The Formation and Early Evolution of the Greensburg, Kansas, Tornadic Supercell on 4 May 2007

HOWARD B. BLUESTEIN

School of Meteorology, University of Oklahoma, Norman, Oklahoma

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ABSTRACT

During the evening of 4 May 2007, a large, powerful tornado devastated Greensburg, Kansas. The synoptic and mesoscale environments of the parent supercell that spawned this and other tornadoes are described from operational data. The formation and early evolution of this long-track supercell, within the context of its larger-scale environment, are documented on the basis of Weather Surveillance Radar-1988 Doppler (WSR-88D) data and mobile Doppler radar data. The storm produced tornadoes cyclically for about 30 min before producing a large, long-lived tornado. It is shown that in order to have forecasted the severe weather locations and times accurately, it would have been necessary to have predicted 1) the localized formation of an isolated convective storm near/east of a dryline, 2) the subsequent splitting and resplitting of the storm several times, 3) the growth of a new storm along the right-rear flank of an existing storm, and 4) the transition from the cyclic production of small tornadoes to the production of one, large, long-track tornado. It is therefore suggested that both extreme sensitivity to initial conditions associated with storm formation and the uncertainty of storm behavior made it unusually difficult to forecast this event accurately.

1. Introduction

During the evening of 4 May 2007 a tornado rated as a category 5 storm on the enhanced Fujita scale (EF-5; LaDue and Mahoney 2006) devastated much of Greensburg, Kansas (GRE); at least 10 people died and 95% of the town was damaged or destroyed (National Climatic Data Center 2007). This tornado, the first to be rated an EF-5 and likely the most intense since the tornado that struck Oklahoma City and Moore, Oklahoma, on 3 May 1999, was at some times as wide as 2.7 km and had a pathlength of 35.4 km. Other tornadoes, some also significant, tracked across portions of southern Kansas, but fortunately did not strike any heavily populated areas and did not inflict damage as extensive as that inflicted at Greensburg. Although severe weather had been anticipated in southwest Kansas on 4 May 2007 [a convective outlook with a moderate risk of severe weather was issued by the Storm Prediction Center at 0600 UTC, over 24 h before the event (information online at http://www.spc.noaa.gov/products/outlook/archive/2007/day1otlk_20070504_1200.html) and a tornado watch was issued almost 3 h before the event (information online at http://www.spc.noaa.gov/products/watch/2007/ww0227.html)], the details of the storm’s evolution were not easily forecasted. Since the Greensburg event was associated with prolific, tornado-producing, parent supercells, it is important to analyze the details of this event to learn how we can better forecast future similar events and to identify specific gaps in our fundamental knowledge of the behavior and dynamics of tornadic supercells.

The purpose of this paper is to document the evolution of the Greensburg storm beginning when its parent storm formed and continuing until the Greensburg storm first approached maturity. The complete details of the subsequent evolution of the Greensburg storm and its tornado, and subsequent storms, however, will not be described here. The data used to analyze the formation and early evolution of the Greensburg storm include Weather Surveillance Radar-1988 Doppler (WSR-88D) data, mobile Doppler radar data, visual observations, and satellite imagery. In addition, wind and thermodynamic data from the standard observing network and the Oklahoma and Texas mesonets, and

Corresponding author address: Dr. Howard B. Bluestein, School of Meteorology, University of Oklahoma, Suite 5900, 120 David L. Boren Blvd., Norman, OK 73072.
E-mail: hblue@ou.edu

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wind data from the National Oceanic and Atmospheric Administration (NOAA) Profiler Network, are also used. Finally, forecasts from experimental numerical prediction models that explicitly simulate convection are considered. The latter is especially relevant as the NOAA/National Weather Service (NWS) moves from a policy of “warn on detection” to one of “warn on forecast” (information online at http://ewp.nss.noaa.gov/wasis2008/AdvancedWASIS-PHI.ppt).

While the synoptic conditions for strong supercells were evident over a relatively broad area, the location(s) of the storm formation and the nature of the subsequent storm evolution were not well anticipated. The initial supercell formed within a region of cirrus overcast, which had some breaks. The presence of cirrus overcast tends to diminish the likelihood of storm formation as surface temperatures are reduced and convective temperatures are not attained (e.g., Bluestein et al. 1987; Thompson and Edwards 2000). In addition, it will be demonstrated that the Greensburg storm evolved from a complex series of splitting cells, whose original parent storm began as an isolated cell near/east of a dryline. In other words, to have forecasted the Greensburg storm precisely, one would have had to have anticipated isolated storm formation ahead of the dryline in a particular location and then realized that storm formation at other locations would be inhibited so that neighboring-storm interactions (Bluestein and Weisman 2000) would be minimal or nonexistent. In addition, one would have had to have anticipated a series of cell splits and correctly ascertained which cell split would have yielded a tornadic supercell. Doing the latter is quite a challenge.

