The 6 May 2010 Elevated Supercell during VORTEX2

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ABSTRACT

An elevated supercell from the second Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX2) on 6 May 2010 is investigated. Observations show that the supercell formed over a stable inversion and was likely decoupled from the surface. Quintessential features of a supercell were present, including a hook echo (albeit bent anticyclonically) and midlevel mesocyclone, and the storm was quasi steady during the observing period. A weak surface cold pool formed, but it was apparently devoid of air originating from midlevels. Idealized modeling using near-storm soundings is employed to clarify the structure and maintenance of this supercell. The simulated storm is decoupled from the surface by the stable layer. Additionally, the reflectivity structure of the simulated supercell is strikingly similar to the observed storm, including its peculiar anticyclonic-curving hook echo. Air parcels above 1 km reached their LFCs as a result of the simulated supercell's own dynamic lifting, which likely maintained the main updraft throughout its life. In contrast, low-level air in the simulation followed an “up–down” trajectory, being lifted dynamically within the stable layer before becoming strongly negatively buoyant and descending back to the surface. Up–down parcels originating in the lowest 100 m are shown to be a potential driver of severe surface winds. The complementary observations and simulations highlight a range of processes that may act in concert to maintain supercells in environments lacking surface-based CAPE.

1. Introduction

On 6 May 2010, the second Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX2; Wurman et al. 2012) collected what may be the first storm-following dataset for an elevated supercell.¹ In this study, convection is considered to be elevated when it receives all of its inflow parcels from above the near-surface layer [i.e., the lowest 500 m AGL, as in Parker (2008)]. This occurs most typically in an environment without surface-based CAPE, but with appreciable instability present aloft (Colman 1990a,b; Grant 1995; Moore et al. 1998; Thompson et al. 2003; Horgan et al. 2007; Corfidi et al. 2008). For a storm to form over a stable layer, parcels with instability aloft need to be lifted to their levels of free convection so that the elevated instability can be released. This can be accomplished, for instance, by frontal lifting (e.g., Moore et al. 1998), interactions between gravity currents and preexisting boundaries (e.g., Carbone et al. 1990), or lifting by density currents, bores, and gravity waves (e.g., Rotunno et al. 1988; Koch and Clark 1999; Marsham and Parker 2006; Morcrette et al. 2006; Ziegler et al. 2010; Marsham et al. 2011). The lifting mechanism likely differs on a storm-by-storm basis and depends on micro-, meso-, and synoptic-scale features. The majority of studies on elevated convection have focused on meso-scale convective systems, which may be due to the lack of elevated supercell observations from past field projects. In this study, we focus on the unique dynamics that may sustain a supercell over a stable layer.

Many convective storms are maintained, at least in part, by lifting at the edge of a cold pool (outflow). However, it is not clear that surface cold pools occur in many elevated systems. It is possible that cold outflow could be produced prior to stabilization (e.g., Parker 2008; Ziegler et al. 2010), or that in situ cooling of the stable layer could occur if the relative humidity is suitably low. Alternatively, Bryan and Weisman (2006) simulated the presence of an elevated cold pool (2–4 km layer) above a stable layer, which worked to continually release elevated instability by lifting air along its leading edge. Bores may also provide sufficient lift to sustain elevated convection; Parker (2008) found in his idealized simulations of MCSs in environments with

¹ A supercell is a thunderstorm with a rotating updraft (e.g., Lemon and Doswell 1979).
low-level cooling that as the convection became elevated, a “gravity-wave-like bore” maintained the convective updrafts. For supercells, the enhanced vertical perturbation pressure gradient acceleration has also been shown to help maintain elevated convection above stable layers (Nowotarski et al. 2011; Billings and Parker 2012). It is possible that storm maintenance mechanisms vary from case to case even for a given storm type.

