An Overview of Environmental Conditions and Forecast Implications of the 3 May 1999 Tornado Outbreak

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ABSTRACT

An overview of conditions associated with the Oklahoma–Kansas tornado outbreak of 3 May 1999 is presented, with emphasis on the evolution of environmental and supercellular characteristics most relevant to the prediction of violent tornado episodes. This examination provides a unique perspective of the event by combining analyses of remote observational data and numerical guidance with direct observations of the event in the field by forecasters and other observers. The 3 May 1999 outbreak included two prolific supercells that produced several violent tornadoes, with ambient parameters comparable to those of past tornado outbreaks in the southern and central Great Plains. However, not all aspects leading to the evening of 3 May unambiguously favored a major tornado outbreak. The problems that faced operational forecasters at the Storm Prediction Center are discussed in the context of this outbreak, including environmental shear and instability, subtle processes contributing to convective initiation, the roles of preexisting boundaries, and storm-relative flow. This examination reveals several specific aspects where conceptual models are deficient and/or additional research is warranted.

1. Introduction

During the late afternoon and evening hours of 3 May 1999, a violent tornado outbreak affected portions of central and northern Oklahoma, and southern Kansas. A total of 69 tornadoes were documented from the 10 tornadic supercells (Fig. 1a) that developed over the southern plains that afternoon and evening (National Climatic Data Center 1999). (Other tornado-producing thunderstorms occurred well after 0600 UTC 4 May 1999, but those events were not a direct continuation of the afternoon and evening outbreak and are not included here.) Long-lived, violent (F4–F5 damage) tornadoes occurred in the Oklahoma City, Oklahoma, and Wichita, Kansas, metropolitan areas, as well as in the small towns of Mulhall and Dover to the north and northwest of Oklahoma City. The number of strong and violent tornadoes on 3 May 1999 was comparable to that of 26 April 1991, the most recent major tornado outbreak to affect Oklahoma and Kansas. The environmental buoyancy and shear profiles on 3 May resembled those of several violent Oklahoma tornado events in the past two decades.

However, the 3 May case presented several difficult forecast problems. Surface plots of observations from the Oklahoma Mesonet (Brock et al. 1995) showed that the initial supercell developed to the east of two weakly convergent drylines, and west of a subtle convergence axis located across southern and central Oklahoma. A thick cirrus overcast reduced heating and boundary layer mixing near and west of the drylines through much of the afternoon. High clouds are generally viewed as a hindrance to convective initiation during the warm season in the Great Plains. However, the initial supercell developed within a break in the thick cirrus overcast, where observed and model forecast soundings suggested that surface temperatures were sufficiently warm to minimize convective inhibition. Additionally, the evolution of the middle- and upper-tropospheric flow was not forecast well by the 0000 and 1200 UTC 3 May 1999 operational models prior to the outbreak. National Demonstration Profiler Network time series (locations mapped in Fig. 1b) revealed the progression of a well-defined jet streak 4–10 km above ground level from eastern New Mexico and western Texas during the late morning of 3 May to western and central Oklahoma that evening. The upper-tropospheric jet streak contributed to greater deep-layer vertical shear than forecast by the operational models, and may have been accompanied by weak large-scale ascent over western Oklahoma during the afternoon.

The intent here is to document the environment and evolution of the 3 May 1999 tornado outbreak from the view of operational forecasters. Field observations by the authors, and other observers, augment the documentation of several tornadic supercells in the outbreak.
We believe this unique perspective offers the opportunity to describe the event from the synoptic scale down to the storm scale, and to focus attention on the difficulties faced by operational severe storm forecasters regarding convective initiation and morphology. It is hoped that concerns raised by this study can help focus future severe storm research, with the goal of benefiting operational convective outlooks, watches, and warnings.

The evolution of the synoptic-scale environment is presented in section 2, with specific emphasis on the distributions of moisture, instability, and vertical wind shear. Convective initiation and the morphologies of the individual tornadic supercells are addressed in section 3. Section 4 discusses typical and atypical aspects of the 3 May 1999 event, while section 5 summarizes the important aspects of this major regional tornado outbreak and directs attention to the most critical operational forecast concerns in its wake.