The synoptic-scale and mesoscale aspects of storm initiation and the storm environment are described in section 2. A detailed description of the storm formation and evolution from a radar perspective is given in section 3. In section 4, guidance from available numerical forecast models will be evaluated in light of what actually happened. Finally, section 5 contains a summary of the most important details of the event and challenges that face forecasters, theorists, and modelers in correctly predicting, understanding, and simulating the Greensburg event and possibly other events.

2. Synoptic-scale and mesoscale environment

a. Upper-air and surface features

An upper-level trough, whose axis was located near the Nevada–Utah border, was positioned far upstream (Fig. 1) both prior to and just after storm initiation; consequently, any strong synoptic-scale ascent associated with the trough was far west of Kansas and the Texas and Oklahoma Panhandles.

There were three distinct surface boundaries in the plains (Fig. 2). (a) A stationary front extended from central Nebraska southwestward across far northwestern
Kansas into southeastern Colorado. West and northwest of the front, winds were generally northerly or northeasterly, and east and southeast of the front winds were southeasterly. The temperature difference across the front was modest (\( \Delta T \approx 5^\circ-10^\circ \text{C} \)). (b) A dryline bulged northeastward from extreme southwestern Kansas, the Oklahoma Panhandle, and the Texas Panhandle. (c) There was a less-well-defined thermal boundary aligned approximately east-southeast to west-northwest across southern Kansas at 1800 UTC (Fig. 2a). This boundary, which was not associated with any earlier convective system, separated cloudy, cooler, moist air to the north from clear, warmer, moist air to the south; there was little if any change in wind direction across this boundary. With time, this boundary propagated northward as a diffuse warm front and became less well defined as sunset approached and solar insolation weakened (Figs. 2c and 2d). The baroclinicity of this boundary was therefore a consequence of differential diabatic heating.

Winds behind the dryline in the Texas and Oklahoma Panhandles backed between 1800 and 2100 UTC (central daylight time (CDT), the local time, is 5 h earlier), and after 0000 UTC most stations reported an easterly component (Fig. 2d) as the dryline retreated. The meteogram traces (not shown) at the Oklahoma mesonet station in Buffalo (in far northwestern Oklahoma, nearest to where storms later formed) indicated a steady pressure fall continuing until \( \approx 0100 \text{ UTC} \), at a rate that is faster \( [2 \text{ hPa} (3 \text{ h}^{-1})] \) than that associated with the diurnal/semidiurnal tidal pressure oscillations \( [1.5 \text{ hPa} (6 \text{ h}^{-1})] \) in this region (Crawford and Bluestein 1997). As a consequence of the backing of the winds behind the dryline in the afternoon and early evening (in response...
to the surface pressure falls that must have been forced by ascending air near the approaching upper-level trough and/or heating at higher terrain to the west), most of the dryline in the Texas and Oklahoma Panhandles was not marked by a band of well-defined convergence/confluence associated with a wind shift as there often is both with and in the absence of strong synoptic-scale forcing (e.g., Hane et al. 1997; Crawford and Bluestein 1997; Hane et al. 2001). The dewpoint increased with time at most mesonet sites in the region (not shown), especially after 2230 UTC, as moister air was advected westward.

There was, however, a localized area of enhanced apparent convergence just west of Pampa, Texas (PPA), at 2100 UTC (Fig. 2b) as a result of a backed wind at PPA; the surface dewpoint had increased over its value at 1800 UTC (Fig. 2a). By 0000 UTC (Fig. 2c), this region of enhanced convergence was no longer apparent. It will be seen subsequently that just east of this area the first storm formed within an hour.

b. Environmental vertical wind shear and thermodynamic stratification

The soundings nearest in space and time to the Greensburg storm and its predecessors were at Dodge City, Kansas (DDC; Fig. 3a) and Lamont, Oklahoma (LMN; Fig. 3b), at 0000 UTC. The DDC sounding was characteristic of the air mass behind the dryline; the boundary layer was dry adiabatic and deep, and extended up to about 625 hPa; the moisture was nearly well mixed in this layer. It is seen from the surface map near the time of the sounding (Fig. 2c) that the dryline was located somewhere east of DDC. Based on the author’s observations southeast of DDC, there was no visible line of convective clouds marking the dryline; the