Heavy precipitation (with associated flash flooding) and large hail are seemingly the most common hazards associated with elevated convection (e.g., Grant 1995; Moore et al. 1998; Horgan et al. 2007), but it appears that severe surface winds are at least possible (e.g., Grant 1995; Schmidt and Cotton 1989; Knupp 1996; Bernardet and Cotton 1998; Bryan and Weisman 2006). The processes behind the production of severe surface winds in elevated convection are somewhat mysterious, with various studies concluding that different storm-scale and mesoscale processes were most important. It has long been presumed that downdrafts of elevated storms cannot penetrate through a stable layer to the surface (e.g., Kuchera and Parker 2006). Parcels are thought to be slowed in their descent to the surface because of the reduction of negative buoyancy, presumably reducing the severe threat. In contrast, Knupp (1996) saw “up–down” trajectories associated with the downdrafts of a microburst-producing storm in Colorado. In such cases, parcels in the stable layer initially rise upward as a result of an upward-directed vertical perturbation pressure gradient. This ascent, in turn, causes these parcels to become strongly negatively buoyant and descend rapidly back to the surface. In this case, inflow air was slightly stabilized under the presence of an echo overhang (anvil shading), causing somewhat negatively buoyant air to be lifted over the cold pool, become more negatively buoyant, and descend rapidly to the surface. Schmidt and Cotton (1989) and Bernardet and Cotton (1998) also reported up–down trajectories associated with severe downdrafts in an elevated squall line and an elevated derecho, respectively. Both studies found that the resulting surface winds exceeded severe criteria as a result of the acceleration of the flow by the horizontal pressure gradient in the stable layer.

It is generally thought that elevated storms are less likely to produce tornadoes, as reviewed in the work of Kis and Straka (2010). Billings and Parker (2012) studied several nocturnal tornado-producing storms they originally presumed to be elevated, but they ultimately concluded they were likely ingesting air from the near-surface layer and were not elevated by definition. Such subtle differences are potentially quite important operationally, since base scans of elevated supercells may have comparable values of vertical vorticity to those in surface-based supercells, and are arguably operationally indistinguishable from surface-based storms on radar.

Nowotarski et al. (2011) produced the first comprehensive study of elevated supercells by using a numerical model to compare surface-based supercells to those over stable boundary layers. Interestingly, most of the simulated “elevated” supercells in the Nowotarski et al. (2011) study were still fueled by parcels from the near-surface layer, except in the case of extreme stability [experiments 500mC, 1kmA, 1kmB, and 1kmC in Nowotarski et al. (2011)]. In addition, midlevel downdrafts were able to reach the surface except in the case with the greatest stability [experiment 1kmC in Nowotarski et al. (2011)]. Within the stable layer, weaker convergence and baroclinic vorticity generation were found by Nowotarski et al. (2011), leading to weaker near-surface vertical vorticity. This is one possible reason for the apparently hindered tornado potential in elevated supercells. Notably, the Nowotarski et al. (2011) study used statically stable boundary layers; in contrast, the 6 May 2010 VORTEX2 case exhibited a well-mixed boundary layer, which nevertheless was capped by a temperature inversion and had no surface-based or mixed layer CAPE. It is uncertain how such differences in low-level stability are likely to affect storm evolution and severity.

Given how few other studies exist, many questions remain unanswered. How are parcels with appreciable instability lifted to their levels of free convection? How is this lifting maintained throughout the storm’s existence? How does the stable layer influence the structure and maintenance of an elevated storm? And, when observed, what is the mechanism behind severe surface wind production in the presence of a stable layer? Observations and idealized numerical modeling will be exploited to explain the processes supporting the structure, maintenance, and severe surface wind production of the 6 May 2010 elevated supercell from VORTEX2. Section 2 presents the observations from the 6 May 2010 elevated supercell, including both the evolution of the preconvective environment and the storm itself. The observations lead to several hypotheses that require further testing, which is undertaken by way of idealized numerical modeling in section 3. The dynamics behind the structure, maintenance, and severe surface wind production of a simulated elevated supercell within the observed environment are investigated and compared to several sensitivity tests. Section 4 summarizes the important conclusions from both components of this study and presents avenues for future work.

2. Observations of 6 May 2010

a. VORTEX2 observational platforms and methods

During VORTEX2, an armada of mobile observational platforms was deployed in an effort to learn more about the dynamics of tornadoes and their parent
supercells (Wurman et al. 2012). Many of these platforms were deployed during the 6 May 2010 case. This case was the first storm to be sampled during the 2010 season of VORTEX2 and was therefore a “shake-down” mission for many of the instruments and mobile Doppler radars. Even so, the novelty of the target storm (easternmost storm in Fig. 1) makes it a dataset worth studying. The environment near the supercell was sampled with RS92 Vaisala radiosondes launched using the mobile GPS advanced upper-air sounding system (MGAUS). Temperature, pressure, and humidity were measured directly and winds were calculated based on the GPS location of the balloon. Surface kinematic and thermodynamic observations in the forward-flank, hook-echo, and inflow areas of the storm were made using mobile mesonets (Straka et al. 1996) and StickNets (Weiss and Schroeder 2008; locations shown in Fig. 1). Quality control and bias correction were performed on these surface data as well as a time-to-space conversion to help eliminate biases associated with mobile versus stationary observations (Skinner et al. 2010).

