C) is given by the green line, and winds are plotted to the right (half barb = 2.5, whole barb = 5, and pennant = 25 m s⁻¹). Hodograph is inserted at the upper left; wind speeds are labeled in knots (kt; 1 kt = 0.51 m s⁻¹). (b) Time (UTC; abscissa)–height (km; ordinate) diagram depicting hourly evolution of the vertical wind profile to the southwest of Norman at Washington, Oklahoma (from the “Purcell profiler”). (c) EVAD from RaXol data collected from volume scans averaged at the first deployment site, just north of El Reno, at five time periods, from 2220 to 2228 UTC; rings shown every 10 m s⁻¹.
The vertical shear in the lowest 6 km was ~25–30 m s^{-1}, which is well in the range supportive of supercells (e.g., Markowski et al. 2003; Thompson et al. 2003, 2007; Mead and Thompson 2011). The sharply curved hodograph at low levels and greater than 400 m^2 s^{-2} 0–3-km storm-relative helicity is supportive of tornadic supercells (e.g., Markowski et al. 2003; Thompson et al. 2003, 2007; Mead and Thompson 2011). Based on data from the nearby profiler southwest of Norman, it is seen that winds backed to southeasterly at ~2100 UTC and picked up in speed thereafter, while the winds at 6 km increased in speed to 25 m s^{-1} by 0000 UTC 1 June 2014 (Fig. 6b). The vertical wind shear profile and thermodynamic environment were therefore supportive of supercells and tornadoes.

A vertical shear profile even closer in time and space than that derived from the Norman sounding at 0000 UTC was obtained just north of El Reno by RaXPol during the first deployment (Table 1, Fig. 7) while the storms were organized in a broken line segment along the front (Fig. 6c). This hodograph was obtained as an extended velocity azimuth display (EVAD; Matejka and Srivastava 1991) east of the convective storms (Figs. 3e–f) averaged from five volume scans at 2220, 2222, 2224, 2226, and 2228 UTC. Most of the EVAD was taken in clear air at low levels or stratiform precipitation aloft. The vertical shear in the lowest 6 km was just over 20 m s^{-1}, which is supportive of supercells, but was not extreme. The EVAD from RaXPol is quite similar to the hodograph from the Norman sounding, with differences likely attributable to differences in time (e.g., the Norman sounding was launched later than the RaXPol data used to construct the EVAD, when the low-level jet likely had intensified more) and space.

The surface front visible in detail in Fig. 1 is also seen at 0000 UTC at 850 hPa, mainly as a wind-shift line extending from southeastern Kansas west-southwestward.
across northern and northwestern Oklahoma (Fig. 7a); the temperature gradient associated with the front is not well represented in southwestern Kansas or the Texas Panhandle, owing to the elevated terrain. The front is also seen at 700 hPa, where there is a strong north-to-south temperature gradient and zone of confluence (Fig. 7b).

At 0000 UTC 1 June 2013, a major synoptic-scale cyclone was located near the northeastern corner of South Dakota at all levels. A front also is apparent in the midtroposphere at 500 hPa in southern Kansas, as the temperature varies from approximately −7°C south of the front in northwestern Oklahoma to −18°C in central Colorado (Fig. 7c). Such a midtropospheric front is characteristic of the cyclonic-shear side of a jet in the baroclinic westerlies (Bluestein 1993); at 500 hPa the airflow was from north-central Utah, southeastward into southwestern Kansas and the northern Texas Panhandle, and then into a jet streak with wind speeds of 45 m s⁻¹ over eastern Kansas and southern Iowa. A consequence of a midtropospheric front on the environment is that it was relatively warm on the anticyclonic-shear side of the jet, which lay over Oklahoma. Often subsidence warming on the anticyclonic-shear side of the jet reduces CAPE and results in an environment less conducive for supercell development. In this case, however, the surface temperature was so warm and the surface dewpoint so high that the relatively warm temperature at 500 hPa of −7°C did not prevent high CAPE from being realized.