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**Fig. 3. NWS rawinsonde observations at (a) DDC and (b) LMN at 0000 UTC 5 May 2007. Temperature scale (°C) is shown along the abscissa and pressure scale (hPa) along the ordinate. Thick solid lines indicated temperature and dewpoint. Wind barbs plotted to the right are as in Fig. 1. Various indices including CAPE (J kg⁻¹) are plotted at the right. Hodographs, based on rawinsonde observations, shown for (c) DDC and (d) LMN at 0000 UTC 5 May 2007. Abscissa and ordinate scales are in m s⁻¹. Wind levels are plotted in hPa, where 500 hPa is approximately 5.7 km MSL, or approximately 5 km AGL.**
Dryline therefore appeared to be diffuse. The LMN sounding (Fig. 3b) was located well east of the dryline; it was characterized by a well-mixed moist layer up to 800 hPa and a shallow, but significant, capping inversion just above. Its convective temperature was \(32^\circ-33^\circ\)C, approximately the same or just slightly warmer than that which was observed just east of the dryline in the eastern Texas Panhandle and in western Oklahoma late in the afternoon (Fig. 2b). The CAPE at LMN was \(\sim 3800 \text{ J kg}^{-1}\) and the virtual-temperature CAPE was nearly \(4000 \text{ J kg}^{-1}\), which are supportive of very strong, buoyant updrafts. Since the temperature gradient at 500 hPa was not strong east of the dryline (Fig. 1), it is reasonable to assume that the environment east of the dryline was characterized by high CAPE over a relatively broad area south of the thermal boundary in Kansas and east of the dryline.

The vertical shear over the lowest 6 km (Figs. 3c and 3d) was in excess of 20 m s\(^{-1}\), which is necessary for isolated supercell development in both an idealized

Fig. 4. (a) Time–height cross section of wind data (plotted as in Fig. 1) from the HVL NOAA wind profiler from 2100 UTC 4 May 2007 until 0200 UTC 5 May 2007, up to 7.5 km MSL (up to 6 km AGL). (b) Hodograph for HVL (see Fig. 6). In (b), winds are plotted in km AGL for 2100 UTC (just before storms formed) and 0100 UTC (while storms were on going and storm evolution was occurring, about 1 h before the birth of the Greensburg tornado) on 4/5 May 2007. Motion of storm N, the Greensburg storm (see Fig. 6), is indicated by “N.”
environment (Weisman and Klemp 1982) and is approximately what has been observed for tornadic supercells in nature (e.g., Bluestein and Parker 1993, Fig. 15; Rasmussen and Blanchard 1998, Fig. 3; Rasmussen 2003). Between 2100 and 0100 UTC, the winds aloft remained from the southwest at speeds increasing from \(\approx 18 \text{ m s}^{-1}\) at 3 km AGL to \(\approx 25–27.5 \text{ m s}^{-1}\) at 9 km AGL (Fig. 4b). However, as the low-level jet (LLJ) appeared just after 0000 UTC, the 500-m and 1-km winds increased from \(\approx 7.5–10 \text{ m s}^{-1}\) from the south or southwest to 15 m s\(^{-1}\) from the south-southeast (Fig. 4a). Thus, the appearance of the LLJ resulted in a marked increase in low-level shear (below 2–3 km AGL). The 0000 UTC DDC hodograph was similar to the 0100 UTC Haviland, Kansas (HVL), hodograph, the latter of which was representative of the environment nearer to and just upstream from the Greensburg storm. The overall hodograph was relatively straight from \(\approx 1\) to \(\approx 6 \text{ km AGL}\), and perpendicular to the hodograph in the boundary layer (Fig. 4b), as is the climatologically average hodograph for supercells (Bunkers et al. 2000, Fig. 5).

Fig. 5. Satellite imagery at about 30-min intervals, depicting the formation and evolution of convective storms prior to the Greensburg tornado from 2045 to 2325 UTC. Times (upper right in each panel) are in UTC on 4 May 2007.
3. Storm formation and evolution

The evolution of radar echoes during the late afternoon and evening of 4 May 2007 over the Texas and Oklahoma Panhandles and southern Kansas is examined mainly from archived level II data from the National Climatic Data Center (NCDC) and from mobile Doppler radar data. The goal is to determine the nature of the origin of the Greensburg storm and the mode(s) of its development and evolution. Satellite observations are used first to document storm formation.

a. Mesoscale organization

At 2045 UTC (Fig. 5), a line of cumulus formed in the southeastern Texas Panhandle near the dryline (Fig. 2b) as surface temperatures reached ~32°C. It is assumed that these cumulus must have had relatively high bases, since the 0000 UTC LMN sounding far to the east of the dryline indicated convective temperature of ~32°C, but for a surface dewpoint ~6°C higher than what was observed near the dryline. Storms did not form along this cloud line. A broken area of cirrus clouds had over spread the area by 2125 UTC (Fig. 5) so that most of the area west of the dryline was in the upper 20s °C, cooler than on the moist side (Fig. 2b).