The evolution of the supercell was sampled by a Shared Mobile and Atmospheric Teaching Radar (SMART-R2; Biggerstaff et al. 2005) and two Doppler on Wheels radars (DOW6 and -7; Wurman et al. 1997); the radar locations can be seen in Fig. 1. Because of their close proximity to the supercell, the mobile radars had better data quality than the conventional Doppler radars nearby [the storm quickly moved out of the range of the Goodland, Kansas, Weather Surveillance Radar–1988 Doppler (WSR-88D), while other mobile radars were degraded as a result of various mechanical issues]. The staff at the National Severe Storms Laboratory (NSSL) edited the vast majority of the SMART-R2 data, including removing the second-trip and spurious echoes and deliasing the velocity data; additional editing, including all of that for the DOW data, was manually performed by the authors. Because of a calibration error, DOW reflectivities were roughly 30 dBZ lower than what was observed by surrounding radars; the DOW reflectivities in this paper have been corrected by the authors. In addition, an issue with the DOW antenna controllers resulted in the loss of some data and inconsistent height readings during sampling. The SMART-R2 also suffered from a mechanical issue causing the reflector to detach from the pedestal mount, which culminated in the reported elevation angle being lower than reality. A dual-Doppler synthesis was attempted between DOW6 and SMART-R2; however, these apparent errors in positioning prevented an accurate one from being produced. We present reflectivity and single-Doppler data from these radars subject to the uncertainties associated with the aforementioned technical issues.

b. Observations

1) EVOLUTION OF THE PRECONVECTIVE ENVIRONMENT

A short-wave trough was located over the western United States with a jet streak present over central Colorado at 0000 UTC on 6 May 2010 (Figs. 2a,b). At this time, a surface low was positioned over southwestern Kansas with a diffuse warm front stretching across central and southeastern Kansas (Fig. 2c). A sounding
north of the front (sounding locations can be seen in Fig. 1) at 1745 UTC showed a considerable inversion around 750 hPa with a well-mixed boundary layer beneath (Fig. 3a), an environment without surface-based CAPE. Strong low-level vertical wind shear was present with a clockwise-looping hodograph and 393 m² s⁻² of 0–3-km storm-relative helicity (SRH). Also of note is the somewhat limited surface moisture over much of the region, with surface dewpoints generally around 12°F–15°F (5°C–59°F) in the warm sector over much of southern Kansas versus 5°F–9°F (41°F–48°F) north of the front in northern Kansas and southern Nebraska (Fig. 2c). This unseasonably dry surface air accounts for the absence of surface-based CAPE. Even though the environment north of the warm front was fairly dry and capped by an inversion, large-scale advection and ascent led to the development of most unstable parcel CAPE (MUCAPE) above the inversion by late afternoon (e.g., Fig. 4), such that storm development could occur in the presence of lift. The analyzed 0–6-km shear vector magnitude exceeded 30 m s⁻¹, further suggesting that convection could become supercellular in this environment (Fig. 3a). Convection initiation occurred north of the warm front in southwestern Kansas and northwestern Nebraska around 2230 UTC. The VORTEX2 team targeted a developing supercell north of McDonald, Kansas (39.8°N, 101.4°W; location shown in Fig. 1), at 2343 UTC and maintained operations on this storm until roughly 0140 UTC 7 May.