2. Synoptic environment evolution

a. Moisture and instability

A mean large-scale trough was located over the western United States at 1200 UTC 3 May (Fig. 2a), with an embedded short-wave trough over Arizona. The large-scale trough amplified over the Rockies by 0000 UTC 4 May, while the embedded short-wave trough
Fig. 3. Conventional station model plots and surface analyses for 3 May 1999 at (a) 0000, (b) 1200, and (c) 2000 UTC. Surface boundary symbols follow standard conventions, isobars are drawn every 4 mb, and the region of >64°F dewpoints is shaded. A smaller-scale analysis of isotherms (every 4°F) and the 68°F isodrosotherm is shown in (d). Dryline boundaries are drawn on the warm side of each resolvable progression from Arizona to western Oklahoma and Kansas (Figs. 2b and 6e). In association with this deepening midlevel trough and southwesterly flow over the southern and central Rockies, a deepening surface low was located over the central high plains, with low-level south-to-southeast flow in the warm sector over Kansas, Oklahoma, and Texas.

A sequence of 3 May surface analyses and data plots that preceded the tornado outbreak is shown in Fig. 3. Surface dewpoints in the middle-to-upper 60°F (18°C±20°C) spread northward from central Texas at 0000 UTC (Fig. 3a), to central Oklahoma by 1200 UTC (Fig. 3b), and southern Kansas by 2000 UTC (Fig. 3c). Figure 3d provides a mesoscale view of surface conditions at 2000 UTC in proximity to the initiation of storms A and B, roughly 30 min prior to the earliest deep convection that can be traced to storm A. Diffuse drylines were analyzed along the moist side of two moisture gradients resolvable in Oklahoma Mesonet data. The western boundary denoted what has been traditionally considered the dryline (e.g., Schaefer 1974; Doswell 1982). The moisture gradient across the eastern boundary was weaker than typically observed with a dryline [e.g., a dewpoint change of 10°F or more between observing stations, after Schaefer (1974)], though the small range of dewpoints (roughly 58°F–61°F) between the two analyzed boundaries was supportive of two transition zones in the surface moisture field. The diffuse boundary structures in Fig. 3 apparently were less important than the fact that dewpoints had increased to the upper 60°F (∼20°C) and surface temperatures had warmed to the middle 80°F (∼30°C) by 2000 UTC in extreme southwestern Oklahoma, which contributed to very large convective available potential energy (CAPE) values and weak convective inhibition (Fig. 4).

The 1200 UTC Norman (OUN) sounding (Fig. 5, background) revealed a moist boundary layer about 1 km deep, beneath an elevated mixed layer located from about 825 to 600 mb. A special 1800 UTC sounding from OUN (not shown) revealed some deepening of the boundary layer from 1200–1800 UTC, with 2°C–3°C warming between 700 and 500 mb. The 0000 UTC 4
May OUN sounding (Fig. 5, foreground) indicated substantial (2°–4°C) warming from 600 mb to the surface during the previous 12 h, with the majority of the warming from 1800 to 0000 UTC confined to the 850–750-mb layer.

Boundary layer depth did not increase substantially after 1800 UTC, although the magnitude of boundary layer dewpoints did increase by about 2°C. Soundings from Fort Worth, Texas (FWD, not shown), revealed low-level warming and moistening similar to that observed at Norman. Both the 1800 and 0000 UTC OUN soundings, modified for surface conditions in proximity to the initial stages of storm A (temperature of 29.5°C and dewpoint of 20.5°C), yielded a mean boundary layer–based CAPE¹ near 5000 J kg⁻¹, with less than 10 J kg⁻¹ convective inhibition. These CAPE and convective inhibition values were in good agreement with short-term model soundings for southwestern Oklahoma. The unmodified mean CAPE in the 0000 UTC OUN sounding was about 2200 J kg⁻¹, with a convective inhibition near 100 J kg⁻¹, although the truncated sounding did not allow a complete estimate of CAPE. The moist profile above 500 mb in the 0000 UTC sounding denoted rawinsonde penetration of the anvil of storm A while that storm was producing tornadoes about 25 mi (40 km) to the west-southwest of Norman.

b. Vertical shear

As destabilization continued over the southern plains, an upper-tropospheric jet streak moved east-northeastward over Arizona and New Mexico from the mean trough position over the western United States. Although the magnitude of this jet streak was not well resolved in the 1200 UTC 3 May soundings, a much clearer indication of the strength of the speed maximum