The first sign of storm development was at 2155 UTC in the northeastern part of the Texas Panhandle (Fig. 5). The first precipitation echo (Figs. 6 and 7) appeared between 2151 and 2201 UTC, ~40-50 km to the east-northeast of a zone of enhanced clear-air echoes associated with a dryline bulge. The existence of a bulge is supported also by the relatively dry air that extended northward into southwestern Kansas (Fig. 2b). From Figs. 7, at 2201 and 2230 UTC, and from Fig. 2b, it is seen that the first storm formed just east of the dryline bulge, though shallow towering cumulus could have been hidden by the cirrus overcast earlier, closer to the dryline (Fig. 5, at 2125 and 2155 UTC). If an updraft ~10 m s⁻¹ had produced the precipitation associated with the first echo, and if the storm depth was ~10 km, then it would have taken air ~15 min to enter the cloud base and exit the top of the storm (producing precipitation along the way). Thus, the first cloud tower could not have been more than ~10 m s⁻¹ × 10³ s = 10 km west of the location of the first echo, which was east of the dryline bulge and therefore probably east of the dryline, not along it.

The first storm is named “A” and subsequent cells are named alphabetically according to the temporal order in which they first appeared. Although it would be best to identify cells by their updrafts (e.g., Bluestein and Weisman 2000), in the absence of better information about the location/existence of updrafts, cells are identified subjectively by the appearance of their precipitation echoes. It is assumed that each discrete echo of ~35 dBZ or stronger is in fact associated with an updraft, even though the updraft may be removed spatially from the location of the radar echo, and may lead the location of the radar echo in time. The evolution of the cells is discussed both from a plot of individual cell...
locations as a function of time (Fig. 6) and from series of WSR-88D plan position indicator (PPI) images of radar reflectivity from Amarillo, Texas (AMA), and DDC; Fig. 7 depicts the evolution on a 30-min time scale, while Fig. 8 depicts the evolution on a finer, 10-min time scale. (When a radar echo has relatively low reflectivity, but also exhibits temporal continuity with an earlier/later echo that has high reflectivity, the cell letter is shown between two parentheses.) While many of the details of the storm's evolution were hidden underneath an anvil canopy (Fig. 5, at 2255 and 2325 UTC), shadows from penetrating tops cast by several cells were evident near sunset (not shown). Since the nature of the identification and tracking of cells was subjective, some alternative interpretations are possible, especially with regard to determining when substructures of fleeting nature may be further identified as additional cells and in locating the centroid of radar echoes. However, it is
believed that the essence of the storms’ evolution has been properly identified in a qualitative sense, even though the precise number of discrete cells may be in small error and the location of each cell may have errors \( \sim 5-10 \) km. Each WSR-88D scan was considered, though not all scans are represented in Fig. 6, in order to render the figure less cluttered.

From Figs. 6 and 9 it is seen that there were three cells—A, C, and E—that began in isolation from other convective cells. Of these, E (also Fig. 5) tracked northeastward and dissipated before interacting with any other cell and without producing any reported severe weather. Cell C collided and merged with cell A early on, just after 2230. Cell A, the mother cell of all subsequent cells on this day in the region, produced two tornadoes near Arnett, Oklahoma, from \( \sim 2321 \) to 2345. By this time, cell A had produced three left-split cells—B, D, and G—the latter split occurring just as the first tornado was being reported. The splitting process on 4 May 2007 was characterized by pulses of new cells appearing adjacent to older cells, especially on the northwestern, northern, or northeastern flanks, but usually separated slightly from the predecessor cell. In this respect, the “split” cells were not classic splits in which a cell actually divides into two (e.g., Bluestein and Sohl 1979), but rather new cells that emanated from an existing cell. The type of splitting observed is probably more characteristic of supercells that split in an environment in which the hodograph has some clockwise curvature, so that splits are not symmetrical with respect to the vertical shear vector. Wilhelmson and Klemp (1981), while successfully modeling a storm in which a cell split more than once, noted that after the first split, subsequent splits occurred in a manner similar to that of the ones from the Greensburg case; that is, the radar echo did not elongate, but rather new, discrete cells formed adjacent to older ones, and the new cells continued to move in the same direction as the parent cells did before the splits. In the Wilhelmson and Klemp (1981) study, the hodograph had low-level curvature and was straight above the boundary layer, as is approximately the case in Figs. 3c, 3d, and 4b.