Along with the midtropospheric jet, an upper-level jet (>37.5 m s⁻¹) was evident at 250 hPa (Fig. 7d) flowing around the base of the cyclone over northeastern South Dakota. Oklahoma was on the anticyclonic-shear side of the jet, so that anvil material from convective storms was not being evaporated from its updraft as vigorously as it would have been farther to the north in Kansas, so that the intensity of the storms in Oklahoma was probably limited.

While CAPE was relatively high and the vertical shear in the lowest 6 km was supportive of supercells, the shear was not as high as observed during some tornado outbreaks (e.g., Grams et al. 2012), and the storms formed on the relatively warm, anticyclonic-shear side of the mid- and upper-tropospheric jet. The storms in central Oklahoma formed near the right-entrance region (right-rear quadrant) of a jet streak at 250 hPa, which was located in northeastern Kansas and northern Missouri (Fig. 7d), and near the right-front exit-region (right front) quadrant of a jet streak at 500 hPa, which was located in the northern Texas Panhandle, as evidenced by the relatively strong wind at Amarillo (Fig. 7c). The location of the storms with respect to the 250-hPa jet streak is consistent with the findings of Rose et al. (2004) and Clark et al. (2009). The location of warmer air at 500 hPa in central Oklahoma is consistent with the finding of weak sinking motion in the midtroposphere in the composite right-front quadrant of a jet streak (Clark et al. 2009); in this case, the jet streak was located at 500 hPa in the northern Texas Panhandle.

In summary, storms formed along a weak stationary front, where lift likely initiated storms, and southwest of the frontal zone, along a dryline, which was marked by a zone of strong horizontal dewpoint gradient, but not a sharp, convergent wind-shift line. Since the first storms did not form along the front, but along a zone of high equivalent potential temperature near the dryline where CIN was close to zero, it is concluded that these storms were probably triggered mainly as a result of air reaching convective temperatures; we cannot quantify from the observations, however, to what extent storm formation may have been caused by a vertical circulation along the dryline.

4. Storm organization and tornado behavior

Upon first deploying RaXPol north of El Reno (Figs. 8 and 9a) in the direct path of a severe thunderstorm warning, we decided to relocate to the south in anticipation of an east-northeast, east, or east-southeast movement of the southern end of the broken convective line of storms. Consideration was given to a more southern target area (Fig. 2f, the storms evolving to the south in southwestern Oklahoma), but we were concerned that, with much hotter air to the south, the CIN might be higher and CAPE lower to the south than it was in central Oklahoma. Furthermore, based on may previous storm intercepts, we have found that the intersection of a front with a dryline is often the most favorable location for tornadic supercells (e.g., Bluestein and Parks 1983; Rose et al. 2004), especially when the linear forcing along a front favors cell mergers and convective-line structures, while forcing along the dryline often favors more isolated cells. RaXPol’s second deployment location, just southwest of downtown El Reno, was directly to the east-southeast of a mesocyclone near the southern end of the broken line of convection and just to the right of the center of the projected path of the mesocyclone and directly in the path of a tornado warning (Fig. 9b).