¹ CAPE values were calculated with the virtual temperature correction described by Doswell and Rasmussen (1994), based on the mean parcel in the lowest 100 mb. Convective inhibition used the “nonvirtual” lifted parcels, which resulted in larger inhibition values.
was provided by several profilers during the day. Time–
height series of profiler winds at Tucumcari, New Mex-
ico, from 1100 to 1600 UTC on 3 May (Fig. 6a), and
later plots from Jayton, Texas (Fig. 6b), Haviland, Kan-
sas (Fig. 6c), and Purcell, Oklahoma (Fig. 6d), generally
showed 20–45-kt strengthening of the flow in the 4–10-
km layer from midmorning over eastern New Mexico
to late afternoon over western and central Oklahoma.
Thereafter, the passage of a speed maximum in the 4–10-
km layer over western Oklahoma was shown clearly by
decreasing wind speeds in the 0000–0400 UTC 4 May
profiler time series at Vici (Fig. 6e). The observed
strengthening of the lower- and middle-tropospheric
flow during the afternoon, generally below 8 km, in-
dicated enhanced supercell potential based on simula-
tions (e.g., Weisman and Klemp 1982; Wilhelmson and
Klemp 1978). In conjunction with modest backing and
strengthening low-level winds noted at Purcell, increas-
ing middle- and upper-tropospheric winds resulted in
sufficient deep-layered vertical shear for supercells by
early afternoon in the warm sector [e.g., the shear term
from the bulk Richardson number (BRN), after Weisman
and Klemp (1982)].

The velocity–azimuth display (VAD) wind profile
from the Frederick, Oklahoma, Weather Surveillance
Radar-1988 Doppler (WSR-88D) (Fig. 7a) revealed suf-
cient deep-layered vertical shear for supercells (e.g.,
BRN shear values around 55 m² s⁻²) near the time and
location of convective initiation. The hodograph was
relatively straight through approximately 4 km (Fig. 7a)
and supported both left- and right-moving supercells
based on observations and numerical simulations (i.e.,
Klemp and Wilhelmson 1978; Wilhelmson and Klemp
1978). Observed motions of the mature left and right
splits associated with storm B (closer to Frederick than
storm A) revealed 0–3-km storm-relative helicity (SRH)
values of 120 m² s⁻² for the right-mover, and −46 m²
s⁻² for the left-mover. Farther east at Purcell, the 0–6-
km shear magnitude increased from about 25 m s⁻¹ (48
kt) at both 1800 UTC (not shown) and 2300 UTC 3
May (Fig. 6d), to 30 m s⁻¹ (57 kt) by 0200 UTC 4 May
(Fig. 7b). BRN shear values also increased from 62, to
123, to 166 m² s⁻² at Purcell at these same times, re-
spectively. Hodographs derived from the Purcell winds
yielded a dramatic increase in 0–3-km SRH values from
roughly 80 m² s⁻² at 1800 UTC to 338 m² s⁻² by 2300
UTC (Fig. 6d). Calculated SRH values at 2300 UTC
were based on the observed motion of storm A from
235° at 12 m s⁻¹ (23 kt) and the surface wind from 140°
at 8 m s⁻¹ (15 kt) at Purcell. The SRH values from the
Purcell profiler were also in close agreement with those
calculated from the 0000 UTC Norman sounding (Fig.
5), and the VAD wind profile from the nearby Twin
Lakes WSR-88D (Fig. 15). The hodographs assumed
pronounced clockwise curvature, all of which strongly
favored right-moving supercells. The large temporal
variability in SRH during the late afternoon hours was
similar to that noted in previous studies of tornadic su-
percell environments by Davies-Jones (1993) and Mar-
kowski et al. (1998, hereafter M98). After the increase
from mid- to late afternoon, SRH values across central
Oklahoma remained in the range of 350–400 m² s⁻²
through the early evening for the observed right-moving
supercell motions to the northeast at approximately 13
m s⁻¹ (25 kt; Fig. 7b).