While cell B was relatively short lived, cell D produced another two generations of left-split cells (F and I), and cell G produced another four generations of left-split cells (Fig. 9). Some of these left-split cells were observed visually (but not photographed) by the author \( \sim 0045-0100 \) UTC from Protection, Kansas, just west of the subsequent location of the University of Massachusetts—Amherst X-band Dual Polarization Radar (XPol) “X-POL” in Fig. 6. These cells (probably J, K, and/or M) were relatively narrow cumulonimbus towers, whose updraft bases were located just to the north of their precipitation cores; they assumed the mirror-image structure of what cyclonically rotating, right-moving cells look like (e.g., Moller 1978, Fig. 2). In the absence of a photograph of any left-split cells on 4 May 2007, a similar image of a left-split cell on 7 April 2008 is shown in Fig. 10. New cloud material kept forming on the northern, forward flank, rather than on the right-rear flank as in right-split cells. Adequate storm-scale data are not available to assess the relative contributions of an upward-directed perturbation pressure gradient force and of density-current lift. Since the cloud base sloped upward with height to the south, however, it appears that density-current lift associated with storm outflow probably played some role, consistent with the Wilhelmson and Klemp (1981) idealized modeling study.

Cell A produced a fourth left-split cell (L) around 0030 UTC. Cell A continued on until \( \sim 0144 \) UTC, while cell L continued on only until \( \sim 0119 \) UTC. Left-split cells were the dominant mode of deep convection for the day until \( \sim 0030-0100 \) UTC, after which cell splitting was no longer observed.

Cell J formed just to the northwest of the second-generation left-split cell H at 2343–2344 UTC (Figs. 6 and 8). Most left-split cells formed just to the northwest of their predecessors; cell J was closest to cell H at 2343 UTC as viewed by the WSR-88D at AMA (not shown). Cell N began, atypically for this day, along the southwest flank of cell J between 0013 and 0038 UTC as series of small cells that congealed into one cell. By 0132 UTC (Fig. 7), cell N had collided with cell J and other radar echoes formed just to the north of N and west of J. Storm collisions, which have been documented (Bluestein and Parker 1993), can sometimes result in enhanced storm development via increased surface convergence (Wurman et al. 2007). The author was located in the precipitation associated with the western side of J and the eastern side of N. This information is provided for the reader in advance of the following discussion on the storm-scale organization. It was cell N that became isolated and eventually produced the tornado that struck Greensburg.

b. Storm-scale organization

At around 0125 UTC (Fig. 11a), the visibility at the author’s location just east of Protection (Fig. 12) increased as rain from cell N passed to the north and the edge of the forward-flank precipitation core became visible, as did a wall cloud, tail cloud, and striated base. About this time a funnel cloud was reported 3–4 mi (4.8–6.5 km) southwest of Sitka, Kansas, or about 23–24 km to the southwest of the author (Fig. 12), but not visible, owing to the long distance to it.
FIG. 7. Radar reflectivity at ~30 min intervals from the AMA (2201–0030 UTC) and DDC (0103–0302 UTC) WSR-88Ds on 4/5 May 2007, showing the formation (the radar is in clear-air mode at 2201 UTC) and early evolution of convective storms and the development of the Greensburg storm (N), which was over GRE at 0302 UTC. When a storm name is given in parentheses, it is just forming or is decaying (the reflectivity is ~ or less than 35 dBZ but shows time continuity with higher reflectivities at later or earlier times, respectively). XPol location is indicated when radar data were being collected or when the radar was being deployed. (Data courtesy of NCDC.)
Mobile-Doppler radar observations are used to document the storm-scale organization of cell N from the time it began producing tornadoes until the beginning of the Greensburg tornado. The Doppler radar used (Bluestein et al. 2007) operates at the X band, so that attenuation in the presence of heavy precipitation can distort the appearance of the radar echoes. The data from this radar are considered for a storm-scale analysis because the radar was much closer to the storm than the nearest WSR-88D, at DDC. Owing to intervening trees, the lowest-elevation-angle scans were contaminated to some extent by ground clutter. Therefore, scans at elevation angles of $\pm 5^\circ$–$8^\circ$ were used for illustrative purposes (representing heights 1–2 km AGL for ranges out to $\sim 15$ km and $\sim 2$–2.5 km AGL out to $\sim 20$ km, i.e., just above cloud base), even though the strongest Doppler velocities associated with tornadoes are expected at lower altitudes.