2) SUPERCELL OBSERVATIONS

Soundings taken in the direct inflow (Fig. 4; see Fig. 1 for sounding locations) of the 6 May 2010 supercell revealed an even more favorable environment than earlier in the day, with MUCAPE of 1000–1200 J kg⁻¹. North-easterly surface winds veering to westerly aloft led to large, looping hodographs, resulting in 0–3-km SRH over 400 m² s⁻². As in the 1745 UTC sounding, each near-storm profile contained a robust frontal inversion near 750 hPa, and lacked surface-based CAPE, suggesting that all resulting convection was elevated. The supercell had the characteristics of the conceptual model of Lemon and Doswell (1979), with a quasi-steady structure that included a hook echo to the southwest of the forward flank (Fig. 5). Midlevel, cyclonic rotation (inbound to the west of outbound) was evident in a column to the east of the aforementioned hook echo (Fig. 6, location annotated by red arrows). Anticyclonic rotation (inbound to the east of outbound) is also present to the west of the mesocyclone in the rear flank (Fig. 6, locations annotated by blue arrows); this will be discussed later.

In the majority of observed supercells (e.g., Markowski 2002), the hook echo bends counterclockwise because of the cyclonic rotation in the main

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**Fig. 2.** NAM analyses of isotachs (kt, where 1 kt = 0.5144 m s⁻¹; shaded), wind barbs, and heights (dam) at (a) 250 and (b) 500 hPa, and (c) surface observations at 0000 UTC 7 May 2010. Approximate locations of the low pressure system (L), dryline (brown line), and cold (blue line) and warm (red line) fronts are also shown.
updraft; however, the hook echo of the 6 May 2010 supercell curved anticyclonically throughout the observing period (Fig. 5). Anticyclonic-curling hook echoes have been observed (Van Tassell 1955; Brandes 1981; Fujita 1981; Fujita and Wakimoto 1982) and may be associated with the anticyclonic member of a vorticity couplet that straddles the hook echo (Brandes 1981; Ray 1976, Ray et al. 1981; Heymsfield 1978; Klemp et al. 1981; Markowski 2002; Grant and van den Heever 2014), which is evident behind the hook echo at a few times (e.g., Fig. 6). Given the severe limitations of the observations, the evolution of vertical vorticity
near the hook echo will be discussed further using simulations in section 3.

Mobile mesonets sampling the inflow environment recorded ambient potential temperature ($\theta_a$) values around 297 K, with a 2–3-K deficit underneath the cell (Fig. 7a). In contrast, equivalent potential temperature ($\theta_e$) values beneath the cell closely matched the ambient inflow values (near 317 K; Fig. 7b). The 0117 UTC
7 May sounding shows that air with \( \theta_e \) near 317 K exists only in the lowest kilometer above the ground (Fig. 7, right). This suggests that the surface cold pool is not driven by descending midlevel air. Instead, rather uniform \( \theta_e \) air at the ground suggests that the surface cold pool was produced by in situ evaporation of precipitation in the near-surface layer. This seems to be consistent with the large ambient dewpoint depressions in the lowest 100 hPa AGL (Fig. 4a). Since equivalent potential temperature is not conserved in the presence of melting, we still consider this result to be speculative and evaluate it further using parcel trajectories in section 3.

Despite the very weak cold pool, the storm eventually grew upscale and became part of a series of line segments spanning northern Kansas (not shown). Near-severe\(^2\) surface winds in advance of these cells were experienced by VORTEX2 team members returning to their lodging in Hays, Kansas, with measured peak gusts of 21 and 24 m s\(^{-1}\) occurring at Hill City (0408 UTC 7 May; 39.4\(^{\circ}\)N, 99.8\(^{\circ}\)W) and Russell (0454 UTC 7 May; 38.9\(^{\circ}\)N, 98.9\(^{\circ}\)W), Kansas, respectively. Unfortunately, these observations occurred after the period of VORTEX2 sampling. We assess possible mechanisms for severe wind production in this environment using simulations in section 3.

3) OBSERVATIONAL SUMMARY

During the evening hours of 6 May 2010 in northwestern Kansas, a supercell formed north of a warm front in a strongly sheared environment without surface-based CAPE. Observed MUCAPE \( \geq 500\text{Jkg}^{-1} \) aloft apparently supported the elevated supercell updraft throughout its life. Mobile mesonet observations of surface \( \theta_e \) suggest that midlevel downdrafts failed to penetrate the stable layer and reach the surface, supporting the notion that the storm was largely decoupled from the air mass below the frontal inversion. Many