3. Convective initiation and storm-scale
observations

All supercells were preceded by the development of
two short-lived convective towers over northwestern
Texas. This convection formed under a relative gap in
the cirrus canopy and between the surface drylines. The
initial area of cumulus is shown in Fig. 8a, and the more
substantial of these towers is noted in Fig. 8b. The cirrus
gap shifted northeastward across southwestern
Oklahoma, wherein additional towering cumulus
formed southwest of Lawton around 2030–2045 UTC
(Fig. 8c). This convection rapidly evolved into a storm
split and the first right-moving supercell, storm A in
Fig. 1a. Storm B developed explosively within a small
cluster of updrafts west of Altus around 2115–2130
UTC (Fig. 8d), just west of the eastern dryline. Figure
8e shows storm A as it moved across the confluen
tion boundary about 20 min prior to the first significant tor-
ando of the outbreak, and as the right split of storm B
developed mature supercell characteristics. These two
storms were the most prolific tornado producers of the
outbreak with a combined total of 35, including the F5
tornado that moved across the southern Oklahoma City metropolitan area and an F4 tornado that hit Abell and Mulhall. All mention of supercells hereafter refers to the right-moving storms of the 3 May 1999 outbreak.

Other damaging tornadic supercells developed south and west of Oklahoma City around 0030 UTC. Storm D tracked from Purcell to Stroud (Fig. 1), resulting in F3 tornado damage in Stroud around 0330 UTC, while storm E produced a tornado with F4 damage in Dover at about 0230 UTC. Storm D appeared to form near the southeast–northwest-oriented confluence line to the southwest of Purcell, while storm E formed near the intersection of the confluence line and the eastern dry-line. Another supercell (storm K) produced a violent tornado on the south side of Wichita, Kansas, around 0130 UTC (Fig. 1).

Supercell character and structure

The central Oklahoma thunderstorms displayed the visual characteristics of classic supercells (e.g., Lemon and Doswell 1979): large, striated updrafts well-removed from the main precipitation core, rotating wall clouds with associated clear slots, and thin precipitation curtains wrapping around the west side of the low-level mesocyclones. Pictures of these storms (Fig. 9) clearly...
Fig. 6. National Profiler Demonstration Network time series of vertical wind profiles and resultant hodographs from (a) 1100–1600 UTC 3 May at Tucumcari, NM; (b) 1600–2300 UTC at Jayton, TX; (c) 1600–2300 UTC at Haviland, KS; (d) 1600–2300 UTC at Purcell, OK; and (e) 0000–0400 UTC 4 May at Vici, OK. Plotted wind barbs are in kt, with time increasing from right to left, and each hodograph is for the latest time in each plot. The heights above ground level of each 3-km-deep layer are labeled along the hodographs. Note that data were unavailable from 2000 to 2100 UTC.
illustrate the classic supercell storm structures. Storm A and storm B each had laminar cloud structures in the low levels, which indicated forced ascent of parcels through the layer of convective inhibition over central Oklahoma noted in the unmodified 0000 UTC 4 May OUN sounding (northeast of where storms A and B developed).

Storms A and B each were tornadic periodically for 4–6 h. Each appeared to reach a state of balance with its environment, during which it was neither detrimen-
FIG. 8. Series of GOES-8 1-km visible satellite images and associated surface station data plots for (a) 1902, (b) 2002, (c) 2045, (d) 2132, and (e) 2202 UTC. Annotations denote the cloud features discussed in the text, with the surface dryline position(s) marked by the open
Continued scalloped lines, and a subtle confluence zone across southern Oklahoma marked by a dashed line. Plotted mesonet observations are within 2 min of each respective satellite image, while standard surface observations are from within the hour before each image.

Thunderstorm outbreak: steep midlevel lapse rates overspread the area, boundary layer moisture increased, and daytime heating further increased instability. In these regards, 3 May 1999 resembled a “synoptically evident” severe weather episode as originally defined by Doswell et al. (1993). However, anticipation of a major tornado outbreak was hindered by poor operational model forecasts of wind speeds in the middle and upper troposphere, and associated vertical shear parameters, over portions of the West Coast and southern plains (Fig. 10).