a. Tornadogenesis and evolution of the parent supercell and other supercells

Although there was a brief tornado reported at 2255–2256 UTC (Fig. 8), the most intense and longest lived of all the tornadoes on 31 May 2013 produced by the El Reno supercell (marked with an A in Fig. 12) began at 2304 UTC to the west-southwest of our deployment spot; the tornado moved initially to the southeast. The entire genesis phase of this tornado was documented by volume scans (Table 1; Fig. 10). By 2305 UTC, the tornado was
FIG. 8. Locations of the deployments of RaXPoI for (a) broad and (b) zoomed-in views, with respect to tornadoes \(T_1-\ldots-T_9\) and other geographic features. (a) All deployment locations \(D_1-\ldots-D_7\) and tornado tracks (color coded, along with NWS enhanced Fujita ratings in the insert) are shown along with the track of RaXPoI (blue). (b) Only deployment spots \(D_1\) (2210–2230 UTC), \(D_2\) (2247–2315 UTC), \(D_3\) (2324–2326 UTC), mobile (2326–2329 UTC), \(D_4\) (2332–2339 UTC), \(D_5\) (2349–2352 UTC), \(D_6\) (0006–0013 UTC), and \(D_7\) (0032–0040 UTC). The red pushpin marked with an M is the location of the El Reno Mesonet site. The large tornado damage path is outlined in orange, with the path taken by the center of the Doppler radar vortex signature given in orange; the anticyclonic tornado path is given in yellow. Damage tracks and enhanced Fujita ratings are courtesy of the NWS.
FIG. 9. Some key features depicted as images captured on a mobile phone, via the RadarScope commercial application for the iPhone. (a) Radar reflectivity factor from KTLX associated with the initial line segment of deep convection at 2235 UTC 31 May 2013. Yellow borders denote severe thunderstorm warning areas and the blue circle denotes location of RaXPol. (b) Doppler velocity at 2249 UTC 31 May 2013, just prior to the first tornado. Red borders show the tornado warning area, the deep purple circle marks the cyclonic vortex signature, and the light purple line marks I-40. (c) As in (b), but for 2318 UTC (when the large tornado was in progress). (d) As in (b), but for 2330 UTC. (e) As in (d), but for radar reflectivity factor. (f) As in (b), but for 2339 UTC (when cyclonic and anticyclonic tornadoes were in progress). The smaller circle marks the anticyclonic vortex signature. The scale in (a) is relatively zoomed out to see all of the storms. The scales in (b)–(e) are zoomed in more than in (a) to see the location of RaXPol with respect to the tornadic vortex signature and hook echo. The scale in (f) is zoomed in the most, to see the location of RaXPol with respect to both the cyclonic and anticyclonic tornadoes.
FIG. 10. Radar reflectivity images (dBZ) from RaXPol at a 4°-elevation angle approximately every 5 min during tornadogenesis of the large, cyclonic tornado at (a) 2256, (b) 2301, (c) 2306, (d) 2311, (e) 2315, and (f) 2324 UTC 31 May 2013. Range rings are shown every 5 km.
marked by a weak-echo hole (WEH)\(^1\) (e.g., Wakimoto and Martner 1992; Bluestein et al. 2004, 2007b; Wakimoto et al. 2011; Tanamachi et al. 2012, 2013), thought to be caused by centrifuging of the most massive radar scatterers (Dowell et al. 2005), at the tip of a hook echo. The storm as it approached at 2258 UTC had a laminar and striated base, suggestive of forced, stable lift (Fig. 11a) (Bluestein 2013). The beginning of the main tornado at 2305 UTC appeared as a large, cone-shaped condensation funnel (Fig. 11b). (Other tornado tracks are indicated in Fig. 8.)

The parent tornadic supercell was semidiscrete when the main tornado formed (Fig. 12a), but separate cells developed upstream of the existing supercell by 2316 UTC (Fig. 12b). By 2329 UTC (Fig. 12c), when the tornado was still large and violent, even newer echoes appeared to the west and southwest. By 2346 UTC (Fig. 12d) a new supercell (marked with a B in Fig. 12) had evolved just 20–30 km to the west of the supercell that spawned the large tornado near El Reno; the tornado had dissipated several minutes earlier. The new supercell tracked eastward (Figs. 12e,f); by 0029 UTC 1 June 2013 a third supercell (marked with a C in Fig. 12) formed