\(^{2}\)The National Weather Service (NWS) criterion for severe thunderstorm winds is \( \geq 25.7\text{m} \text{s}^{-1} \).
Fig. 6. Reflectivity (dBZ) and radial velocity (m s\(^{-1}\)) from SMART-R2 at (a),(b) 11.4° at 0046:45 UTC; (c),(d) 10.0° at 0046:32 UTC; (e),(f) 7.8° at 0046:18 UTC; and (g),(h) 5.6° at 0045:53 UTC 7 May 2010. Approximate locations of the midlevel mesocyclone (red) and mesoanticyclone (blue) are marked with arrows. Height of mesocyclone is labeled in each velocity image. Labeled reflectivity values on the lhs color scale are -24, -11, 2, 15, 28, 41, and 54 dBZ. Labeled velocity values on the rhs color scale are -18, -12, -6, 0, 6, 12, and 18 m s\(^{-1}\).
quintessential supercell characteristics were observed, including a hook echo to the southwest of the weak-echo region and a mesocyclone in the midlevels. Additionally, an unusual anticyclonically curved hook echo was present and may be related to the mesoanticyclone in the storm’s rear flank. Given the limitations of the VORTEX2 datasets (as the first case in 2010, there were a number of unexpected radar quality issues), it is difficult to delve much farther into the storm’s governing dynamics. Therefore, further study of this supercell was...
undertaken using idealized model simulations, the results of which are discussed in section 3.

3. Numerical model simulations

a. Model setup

Although unprecedented, the observations for this case are still limited in duration and scope. Therefore, an idealized model simulation was designed to advance our understanding. Simulations in this study were run using Cloud Model 1, version 16 (CM1; Bryan and Fritsch 2002), with the new, higher-order parcel interpolator from version 17. The model domain for each simulation was 165 km × 165 km × 16 km with a horizontal grid spacing of 250 m. The grid moved at a constant speed throughout the simulation to keep the supercell centered. A stretched vertical grid (64 levels), with spacing ranging from 100 m in the lowest 3 km to 500 m above 9 km, was used to better resolve the stable layer near the surface. A damping layer was put in place above the tropopause (11.5 km to the model top) and along each of the lateral boundaries (20-km width) to prevent reflection off the model top and sides, respectively. Clouds and precipitation were represented using the NASA Goddard version of the Lin et al. (1983) microphysics scheme (Tao et al. 1989; Tao and Simpson 1993). Each simulation was run out for 4 h with a large time step of 2.0 s and output every minute over the analysis times. The Coriolis force, surface fluxes (including drag), and radiational effects were neglected.

A horizontally homogenous environment was implemented at model start using a sounding representative of the observed environment (Fig. 4), created by taking a straight average, with some caveats, of two inflow soundings (0039 and 0117 UTC; green and red thermodynamic profiles in Fig. 4, respectively) combined with the wind profile from another inflow sounding (0106 UTC; pink hodograph in Fig. 4). The thermodynamic profiles were based on the 0039 and 0117 UTC soundings because they were taken closer to the storm and farther into the postfrontal air mass; the 0106 UTC sounding appeared to be unrepresentatively dry aloft (not shown). Above 4 km AGL, the 0117 UTC sounding was ignored as a result of its trajectory into the forward flank of the supercell, leading to a profile that was near saturation throughout and not indicative of the true inflow environment. The moisture profile for the blended sounding was not allowed to exceed a relative humidity of 90% at any height to prevent the presence of moist absolutely unstable layers that would spark excess precipitation. The 0106 UTC sounding was employed for the model wind profile because of worries that the wind profiles in the 0039 and 0117 UTC soundings could be convectively contaminated farther aloft as those soundings drifted closer to the storm [see, e.g., the near-supercell wind perturbations analyzed by Parker (2014, his Fig. 7)]. This amalgamation of near-inflow thermal profiles with the far-field winds allowed for a realistic inflow environment that sustained convection in the model without convective contamination of the winds.