Large-scale pattern recognition, based on numerical model forecasts, did not necessarily suggest a major tornado outbreak would ensue the evening of 3 May 1999. The perceived problems with the pattern were derived from operational model forecasts of, at best, modest vertical wind shear (Fig. 11a), and surface analyses that did not provide a clear focus for convective initiation. However, the combinations of large buoyancy (in observations and model forecasts), and vertical shear parameters derived from regional profiler data and soundings during the afternoon and evening of 3 May (Fig. 11b), were consistent with other historical tornado outbreaks. To further illustrate this point, 0000 UTC hodographs for OUN and OKC are compared in Fig. 12 for four violent tornado events in central and northern Oklahoma: 4 May 1999, 27 April 1991, 27 April 1984, and 23 May 1981. In each event, the sounding site was uncontaminated in the low levels by convective outflow, although thermal lapse rates on the 3 May sounding decreased as the radiosonde entered the anvil of storm A at the 551-mb pressure level. Each sounding showed a deep layer of strong positive buoyancy (not shown), and the hodograph structures were remarkably similar in the lowest 3 km. Interestingly, each of the hodographs displayed a pronounced “kink” within the 1.0–1.5-km above ground level (AGL) layer, where a combination of weak veering with height and strong speed shear suddenly changed to strong veering with little speed change. This feature enhances storm-relative inflow in the boundary layer, and may be related to low-level mesocyclone intensity and associated tornado potential based on the results of model simulations by Wicker (1996).

b. Convective initiation and low-level boundaries

Surface analyses revealed a dryline on the mesoalpha to synoptic scales, which appeared to be a diffuse double-dryline structure in finer-scale analyses over northwestern Texas and western Oklahoma by midafternoon on 3 May 1999 (Fig. 3). A deep surface low existed well to the northwest of Oklahoma. However, surface convergence along each 3 May dryline was ill-defined, and the moisture gradient was not particularly large across either dryline. Ziegler and Hane (1993) suggested that boundary layer convergence is critical for maintaining a pronounced moisture gradient across a dryline, as well as for thunderstorm initiation along it. In fact, the initial supercell developed about 50 km east of the easternmost surface dryline position (Figs. 3d and 8c). Also, a large plume of cirrus developed in the lee of the Rockies over eastern New Mexico by midmorning on 3 May. The cirrus overspread much of the Texas Panhandle and western Oklahoma by early afternoon (Fig. 8a, and raised questions about continued heating and mixing in the boundary layer through midafternoon. Small temperature decreases and dewpoint increases across the central and southern Texas Panhandle from 2000 to 2200 UTC (Figs. 8b and 8e), combined with small temperature increases and dewpoint decreases farther north in the Oklahoma Panhandle, indicated shading and weaker vertical mixing under the dense high cloud canopy across the Texas Panhandle. The widespread high clouds and lack of convergence in the dryline regions introduced considerable uncertainty regarding the timing and location of convective initiation. However, the gap in the cirrus appeared to be crucial in allowing continued surface heating and mixing to maintain weak enough capping for the initiation of both short-lived cumulonimbi in northwestern Texas and long-lived storms A and B in southwestern Oklahoma.

The 0000 UTC OUN sounding, modified for surface mesonet observations in southwestern Oklahoma, showed little convective inhibition and a level of free
convection (LFC) within 2 km of the ground. This thermodynamic profile suggested that relatively weak mesoscale lift may have been sufficient to initiate deep convection. Forecast fields from the operational Eta Model runs at 1200 and 1800 UTC 3 May showed divergence in the 300–200-mb layer and weak quasigeostrophic forcing for ascent over western Oklahoma from 500 to 300 mb between 1800 UTC 3 May and 0000 UTC 4 May. The model also suggested weak ascent from 850 to 700 mb, although ascent over western and central Oklahoma may have been offset by the warming noted in this layer from 1800–0000 UTC at OUN. Upper-tropospheric divergence and possible weak inertial instability in the upper troposphere (Fig. 13) may have also contributed to a favorable environment for strong storm-top divergence and sustained updrafts, perhaps similar to the arguments presented by Blanchard et al. (1998).

On a smaller scale, storm A appears to have formed near the updraft portion of a large horizontal convective roll (HCR) in the boundary layer. Though no cloud streets were observed in 1-km visible satellite imagery, the apparent HCR was denoted by a meridional band in 1.5° elevation base reflectivity data from the Frederick radar site; the radar site was located about 25 km west of where storm A began (Fig. 14). The fine line was oriented parallel to the boundary layer flow, and Wilson et al. (1994) have established coincidence between convective boundary layer updrafts and reflectivity fine lines in clear air mode. This feature conformed to radar and shear-based characteristics for HCRs established by Weckwerth et al. (1997), who also found that HCRs are sometimes not apparent as visible cloud streets. The depth of the feature (1400–1500 m above ground level) appeared to be within a few hundred meters of the LFC height derived from modified soundings (roughly 1700 m above ground level); therefore, an HCR updraft was a potential mechanism for convective initiation. While such a mode of thunderstorm formation may be common, the presence and depth of HCRs are difficult to anticipate and detect in an operational setting. Forecasts for convective initiation will necessarily have large uncertainty when HCRs or other subtle boundary layer processes dominate, given the sensitivity of moist convection to temperature and moisture in the boundary layer (e.g., Crook 1996).