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\(^1\) The WEH is sometimes described as an “eye” or “ring” or “low-reflectivity eye” (LRE) of high reflectivity (e.g., Wurman and Gill 2000; Bluestein and Pazmany 2000; Bluestein et al. 2007a; Tanamachi et al. 2007; Wurman et al. 2010; Wurman and Kosiba 2013).
FIG. 12. Radar reflectivity images (dBZ) from KTLX at a low elevation angle depicting the evolution of the El Reno tornadic supercell (marked with an A) and neighboring storms (marked with a B and C), every ~15 min, at (a) 2302, (b) 2316, (c) 2329, (d) 2346 UTC 31 May 2013, (e) 0001, (f) 0015, (g) 0029, and (h) 0042 UTC 1 Jun 2013. Range rings are shown every 10 km. White (black) dot denotes location of the El Reno Mesonet site (RaXPol deployment site). Arrows point to hook echoes. Elevation angles are (a) 0.9°, (b) 1.3°, (c) 0.5°, (d) 1.3°, (e) 0.5°, (f) 0.9°, (g) 0.9°, and (h) 0.5°.
Each of the three supercells evolved from cells that developed at the same location, ~80 km to the west-northwest of the Twin Lakes radar (KTLX), to the southwest of El Reno, near the cold front, which was reinforced by an outflow boundary from the convective storms. This behavior—the formation of successive supercells at a fixed location (colloquially referred to as an anchor point)—has been noted by storm chasers in other cases of tornadic-supercell development near the dryline and its intersection with a front or outflow boundary, but to the best of the authors’ knowledge has not been well documented in the refereed literature. Bluestein and Parker (1993) found that backbuilding, discrete propagation in the direction opposite to that of storm motion (e.g., Newton and Fankhauser 1964), often occurs near the dryline, and has been associated with successive development of convective storms in developing squall lines (e.g., Bluestein and Jain 1985). Between 2300 and 0100 UTC, the supercells became increasingly separated from the line of convection to the north (Figs. 3g–i).

Surges of outflow from the first two supercells were recorded at the El Reno Mesonet site several minutes after 2300 UTC and also just after 2345 UTC (Fig. 13). Each surge was associated with a maximum wind speed of ~20 m s$^{-1}$ from the north-northeast. The first surge occurred just after the pressure had begun to rise rapidly (~3 hPa in ~20 min) and both temperature and dewpoint had begun to fall as the air became nearly saturated. From Fig. 12a it is seen that the first surge occurred within the forward-flank core of the first supercell, while the tornado was ~10 km to the southwest. It is therefore likely that the first surge marked the passage of a gust front originating in the forward-flank downdraft region of the supercell. Visual evidence of a gust front/relatively cool outflow to the north in the forward-flank downdraft region is found in Fig. 14a: looking to the west and northwest of our deployment spot toward a region just to the south of the El Reno Mesonet site, ~5 min or so after the north-northeasterly surge had passed by the site, a tail cloud is seen...
extending to the north of the supercell’s wall cloud (which had rapid southward-moving cloud tags) and a greenish-tinged precipitation area is seen to the north. Tail clouds are thought to be evidence of cooler air behind the forward-flank precipitation region of a supercell rising up into the storm’s main updraft (Rotunno and Klemp 1985; Atkins et al. 2014). There is numerical evidence that developing supercells that encounter a cool boundary (e.g., from outflow from earlier convective storms or from a preexisting cold front) produce low-level mesocyclones more quickly and these low-level mesocyclones are more persistent, so that the potential for tornadogenesis is greater (Atkins et al. 1999). There is also numerical evidence that when the parent supercell tracks along the boundary or more toward the warm side, low-level mesocyclones are more intense (Atkins et al. 1999).

Fig. 14. As in Fig. 11, but for (a) 2309 (a wide-angle view to the west of the tail cloud at the right and a tornado indicated with an arrow) and (b) 2314 UTC. Vortices are indicated by arrows within or near the large tornado, to the southwest. One of these may be the satellite vortex noted in Fig. 15. [Photographs courtesy of H. Bluestein.]
In the case of the El Reno supercell of 2013, it did interact with outflow from the north and it did track along or to the right of the cooler air, but the outflow may have been only on the storm scale, since there was no direct evidence of any meso- or larger-scale boundary.

The second surge occurred when the pressure was rising, but not rapidly; the air was nearly saturated and the temperature and dewpoint rose slightly rather than fell rapidly. From Fig. 12d, it is seen that the second surge occurred within the forward-flank core of the second supercell. So, in both instances the surge was associated with outflow under the forward flank of the core of a supercell, but the thermodynamic behavior was different. In this case, unlike the first surge, the second surge occurred to the rear of a preexisting cold pool. The second surge is reminiscent of secondary, internal downdraft momentum surges in supercells, but in this case the second surge came from the forward flank of a different supercell rather than from the rear flank of the same supercell [e.g., see the summary in Skinner et al. (2014)]. It may be that this surge was dynamically driven, but we find evidence only that it was probably not thermodynamically driven.