It took approximately 1 h ($\sim 0038$–$0138$ UTC) for cell N to develop into a tornadic supercell, which is not atypical, but faster than the mean time for tornadic supercells that begin near the dryline (Bluestein and Parker 1993, Table 2). At 0127 UTC, a weak cyclonic-shear signature [yellow (receding) to green (approaching) Doppler velocities] is evident at the location of a narrow hook echo on the rear flank of the storm (Fig. 13), which can be tracked ahead to the damage path of tornado 1 (Fig. 12), which began at 0132. A second hook echo and a weak cyclonic-shear signature (Fig. 13) located $\sim 6$ km to the east-southeast of tornado 1 (at 0145 UTC) were associated with the remains of a nearly temporally coincident tornado 2, which had ended at 0139 UTC. Tornado 3 then formed at 0148 UTC, $\sim 6$ km to the east-southeast of the remains of tornado 2 (Figs. 12 and 13). Tornado 4 formed at about the same time, $\sim 8$ km to the west-northwest of tornado 3. So, tornadoes 1 and 2 were paired, about 6–8 km apart, and tornadoes 3 and 4 were paired also $\sim 8$ km apart. The remains of tornado 4 are evident at 0200 UTC in Fig. 13, along with the appearance of a new tornado 5, to the east-southeast of tornado 4 by $\sim 10$ km. Thus, between $\sim 0130$ and 0200 UTC, four tornadoes were produced cyclically and in pairs. Cyclic mesocyclogenesis has been documented using mobile Doppler radars (e.g., Beck et al. 2006) that also showed as many as four simultaneous circulations (French et al. 2008).

Because there were multiple circulations, the motion of the radar echo mass is not accurately representative of the updrafts in the parent storm: While new circulations kept forming farther and farther to the east with respect to the parent storm, new updrafts must have also kept forming farther and farther to the east. As older circulations and updrafts migrated westward with respect to the reflectivity core as in Dowell and Bluestein
In addition to the radar-observed process of cyclic tornado production, the following intriguing concurrent satellite observation is noted. At the time the cyclic tornadogenesis began, a thin line of cloud at the cirrus level and apparent in the visible satellite imagery (2002a,b), Beck et al. (2006), and French et al. (2008), the overall motion of the core was probably to the left of the actual motion of the main updraft. In Fig. 12, it is seen that the damage tracks of tornadoes 1–4 were to the left of the core motion (Figs. 4b and 6).
(Fig. 14) was noted extending from storm N south-westward through the Texas Panhandle into southeastern New Mexico. This line did not coincide with the location of the dryline as it was retreating westward (Figs. 2c and 2d). Cell N at 0130 UTC was the southernmost cell along the cloud line at this time. It is not known if this was coincidental or if there was a causal relationship.

Following this 30-min period of paired, cyclic tornado genesis, tornado 5 developed into a wider, more intense, and longer-track tornado, striking Greensburg almost 50 min later. Unlike tornadoes 1–4, tornado 5 (at its outset) was flanked by a broad (~7 km x 7 km) area of approaching (green) Doppler velocities. Other instances in which a period of cyclic tornado genesis was followed by a longer-track, wide tornado have been observed [e.g., Speheger et al. (2002), tornadoes A1–A8, followed by A9, the Oklahoma City–Moore tornado; Dowell and Bluestein (2002a,b)].

Finally, it is noted from Fig. 12 that during the time there was cyclic tornado genesis (tornadoes 1–4), the damage paths were to the left of the storm motion (toward the north or north-northeast), as noted earlier, but after cyclic tornado genesis, when there was just one, large tornado, the damage path was more closely aligned with the track of the parent storm (Figs. 6 and 12), to the northeast. Although the radar echo of storm N was not tracked (in this paper) after 0148 UTC, it can be seen from Figs. 6 and 12 that storm N turned to the right thereafter as it headed toward Greensburg (a line extending from cell N at 0144 UTC in Fig. 6 to GRE would be to the right of the earlier track and about at the same heading as that of the damage path of tornado 5).

4. Numerical model guidance

To determine how well available numerical model guidance was able to pinpoint the location of the Greensburg storm’s formation and to represent its evolution, three sources of model data available in real time to the author are considered. These were not the only model guidance available, but they are considered...
because they represent rather different types of model forecasting methods. First, it is seen from the output from the 12-h National Centers for Environmental Prediction (NCEP) 12-km North American Mesoscale (NAM) model valid at 0000 UTC on 5 May (Fig. 15) that an intense trough at 500 hPa was forecasted to be far upstream from southwest Kansas. The model forecasted an area of small amounts of precipitation in far northwestern Oklahoma, as was observed. However, it also forecasted a broad region of light precipitation for west-central and northwest Kansas, which did not occur. The region where cell N formed was approximately northeast of a dry bulge in the 850–500-hPa layer, into a north–south axis of strong potential instability (lifted index of $\sim -6$ or less). At 850 hPa, the region where cell N formed was northeast of a thermal ridge. Overall, considering that moist convective processes were not resolved, the guidance from the NAM was reasonable over northwest Oklahoma, but to a lesser extent to the north, where there was little or no precipitation.