Finally, visible satellite imagery revealed a series of billow clouds oriented meridionally over northern Texas and south-central Oklahoma (Fig. 8a), to the east of where storms A and B developed. These clouds were associated with a relatively shallow, capped boundary layer with backed surface winds compared to northwestern Texas and southwestern Oklahoma. As the southwestern Oklahoma storms approached central Oklahoma, they encountered the confluent zone and associated backed surface winds, which yielded 0–3-km SRH values of 300–400 m² s⁻² (Figs. 6d and 7b). Both storms A and B produced brief tornadoes near and west of the subtle confluence boundary, but tornadoes became more numerous and progressively more intense as the storms moved through the region of increased SRH and lower LCL heights to its east. The air mass in this region, which coincided with the location of the billows prior to being obscured by high clouds, still contained surface-based CAPE values in excess of 3000 J kg⁻¹. However, the longevity of the storms after 0000 UTC was in question operationally because of the stronger cap well east of the confluence boundary. Indeed, storm A ceased producing tornadoes by 0130 UTC (less than an hour after devastating the southern Oklahoma City area), and dissipated shortly after 0200 UTC (see Fig. 9e). To complicate matters further, storm D reached peak intensity only 30 mi south of where storm A dissipated. These observations underscored the difficulties faced by forecasters in anticipating storm longevity and tornado production in an evolving environment with subtle, but important, variations.

The tendency of storms A and B to produce significant tornadoes after crossing the surface boundary appears to be consistent with the observations of M98, although the confluence boundary in this case was subtle, even in the relatively dense Oklahoma Mesonet surface observations. Another storm–boundary interaction occurred farther north, when the Wichita supercell (storm K) crossed a boundary marked by a westward-moving band in 0.5° elevation reflectivity data, thereafter producing a violent tornado.

c. Variations from classic supercell structures

At least two possible cases of destructive storm interference were observed during the tornado outbreak, namely with storms D and K. The Wichita supercell (storm K) weakened substantially after being overtaken by a large area of thunderstorms from the southwest, while the Purcell–Stroud supercell (storm D) also was structurally altered by convection that formed and merged with the supercell from the southwest. Trends
in radar imagery suggested that storm D evolved in a manner quite similar to the other violent tornadic supercells in central Oklahoma until the disruption of its classic structure by the merging precipitation areas.

In addition, not all wind profile parameters were strongly supportive of classic supercells with major tornadoes. A model-forecast weakness in storm-relative winds in the middle troposphere was seen as a potential limiting factor for significant tornadoes at the onset of the outbreak. It has been suggested by Brooks et al. (1994) and Thompson (1998) that weak storm-relative winds in the middle troposphere can lead to excessive cold outflow generation in the rear-flank region of a supercell. This cold outflow can then undercut the mid-
level mesocyclone prior to the formation of significant tornadoes, if not balanced by sufficiently strong storm inflow. Both the Purcell profiler at 2300 UTC 3 May and the Norman sounding at 0000 UTC 4 May revealed storm-relative winds in the 4–6-km layer that appeared to be marginal for sustained supercells with significant tornadoes (after Thompson 1998). However, actual wind speeds in the 3–6-km layer increased from 17 to 25 m s$^{-1}$ (about 35–50 kt) by 0200 UTC (Fig. 7b). Storm-relative winds derived from the Twin Lakes WSR-88D VAD wind profile, located about 15 mi north-northeast of the Purcell profiler, also were similar to those from Purcell. The severity of the 3 May tornado outbreak tended to increase in conjunction with increasing midtropospheric storm-relative flow at the Purcell profiler site from 2300 to 0200 UTC, and as multiple supercells encountered the region of enhanced SRH to the northeast of the subtle confluence boundary.