It is also noted from Fig. 13 that while convective storms were evolving ~20–30 km to the northwest, west, and southwest of the El Reno Mesonet site (Fig. 3), a relatively slow drop in pressure (~3 hPa between 1800 and 2130 UTC) was followed by a rapid drop in pressure of ~2.5 hPa in just 30 min prior to 2200 UTC. No significant changes in temperature, dewpoint, or wind direction were noted, but there was a brief maximum in wind speed of ~13 m s\(^{-1}\) while the pressure was falling. It is hypothesized that the rapid pressure fall might have been due to the buoyancy associated with latent heat release in the growing updrafts in the developing convective storms downstream with respect to the surface wind.

b. Interaction between the main tornado and a satellite tornado

The short-term evolution, beginning ~10 min after the first surge at the El Reno Mesonet site, of the hook-echo region of the first supercell, is seen in RaXPol data depicted in Fig. 15. A WEH is seen, surrounded by a ring of enhanced reflectivity thought to be associated with the tornado debris cloud (e.g., Wurman and Gill 2000; Bluestein et al. 2007a). A primary band of reflectivity, some of which is broken up into smaller segments, is wrapped around an outer ring of enhanced reflectivity. Preceding the northeastern portion of this band is a narrow zone of weak reflectivity or a notch (e.g., Bluestein et al. 2007a,b). Near the region of the hook echo, where the notch becomes so narrow that it is difficult to resolve anymore, a WEH appeared at 2313:54 UTC (Fig. 15a), coincident with a tornado vortex signature in Doppler velocity (not shown), ~3–4 km from the center of the main tornado. This WEH therefore seemed to be associated with a second, satellite tornado (marked by an arrow in Fig. 15), which rotated cyclonically around the main tornado and dissipated as it appeared to be absorbed into it. The second tornado therefore appeared to form separately from the main tornado (Edwards 2014) and moved southward and southeastward around the main vortex.

RaXPol at this time was northeast of the tornado (Figs. 8a and 9c), and it was assumed that the tornado would continue moving to the southeast; we realized that we were at risk from strong outflow from the north and large hail, and we would lose sight of the tornado inside a core of precipitation if we remained where we were. We chose not to relocate RaXPol to the south, because we were concerned about the possibility that we would cross directly into the path of the tornado. We also recognized that a continuing intercept strategy to the south would require traveling on minor, rural roads and would have limited options if the tornado eventually crossed the Canadian River to our southeast (see the southern edge of Fig. 8a). Instead, we chose to drive east and north to Interstate-40 (I-40) and move quickly to the east for safety and to reposition to the east or northeast of the tornado, which would not be visible during our repositioning owing to intervening precipitation. After the third deployment (Fig. 8), we were surprised to see that the tornado vortex signature was due west of us on I-40 (Figs. 8 and 9d,e), not to our southwest as anticipated. The first author mistakenly thought that the original tornado may have dissipated and a new one had formed to the east or northeast, something that happens during some types of cyclical tornadogenesis (Adlerman and Droegemeier 2005). In fact, the main tornado, which had been moving to the southeast, turned to the left and moved eastward and then northeastward while we were in transit between the second and third deployment locations.

c. Multivortex behavior

As the isolated satellite vortex presented in Fig. 15 had revolved partway around the main tornado, several subtornado-scale vortices (subvortices) appeared (Fig. 14b) near the time the crew moved to the third deployment spot (Fig. 8). [From chaser videos (not shown) it is seen that there were multiple vortices during our second deployment, though RaXPol was apparently not close enough to resolve any of the subvortices then.] The satellite vortex and a second, weaker vortex were resolved by RaXPol while the radar truck was in transit (Fig. 16). There are two extrema in inbound Doppler velocity to the
FIG. 15. Revolution of a satellite tornado around the main tornado as depicted by radar reflectivity imagery (dBZ) from RaXPol at a 4°-elevation angle approximately every 15 s, at (a) 2313:54, (b) 2314:09, (c) 2314:25, (d) 2314:41, (e) 2314:57, and (f) 2315:12 UTC 31 May 2013. The WEH associated with the satellite tornado is indicated with an arrow. Range rings are shown every 1 km.
east of the center of the tornado (Fig. 16c). The maximum inbound Doppler velocity farther from the center of the tornado is of tornadic strength (∼55–60 m s⁻¹) and covers a relatively large area outside of the smaller-scale vortex, whose asymmetric Doppler velocity couplet is only ∼250–300 m in diameter; this outer wind maximum may be associated with the rear-flank downdraft (Fig. 16c, the blue region at the right). The asymmetric flow at the greater ranges from the center of the tornado seems to indicate that the rear-flank downdraft was extremely intense, although from ρhv data (Fig. 16b), in which the debris signature is centered on the two vortices, it appears that the outer inbound velocity maximum may have been associated with a larger-scale vortex as winds associated with the rear-flank outflow were superimposed on a 2.5–3-km-diameter, larger-scale vortex of tornadic intensity.