The 3-km Advanced Research Weather Research and Forecasting (ARW) model forecasts with explicit moist convection produced by the National Center for Atmospheric Research/Mesoscale Microscale Meteorology [NCAR/MMM; see Weisman et al. (2008) for a discussion of earlier experiments] and initialized the previous evening at 0000 UTC (Fig. 16) predicted an isolated, intense convective storm in west-central Kansas, which then propagated to the east-northeast. Convective activity in northwest Oklahoma was not predicted at all until perhaps 0100 UTC, along the Kansas border. The latter cells did not develop into intense convective storms. The 24-h ARW therefore got the “right idea” (production of an intense, isolated storm), but in the wrong location and at the wrong time. The MMM ARW model did forecast a dry bulge between 2100 and 0000 UTC at the surface (Fig. 16, bottom panel). The dry bulge at 2100 UTC was forecast to be too far southwest in Kansas (cf. Fig. 2b). The model did show an area of enhanced moisture that extended back to the west at 2100 UTC in the northeastern Texas Panhandle; this area of enhanced moisture propagated to the north and was located along the border between southwest Kansas and the Oklahoma Panhandle at 0000 UTC. The model may therefore have been correctly predicting the westward extension of the surface moist layer near Pampa (Fig. 2b) at 2100 UTC, which was near where the first storms formed.

Ensemble forecasts using the 4-km ARW were run by the Center for Analysis and Prediction of Storms (CAPS) at the University of Oklahoma using 10 ensemble members that represented perturbed states and variations in the grid-scale microphysics and in land surface and PBL physics (Kong et al. 2007). Unlike the
MMM ARW, the CAPS ensemble system was initialized at 2100 UTC. The 24-h forecast valid 2100 UTC, just before cell A developed, showed that some ensemble members produced storms in southwestern and western Kansas, which did not actually occur (Fig. 17). The 27-h forecast valid at 0000 UTC had some ensemble members develop convective storms in various locations from western Oklahoma to northwestern Kansas. By 0000 UTC, however, no storms had formed yet in Kansas. However, the 3-h forecast valid at 0000 UTC had some ensemble members produce convective storms in northwest Oklahoma, as actually did happen. The 6-h forecast valid at 0300 UTC had some ensemble members produce convective storms around northwestern Oklahoma and southwestern Kansas (and northcentral Kansas), as was observed. It is apparent that the short-term ensemble forecasts captured the event better than the longer-range ensemble forecasts or the 24-h control run, but they were not timely enough for advanced guidance.

5. Summary and conclusions

The Greensburg, Kansas, storm of May 2007 had a complex origin, as the discrete growth on the southwest flank of a fourth-generation left-split cell from a mother cell (A) that had also produced three other first generations of left-split cells. Only two other isolated cells (C and E) formed, while the former was short lived as it merged with cell A and the latter did not evolve long or produce severe weather. As such, one wonders whether if the original, isolated cell A had not formed, whether or not there would have been a large tornado that would have struck Greensburg or anywhere else. In the spirit of E. Lorenz’s famous talk in 1972 at the 139th meeting of the American Association for the Advancement of Science, entitled (by P. Merilees) “Predictability: Does the flap of a butterfly’s wings in Brazil set off a tornado in Texas?,” perhaps a secondary title of this paper could be “Does the formation of an isolated cell in the Texas Panhandle near the dryline set off a tornado in Kansas?”
It is clear that to have adequately predicted the complex evolution of the Greensburg storm, the physics of convective processes on the substorm scale would have to have been represented very accurately.

If the Greensburg event was so sensitive to the initial conditions, what hope is there for numerically forecasting such an event? The numerical forecasts using the ARW described herein were done using 3- and 4-km grid spacings. Weisman et al. (1997) have demonstrated that 1-km resolution is needed to simulate line convective systems realistically, while Bryan et al. (2003) have recommended a grid spacing ~100 m. It is not anticipated, then, that any of the existing models can adequately forecast storm-scale events even if the initial conditions were perfect, unless the horizontal resolution was increased. However, even with relatively coarse horizontal resolution, it has been demonstrated that existing models can provide useful guidance on the mode of convection that will occur (Weisman et al. 2008).

The mode of tornadogenesis (cyclic versus noncyclic) seems to have had a relationship with the tornado motion. In the case of the former, the motion of the tornadoes was to the left of that of the parent storm; in the case of the latter, the motion was along with the parent storm. When the transition from cyclic to noncyclic tornadogenesis occurred, a larger-scale region of approaching Doppler velocities from the parent storm was evident. These findings are consistent with those of Dowell and Bluestein (2002b) in that there was apparently a mismatch between vortex motion and storm motion during cyclic tornadogenesis and an approximate match between vortex motion and storm motion during noncyclic/long-lived tornadogenesis. It was speculated in Dowell and Bluestein (2002b) that the latter occurred when stronger outflow from the storm forced the vortex along more quickly.