Interestingly, the wind profiles from the Purcell profiler, the Norman sounding, and the Twin Lakes radar site all conformed to common “proximity” standards for storm A (e.g., Darkow 1969; Brooks et al. 1994). However, storm A was located 25 mi (40 km) or more to the west of these sites until about 2330 UTC and, therefore, somewhat closer spatially to the intensifying flow fields observed at Tucumcari and Jayton earlier in the day. A dramatic short-term increase in vertical wind shear and storm-relative winds also was indicated by the Twin Lakes WSR-88D VAD wind profile from 2348 UTC 3 May to 0018 UTC 4 May (Fig. 15), as storm A approached and passed only 10 km northwest of the radar site. This observation supports simulations by Weisman et al. (1998), which yielded 0–6-km shear perturbations of 8 m s$^{-1}$ (about 15 kt) or more extending nearly 30 km ahead of a supercell for hodographs comparable to those observed in central Oklahoma. The general increase in wind speeds and vertical shear during the period 2300–0200 UTC, as well as the potential influence of storm A on the Twin Lakes VAD wind profile, suggest that there may not have been a single wind profile that correctly characterized the environment of storm A.

5. Summary and implications

The 3 May 1999 event in Oklahoma showed some general large-scale characteristics of historical severe thunderstorm outbreaks in the southern and central plains. A mean trough was located over the Four Corners area with a deep high plains surface low and an unstable warm sector. An embedded mid- to upper-tropospheric jet streak moved east-northeastward from the mean trough during the afternoon of the outbreak. However, the embedded jet streak was not resolved well by the operational models prior to the outbreak, and surface convergence in proximity to the drylines was ill-defined through the afternoon. Additionally, a large area of high clouds overspread the drylines and warm sector during the afternoon, which complicated forecasts of convective initiation.

There are several important points to be learned from the 3 May outbreak. First, this event illustrates that outbreaks of strong and violent tornadoes are not necessarily associated with what many operational forecasters would consider to be the most evident large-scale patterns in numerical model output. While the potential for a severe weather episode was anticipated by Storm Pre-
Between its development and violent tornado production, storm A appeared to shift markedly from the high-CAPE and low-shear extreme of the Johns et al. (1993) diagram toward the midrange of the parameter space (Fig. 16). Also, the kinematic observations in central Oklahoma from 2300–0200 UTC suggested that storm characteristics may not always be predictable, especially when the mesoscale storm environment is changing with time. This problem is compounded when a storm simultaneously affects nearby observations as evident in the Twin Lakes VAD wind profile data just after 0000 UTC (Fig. 15), that is, when the storm’s influence may be superimposed on changes in the background wind profile. This illustrates the need for operational forecasters and researchers alike to consider the CAPE–shear parameter space for any given thunderstorm as fluid with respect to time, not static. The concept of significant, deep tropospheric environmental evolution throughout an individual storm’s lifetime should become a fundamental inclusion in storm-scale numerical modeling studies of supercells.

When forecasting a threat of tornadoes, the mode of convective initiation and the number and spacing of supercells that form are critical to the number of tornadoes expected. In the same mesoscale region, several supercells may develop in association with different forms of boundaries. These boundaries vary in detectability when using conventional data sources, and storms may form where there are no apparent boundaries. The initial storms in the 3 May 1999 outbreak evolved into tornadic supercells that each lasted several hours, with no early transition to a squall line or other convective mode. Storm spacing and motions were such that the supercells remained in an environment of favorable vertical shear and instability for several hours without numerous storm collisions, thus allowing the supercells to produce a large number of tornadoes.

The predominance of a supercell convective mode and lack of a squall line on 3 May 1999 may have been attributable to the lack of strong low-level convergence...
near the dryline(s). It is conceivable that the outbreak would not have materialized in such intense or prolific form had the convergence been stronger along a consolidated dryline, and had numerous storms formed simultaneously and merged into a larger-scale convective system in the weakly capped environment over northwestern Texas and western Oklahoma during the afternoon. However, on forecast times greater than mesoscale, that same lack of convergence in the area of drylines suggested that supercells might not develop at all, consistent with the ideas of Ziegler and Hane (1993). This presented a major forecasting challenge in two ways: 1) narrowing the spatial threat for supercell concentration on the synoptic timescale (roughly 12 h or more), and 2) diagnosing and nowcasting subtle mesoscale features not associated with distinct thermodynamic discontinuities (such as fronts, singular drylines, outflow boundaries, etc.).