A subvortex was seen at 2325 UTC on the northern fringe of the main tornado (Fig. 17). The condensation funnel associated with this subvortex appeared to lean radially outward with height, as is seen in some multivortex tornadoes and laboratory models of multivortex tornadoes [e.g., Bluestein and Pazmany (2000), their Fig. 2; Church et al. (1979)], probably as a result of frictional inflow near the ground. Rotunno (1984) showed that subvortices should lean outward slightly and in the direction of the azimuthal component of the wind; it may be inferred then that projected onto a vertical plane they

FIG. 16. Radar imagery from RaXPol at a 3°-elevation angle, at 2315:10 UTC 31 May 2013, depicting two subvortices in the El Reno tornado at a low elevation angle. Range rings are shown every 1 km. (a) Radar reflectivity factor (dBZ); (b) copolar cross-correlation coefficient; (c) dealiased Doppler velocity (m s⁻¹), for NCP > 0.2, with the motion of the radar truck removed; and (d) spectrum width (m s⁻¹). The circles surround the subvortices in the multivortex tornado based on weak-echo holes/curls in reflectivity [in (a)], lobes of reduced ρhv [in (b)], cyclonic vortex signatures [in (c)], and lobes of enhanced spectrum width [in (d)].
can appear to lean outward with height. The broad extent of the main tornado condensation funnel is striking and will be discussed in more detail in another paper.

At the start of our third deployment, the RaXPol Doppler velocity field exhibited characteristics mainly of a single-vortex tornado. Following the brief third deployment, data were collected as the radar was in motion to the east. Near 2328 UTC, while the radar truck was moving, the tornado contained multiple vortices (Fig. 18). High-resolution RaXPol data show two WEHs and a third enhanced area of reflectivity (Fig. 18a), each of which is associated with a separate cyclonic vortex signature (Fig. 18b). Each subvortex was characterized by relatively high spectrum width (Fig. 18c); the maxima in spectrum width together formed a ring of relatively high spectrum width (Fig. 18c). High spectrum width can indicate turbulence and high shear within each radar volume. A tornado debris signature defined by low $r_{hv}$ (Fig. 18d) was coincident with the ring of enhanced radar reflectivity associated with the WEHs, etc., with the cyclonic vortex signatures, and with the ring of enhanced spectrum width; it seems likely that the presence of debris of varying sizes, shapes, and movement speeds resulted in a very complex Doppler spectrum. Convex lobes of low $r_{hv}$ (Fig. 18d) were coincident with the cyclonic vortex signatures, including a fourth, much weaker one in the southeastern quadrant of the tornado. There is therefore evidence of four subvortices within the eastern half of the parent tornado vortex.

The detailed behavior of these multiple vortices is currently being analyzed and will be reported upon in future publications. Wurman et al. (2014) attribute the deaths of storm chasers to a subvortex (or subvortices) similar to those shown in Fig. 18. The subvortices were embedded within a larger-scale cyclonic vortex, whose Doppler velocity couplet was $\sim 1.5$ km wide (Fig. 18b).

d. Formation of an anticyclonic tornado

The last feature to be noted in this paper is a rare anticyclonic tornado (labeled $T_3$ in Fig. 8). This anticyclonic tornado began near 2329 UTC while RaXPol was moving eastward on I-40. It was first spotted to our southwest as a funnel cloud, underneath which there was a rotating debris cloud (not shown). Data collection at the fourth deployment location began $\sim 3$ min after the anticyclonic tornado had formed, though data were collected during tornadogenesis while the radar truck was in transit (not shown). At the fourth deployment location, the weakening, yet extremely large, main, cyclonic tornado and the anticyclonic tornado were both probed by RaXPol (Figs. 9f and 19).