Other significant issues that this case study raises are as follows:

**FIG. 12.** Storm damage tracks from five tornadoes on 4/5 May 2007. See the text for explanations. Beginning and ending times estimated in CDT from DDC WSR-88D vortex signatures and their correlation with the damage paths (M. Unscheid, NWS, Dodge City, KS, 2008, personal communication). Location of the University of Massachusetts XPol is at “R.” (Data provided by the NWS at Dodge City, KS.)
1) What causes some splitting storms to split again repeatedly, while others split only once? What are the relative roles of environmental shear, outflow (cold pool) intensity, and environmental stability? The process of the emitting of left-split cells in this case looks similar to that of the discrete propagation in ordinary cells, owing to periodic regeneration from outflow in multicell convection (Fovell and Dailey 1995; Fovell and Tan 1998). In the case of the predecessors to the Greensburg storm, the period between new pulses was relatively long so that each successive pulse produced a separate new storm, and the new cell growth occurred downstream from the predecessor storm. The formation of new cells on

Fig. 13. (left) Radar reflectivity (dBZ) for the times shown in UTC on 5 May 2007 and (right) Doppler velocity (m s$^{-1}$) the elevation angles. Range markings are given every 5 km. Hook echoes/vortex signatures 1–5 marked by arrows (reflectivity) and circles (Doppler velocity). See the text for a more detailed description of signatures 1–5. Data from the University of Massachusetts XPol at the locations noted in Figs. 6 and 12.
the forward flanks of storms on 4 May 2007 is consistent with Rotunno–Klemp–Weisman (RKW) theory (Rotunno et al. 1988) in that the generation of horizontal vorticity along the northern edge of cool outflow from a convective cell (i.e., in approximately the easterly direction) is of the opposite sign of the horizontal vorticity imported from the environment (i.e., in approximately the westerly direction; see the low-level, southerly/south-southwesterly vertical shear depicted in Fig. 4b at 0100 UTC) (Fig. 18). In this case, deep lifting and hence the triggering of new cells is most likely on the northern sides of cells. It is not known, however, to what extent the regeneration processes studied in multicell convective systems can be carried over to isolated cells in a supercell environment.

2) What controls the difference between convective storms that exhibit cyclical development from those that do not? Adlerman and Droegemeier (2002, 2005) have demonstrated how sensitive the mode of cyclical development is to the environmental shear profile and to the model physics and grid spacing.

Fig. 14. Satellite imagery, as in Fig. 5, but at 0125 UTC 5 May 2007.

Fig. 15. Four-panel 12-h forecast from the NCEP 12-km NAM at 0000 UTC 5 May 2007. Shown are the (top left) 850-hPa height contours (white) and temperature (color shades); (top right) 300-hPa height contours (white) and wind speeds (color shades); (bottom left) surface isobars (light blue), 1000–500-hPa thickness contours (yellow), and 12-h total precipitation (color shades); and (bottom right) 850–500-hPa mean relative humidity (color shades) and lifted index (white contours).
Early on, circulations were relatively small in scale and there was no evidence of widespread approaching Doppler velocities; later, when the Greensburg tornado (tornado 5) formed, there was Doppler radar evidence of a broader circulation. The first tornadoes in cell A formed before the environmental shear had increased as the LLJ formed. The shear had increased significantly even before the Greensburg tornado had formed, when only smaller and weaker tornadoes were produced cyclically. Thus, while an increase in low-level shear may be necessary for strong tornadoes, it apparently is not for weaker tornadoes.

3) What was the mechanism by which some storms could form on the rear flanks of parent storms while others formed on the forward flanks? What are the relative roles of vertical perturbation pressure gradients and gust front–density current lift?

The ARW predictions were made using initial data 24 h or more before the forecast time, owing to the limited availability of computer time (M. Weisman, NCAR, 2008, personal communication). It would be desirable to judge the accuracy of forecasts made on the basis of observations only 12 h prior to the forecast time.
It is hoped that this study will stimulate other studies involving the sensitivity of storm behavior and tornadogenesis to initial conditions and model physics and computational parameters, and fundamental studies of the dynamics of left-split cells in highly sheared environments.

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imagery courtesy of NCAR. Figure 15 courtesy of Unisys. Figure 16 data courtesy of NCAR/MMM. Figure 17 courtesy of F. Kong (CAPS).

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**FIG. 18.** Depiction of horizontal vorticity (curved streamlines) generated baroclinically in the +x direction (denoted by “+”) along the northern edge of a cold outflow boundary and horizontal vorticity in the −x direction (denoted by “−”) in the environment in the lowest 500 m (associated with vertical shear; wind vectors are plotted to the right as a function of height), based on the hodograph shown in Fig. 4b at 0100 UTC.


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