In this case, subtle clues/precursors to storm initiation were present primarily in high-resolution visible satellite imagery (e.g., the cumulus towers beneath the cirrus hole) and nearby WSR-88D data (e.g., the possible HCR). The complex dryline, confluence zone, and HCR structures identified during this event each demonstrate the importance of high-resolution, lower-tropospheric observational data to the real-time diagnosis of subtle boundaries and their subsequent effects on storm initiation and organization. As illustrated in our analyses...

Fig. 13. Eta Model initial analyses of 300-mb geopotential height (dam), divergence (solid lines $1 \times 10^{-5}$ s$^{-1}$, dashed lines represent convergence), and negative absolute geostrophic vorticity (shaded area, with gradations at $0$, $-1 \times 10^{-5}$ s$^{-1}$, $-5 \times 10^{-5}$ s$^{-1}$, and $-10 \times 10^{-5}$ s$^{-1}$) at (a) 1800 UTC 3 May 1999, and (b) 0000 UTC 4 May 1999. The shaded areas of negative absolute geostrophic vorticity represent areas of inertial instability. (Courtesy of D. Schultz, National Severe Storms Laboratory.)

Fig. 14. Filtered “clear air mode” radar reflectivity data (dBZ) at the 1.5$^\circ$ elevation angle from Frederick at 2012 UTC 3 May 1999. The annotation identifies the reflectivity signature of a possible HCR immediately east of the radar site. Reflectivity values less than 0 dBZ have been removed for clarity.

Fig. 15. Time series from the Twin Lakes WSR-88D VAD wind profile from the period 2348 UTC 3 May to 0018 UTC 4 May showing the increase in midlevel wind speeds as storm A moved approximately 10 km northwest of the radar site.
tical shear (e.g., 0–6-km shear or BRN shear), and strong
characterized by large CAPE, strong deep-layered ver-
supercells in an otherwise highly favorable environment
preclude significant tornadoes, given the initiation of
absence of such boundaries in the warm sector does not
outbreak. This illustrates that the apparent weakness or
were either ill-defined or not present during the 3 May
et al. 1999). However, low-level baroclinic boundaries
strength and orientation of low-level boundaries (Atkins
face boundaries (M98). Furthermore, recent numerical
Tornadoes Experiment occurred in association with sur-
version of the Verification of the Origins of Rotation in
A majority of significant tornadoes during the 1995
version of the Verification of the Origins of Rotation in
Tornadoes Experiment occurred in association with sur-
facing boundaries (M98). Furthermore, recent numerical
simulations indicated strong relationships between me-
socyclone intensity and longevity, and the baroclinic
occur on the mesoscale, or even storm scale. Profiler
and radar-derived wind data each greatly aid the short-
term forecast process by providing time sampling be-
tween synoptic rawinsonde launches. However, many
supercell-type and tornado forecast parameters are sen-
sitive to small changes (2–5 m s⁻¹) in wind vectors and
storm motion, which lends these parameters to misin-
interpretation. Still, the more general combinations of
CAPE and vertical shear clearly supported tornadic su-
percells by the late afternoon and evening of 3 May
1999. This suggests that the background environment,
as opposed to just storm-scale variations, can be a dom-
inant controlling factor in regional tornado outbreaks.

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![Fig. 16. Scatterplot of lowest 100-mb mean-parcel CAPE (abscissa)
vs 0–2-km SRH (ordinate) for the dataset of 242 significant (F2
damage or greater) tornadic supercells collected by Johns et al.
(1993). Locations of various estimates of CAPE and 0–2-km SRH
from 3 to 4 May 1999 are overlain for comparison, using (a)
unmodified (truncated, with incomplete CAPE) 0000 UTC OUN sound-
ing, (b) 0000 UTC OUN sounding with 1800 UTC thermal profile
attached above truncation, and (c) 1800 UTC OUN sounding modified
using these input conditions as an idealized estimate of proximity
to conditions of the genesis point of storm A: 2100 UTC Tillman County
mesonet wind, temperature, and dewpoint; and Frederick velocity–
azimuth display winds between surface and 2 km AGL.]