The cyclonic tornado was very broad, marked by a $\sim 5$-km-wide ring of enhanced reflectivity (Fig. 19a) and a polarimetric debris signature of low $Z_{DR}$ (Fig. 19b) and $r_{hv}$ (Fig. 19c). The cyclonic vortex signature associated with the cyclonic tornado was very broad: the distance between in- and outbound maxima was $\sim 2$ km. Based on Doppler velocity (the width of the approximately
±30 m s\(^{-1}\) velocity couplet) and the width of the polarimetric debris signature (an upper bound), the tornado was at least 4 km wide, or at least 2.5 mi wide, making it one of the widest, if not the widest, ever documented. Centrifuging radially outward of debris (Dowell et al. 2005) could have caused the unusually wide debris signature.

There is some hint in Fig. 19c of the vestige of a smaller-scale, but weak Doppler velocity couplet (brown–green couplet near the center), which may be the remains of the main tornado seen in Fig. 16. It appears as if a mesocyclone-scale tornado persisted, while a small-scale tornado apparent at the center, seen in Fig. 16, had weakened considerably or dissipated by the time of the Fig. 19 scan.

The anticyclonic tornado formed along the rear-flank gust front, which was marked by swirls in reflectivity (Fig. 19a) and by vortex signatures (Fig. 19d). The southernmost vortex had a strong anticyclonic vortex signature along with associated polarimetric debris signatures in both \(Z_{\text{DR}}\) and \(\rho_{hv}\). Like other observed cyclonic–anticyclonic tornado/vortex pairs in supercells (e.g., Wurman and Kosiba 2013), these counterrotating tornadoes moved along different paths (e.g., Snyder et al. 2007). The anticyclonic tornado turned to the right and curved to the southeast and south. From the standpoint of tornado warnings to the public, it can be confusing to issue warnings for both tornadoes. Figure 9f shows that the area warned for tornadoes included both the

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**Fig. 18.** Radar imagery from RaXPol at a 2°-elevation angle, at 2327:54 UTC 31 May 2013, depicting the multivortex stage of the El Reno tornado at a low elevation angle. Range rings are shown every 1 km. (a) Radar reflectivity factor (dBZ); (b) dealiased Doppler velocity (m s\(^{-1}\)), for NCP > 0.2; (c) spectrum width (m s\(^{-1}\)); and (d) copolar cross-correlation coefficient. The arrows point to the subvortices in the multivortex tornado based on WEHs/enhanced reflectivity [in (a)], cyclonic vortex signatures, lobes of enhanced spectrum width [in (c)], and lobes of reduced \(\rho_{hv}\) [in (d)].
cyclonic and anticyclonic tornadoes, but there was no indication that the motion of each tornado differed from that of the other.

A view of RaXPol data zoomed in on the anticyclonic tornado is shown in Fig. 20. The reflectivity pattern exhibited an anticyclonic hook with an inner debris ring (Fig. 20a).

Tornadic debris signatures of low $Z_{DR}$ (Fig. 20b) and low $\rho_{HV}$ (Fig. 20c) were coincident with the tornado, which was marked by Doppler velocity shear of $\sim 75$ m s$^{-1}$ over $\sim 500$ m (i.e., $\sim 0.3$ s$^{-1}$ vertical vorticity). Based on the width of the debris signature in $\rho_{HV}$, and the width of the Doppler wind speeds in excess of $\sim 35$ m s$^{-1}$ (Figs. 20c,d), the anticyclonic tornado at this time was $\sim 750$ m wide.