Through experimentation, a warm bubble was found to be insufficient to produce a sustained storm because of the interactions with the environmental stable layer. In nature, frontal lifting likely provided ascent aloft to support initiation. As a proxy, convergence was initialized over the first 30 min of the simulation following the technique of Loftus et al. (2008). However, given the elevated nature of the expected convection, the applied convergence was maximized near the top of the stable layer instead of at the surface, as Loftus et al. (2008) originally envisioned. The equations for velocity potential from Loftus et al. (2008) were used but with a maximum divergence (set to −5.0 × 10⁻³ s⁻¹), taking place at (xc, yc, zmax), where xc and yc are the horizontal points at the center of the domain and zmax is set to 2000 m. With convergence maximized near the top of the stable layer, the storm developed and survived long enough for analysis of its structure and maintenance. Although this ostensibly forces the initial storm to feed primarily on elevated air, the environment is not supportive of surface-based convection (zero surface-based CAPE; Fig. 4), so the elevated forcing for initiation simply expedites convective development in a realistic scenario that closely resembles the VORTEX2 observations.

To further understand the origins and fates of parcels within the elevated supercell, passive tracers and trajectories were used in the model. Tracers were initiated at the start of the model in 500-m-deep layers from the surface up to 2.5 km. These tracers are both passively advected and smoothed by the full scalar equations in the model; fractional concentrations (0–1) of their original value reveal the displacements of air in each of these layers over time. In addition, massless parcels were introduced 2 h into the simulation to give a more explicit description of specific parcel trajectories. A total of 2190584 parcels were released near the supercell over a 123 km × 145 km × 6 km grid. The parcel trajectories are integrated forward using the model

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3. These soundings, used individually, were not able to sustain convection in the model. As a result of the poor representation of the thermodynamic environment, a NAM-initialized WRF simulation was also unsuccessful in simulating this storm.
winds on the native model time step and are saved every 30 s.

b. Simulation results

As in the observations, the simulation produced a supercell storm that was quasi steady through 3 h. Because of this, for brevity, the present analysis will focus on the storm 2 h into the simulation (Fig. 8) unless otherwise noted. At this time, the storm is representative of other times in the simulations and parallels the observations. Beneath the stable inversion, there is a lack of significant vertical motion (Fig. 8, purple contours); however, appreciable wind perturbations are evident (Fig. 8; 0.5 km AGL). At 1 km AGL and above, strong downdrafts are evident in the rear flank, nearly collocated with the hook echo (dashed purple contours in Fig. 8); the dynamics and effects of this descent will be discussed later in this section.

In the low levels, the storm has a somewhat diffused appearance with reflectivity structure that is not strikingly supercellular. Although this may be perpetuated in the

Fig. 8. Plan views of reflectivity (shaded; dBZ), ground-relative wind vectors, and vertical velocity greater (less) than 5 (–5) m s$^{-1}$ shown in solid (dashed) purple contours 120 min into the simulation at (from top left to bottom right) 3.0, 2.5, 2.0, 1.5, 1.0, and 0.5 km AGL. The potential temperature deficit at the surface is shown with the 0.5 km AGL reflectivity (contoured in dashed blue at –1 and –3 K).
model by diffusion, the observed storm had a similar look on radar at lower tilts (at altitudes below what is shown in Figs. 5 and 6). This may be a partial effect of the decoupling of winds from the storm beneath the inversion, such that precipitation falling through the subinversion layer is redistributed by a flow field that is not characteristic of the storm aloft. At and above the stable layer, the main updraft appears to the east of a prominent anticyclonic hook echo (Fig. 8), bearing great resemblance to the observations (e.g., Fig. 6). The winds have a clear cyclonic (anticyclonic) curvature to them to the east (west) of the hook echo, especially at and above 2.0 km (Fig. 8), indicating substantial cyclonic (anticyclonic) vorticity that is present in the updraft (downdraft). By 3.0 km AGL, the appearance of reflectivity, along with cyclonic vorticity in the updraft, conforms to the typical supercell model (Fig. 8).

A key motivation for this study was to understand the origin of the updraft air in presumably elevated supercells. At 5 km, the highest concentration of passive tracer in the updraft is from above 1.5 km, with no tracer arriving from the lowest 1 km (Fig. 9). As noted previously, no CAPE exists in the lowest 1 km, so this result makes physical sense. Some lower concentrations from the 1–1.5-km layer appear west of the main updraft, suggesting that low-level parcels, possibly with rather small instability (CAPE < 500 J kg⁻¹), are still being lifted to

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**Fig. 9.** Plan views of tracer concentration (shaded) and vertical velocity every 10 m s⁻¹ (purple) 120 min into the simulation at 5 km AGL. Tracer originates in the 0.05–0.5-, 0.5–1.0-, 1.0–1.5-, 1.5–2.0-, and 2.0–2.5-km layers with a maximum concentration of 1. The 35-dBZ contour is shown in black for reference. The dashed gray lines in the top-left frame correspond to the locations of the updraft cross sections in Fig. 17.
an appreciable altitude (10.6% of analyzed updraft parcels came from the 1–1.5-km layer). In many cases, this air (with substantial CIN) is strongly negatively buoyant during ascent and is accelerated back to its original height or lower [i.e., up–down trajectories, as described in section 1; cf. Knupp (1987)]. This regime of trajectories will be discussed further momentarily. To further examine the origins of updraft air, parcels having acquired $w > 10 \text{ m s}^{-1}$ at 5 km in a 30 km $\times$ 40 km box around the updraft within 2–2.5 h into the model simulation were identified for analysis (6580 parcels in total). Placing the strict thresholds on $w$ and the parcel’s horizontal location prevented the inclusion of parcels that were not representative of the updraft. All of the parcels originate above 1 km in height, with the majority coming from above 1.5 km (Fig. 10a). This matches both the environmental CAPE profile (Fig. 4) and the lack of significant vertical motion below the inversion height.

To look more closely at typical low-level parcel behavior, the aforementioned 6580 updraft parcels were averaged within each 250-m-deep layer (0–250 m AGL, 250–500 m AGL, etc.) based on their heights of origin (Fig. 11a); note that the plotted potential temperature perturbations (shaded field in Fig. 11a) are time averaged and thus not necessarily representative of an individual parcel’s instantaneous buoyancy at each point. Below ~1.2 km, parcels follow an up–down flow branch, becoming strongly negatively buoyant after being lifted to an altitude as high as 4 km and then moving out of the zone of lifting (Fig. 11b). The zone where parcels have ascended into the warm inversion layer is clearly visible as a cold pocket centered at $x = 10 \text{ km}, z = 2 \text{ km}$ in Fig. 11. Above 1.2 km, each averaged trajectory rises into the main updraft.

To further isolate the behavior of the up–down trajectories that travel through the anticyclonic hook-echo region, we identified and averaged the 11 such parcels that arrived there with $z > 0.02 \text{ s}^{-1}$ during a 5-min period centered on $t = 2 \text{ h}$. As a result of the impressive directional and speed shear in the wind profile (Fig. 4), the ambient environmental parcels originate with appreciable streamwise horizontal vorticity (Fig. 12a). Streamwise horizontal vorticity will be reoriented into the vertical as a result of gradients in vertical velocity along the parcel trajectory, as shown in

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**Fig. 10.** Normalized (a) origins of updraft parcels in 100-m-deep layers and (b) destinations of downdraft parcels originating above 1 km AGL 120 min into the simulation (or 30 min after model restart). Updraft parcels were defined as those with $w > 10 \text{ m s}^{-1}$ at 5 km and above in a box centered around the storm. Downdraft parcels were defined as those with $w < -3 \text{ m s}^{-1}$ in the lowest 2 km AGL at any point during their life and were then sorted to only include those that originated above 1 km AGL.
Fig. 12b. In general, supercell mesocyclones are fueled by tilting and subsequent stretching of this horizontal vorticity, yielding large values of positive $\zeta$ during parcels' period of rapid ascent (Davies-Jones 1984). Similarly, tilting of ambient horizontal vorticity occurs along the up–down trajectories, creating a small amount of positive $\zeta$ initially (Fig. 13). Because of their negative buoyancy, however, stretching is weak for these up–down parcels, limiting the amount of positive $\zeta$ produced (Fig. 13). Therefore, updraft parcels (as opposed to the up–down trajectories) dominate the center of positive $\zeta$ in the parent storm (i.e., the mesocyclone).