5. Summary and conclusions

The supercell on 31 May 2013 that spawned a large and violent tornado near El Reno, Oklahoma, began along the southern end of a line segment of convective storms in an environment conducive for the formation of supercells. The 0–6-km vertical shear was not extremely large. In cases such as these, the hypothesis that the intensity of low-level mesocyclones (and by inference, tornadoes) is maximized for unique combinations of low-level vertical shear and the intensity of the cold pool produced by the convection (Markowski and Richardson 2014) might help explain why the deep vertical shear by itself may not have been a good predictor of a violent tornado.
The northern portion of the line segment that developed west and northwest of the Oklahoma City area formed along a front, where it was possible that forced lift may have played a role in storm initiation, as evidenced by surface temperatures and dewpoints near (based on RAP analyses) or below the convective temperature (based on Mesonet data and soundings). Along the southern end of the line, however, the storm was not along a boundary and it was found that insolation in a very moist air mass ahead of a dryline probably played a dominant role in storm initiation. Convective storms that formed along the dryline to the south across southwestern Oklahoma were probably suppressed by a strong cap and drier air aloft, as evidenced by the much warmer and drier surface temperatures in the well-mixed air to the west of the dryline.

The mean vertical shear vector below 6 km AGL was oriented in the east–west or west-northwest to east-southeast direction (Fig. 6), while the front was oriented in the northeast–southwest direction initially (Figs. 1a–e); however, with the appearance of cool outflow from the convective-line segment to the north, the front had become oriented more in the north–south direction near the storm (Fig. 1f). The southernmost, cyclonically rotating, right-moving storm presumably had the best chance for remaining at least semi-isolated and therefore being more long lived, since the mean vertical shear vector was aligned approximately normal to the boundary along which storms were forced (Bluestein and Weisman 2000). Any neighboring storms to the north, along the line segment near the front, would probably have not developed into discrete supercells owing to storm collisions and outflow mergers between left- and right-moving cells. So, the localized nature of the tornadoes on 31 May 2013 may have been a result of supercell formation having been suppressed to the north and south of the intersection of the dryline and front.

Fig. 20. As in Fig. 19, but for the (zoomed in) anticyclonic tornado.
Two semidiscrete supercells subsequently developed to the west of the El Reno tornadic supercell, one after the other, at approximately the same location, along the intersection of the dryline and front. This behavior might be explained by the movement of the first storm away from the warmest and moistest air, leaving behind a region warm enough and moist enough for storm initiation, once the cold pool from the preceding storm had moved eastward. Convergence along the retreating cold pool boundary may have made this location a preferred one for convective initiation.

Another intriguing behavior of the primary tornadic supercell was the formation of a strong anticyclonic tornado along the tail end of the rear-flank gust front, as has been documented in a number of other rare cases (e.g., Brown and Knupp 1980; Fujita 1981; Bluestein et al. 2007a; Snyder et al. 2007; Wurman and Kosiba 2013). In addition to those cases already noted in the literature and at conferences, the authors have recently compiled other mobile Doppler radar datasets for several other similar cases. Straka et al. (2007) and Markowski et al. (2008) have suggested that the source of vorticity of the anticyclonic member of a cyclonic–anticyclonic vorticity pair along the right flank of a supercell might be baroclinically generated vorticity along the rear-flank gust front that is tilted upward at its southern end (in the Northern Hemisphere).

Detailed radar analyses of the formation of the El Reno tornado of 2013, the characteristics of both the cyclonic and anticyclonic tornadoes, and comparisons of low-level RaXPol wind data with damage surveys are in progress and will be discussed elsewhere. Both idealized numerical simulations and real-data assimilation experiments should be carried out to undertake more thorough investigations of the physical mechanisms responsible for the behavior of the parent storm.

Acknowledgments. This study was supported by NSF Grants AGS-0934307 and AGS-1262048. John Meier at the Advanced Radar Research Center at the University of Oklahoma (OU) maintained RaXPol for use during our field experiment and OU provided matching funds for the MRI grant that funded the construction of the radar. This research was performed while the second author held a National Research Council Research Associateship Award at the National Severe Storms Laboratory. Gabe Garfield and Doug Speheger from the Norman NWSFO provided some of the tornado track information from the El Reno storm. Mike French (NSSL) provided a useful informal review. Corey Potvin (NSSL) and two anonymous reviewers provided very helpful comments.

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