Negative buoyancy drives the up–down parcels quickly toward the surface, whereupon tilting causes the vorticity vectors to be redirected downward (Figs. 12b and 13).
Substantial “negative” stretching of the air over time results from this downward acceleration, producing large values of anticyclonic $\zeta$ in the descending branch of these trajectories (Fig. 13). In addition to the production of vertical vorticity by tilting and stretching, baroclinic generation of horizontal vorticity is also present (e.g., Rotunno and Klemp 1985); the parcel budgets for horizontal vorticity do not reconcile very well, but the sign of

FIG. 12. (a) Vertical profile of the ambient streamwise vorticity (s$^{-1}$) from the model simulation with the origin layer for averaged parcels highlighted in the dashed green box, and (b) parcel trajectory (black) and vorticity vectors (teal) viewed from the south-southeast for an averaged up–down trajectory that descends behind the hook echo (pink trajectory in Fig. 14).
the baroclinicity on average produces antistreamwise horizontal vorticity (i.e., it decreases the initially large values of streamwise vorticity during the trajectory history; not shown). The history of vorticity vectors along the averaged trajectory (Fig. 12b) is consistent with the budget, showing that the primary process contributing to strong anticyclonic vorticity in the hook-echo region is the reorientation and subsequent stretching of initial ambient streamwise vorticity along the up–down trajectories that flow into that part of the storm.

Interestingly, this persistent negative $\zeta$ along the up–down trajectories, acquired upon descent into the rear-flank, results in pools of negative vorticity to the west of the hook echo (Fig. 14a). Trajectory analysis confirmed that this was the primary method of negative vorticity generation as parcels originating in the mid-levels were not present near the vorticity minimum (not shown). At 3 km AGL, this anticyclonic vorticity (approaching $-0.05 \text{ s}^{-1}$) is much greater in magnitude than the positive vorticity present in the updraft sector (0.01–0.015 \text{ s}^{-1} centered at $x = 15 \text{ km}$; Fig. 14). Descending precipitation then wraps around the stronger, anticyclonic zone of vorticity, resulting in the peculiar anticyclonic shape of the simulated hook echo that was also seen in the observed storm (Fig. 14a).

Although we have demonstrated the origins and behavior of updraft and up–down air in the low levels, the persistent lifting mechanism leading to the steadiness of these flow branches is still unclear. The buoyancy field and dynamic and buoyant perturbation pressures (Rotunno and Klemp 1982) were calculated using the solver described by Coffer and Parker (2015, 2017) and Davenport and Parker (2015). A west–east cross section through the main updraft shows a strong dynamic low to the east of the updraft (Fig. 15a; $x = 18 \text{ km}, z = 4.5 \text{ km}$). This is fueling upward dynamic accelerations from near the inversion to roughly 6 km (Fig. 15, red hatching). Combined with negligible upward buoyant accelerations below 3 km, it is implied that the supercell’s own dynamic lifting plays a major role in both flow fields, lifting air below the inversion into the ascending branch of the up–down trajectories, and lifting parcels above the inversion to their LFCs.

In addition to the dynamic lifting, an area of negative buoyancy is present at the inversion top (Fig. 15a; $x = 12 \text{ km}, z = 2 \text{ km}$), which may provide additional isentropic upglide where the potential temperature surfaces are domed upward, much as with the elevated cool pool noted by Bryan and Weisman (2006). This negative buoyancy anomaly is associated with the zone of parcels from the stable layer that have ascended in the up–down flow branch. At the same latitude and longitude of this buoyancy minimum, ridges and troughs are present in the potential temperature surfaces with vertical velocity perturbations located roughly one-quarter wavelength out of phase with these ripples (Fig. 16), which is indicative of gravity waves. These features are persistent and continually adjacent to the simulated storm. These waves represent perturbations produced by the storm that can potentially contribute to the maintenance of the updraft by causing the upglide of parcels to their LFC.

In the observations (section 2, Fig. 7) we noted a weak surface cold pool exhibiting a lack of a significant $\theta_e$ deficit. The model data reveal a similar phenomenon. Over the course of 120–180 min in the simulation, the $\theta_e$ deficit in the surface cold pool never exceeds 2 K, while $\theta$ deficits are typically 2–3 K (Fig. 17). Much as in the observations, the base-state vertical profile of $\theta_e$ in Fig. 17 suggests that this cold pool air could have only descended from below roughly 1 km AGL. Exploration of the downdrafts by way of parcels reveals that downdraft trajectories ($\omega \approx -3 \text{ m s}^{-1}$) originating above 2 km AGL indeed do not reach the surface (Fig. 10b), not even in the area of $\theta$ surplus in the forward flank ($x = 15–20 \text{ km}, y = -9$ to $-14 \text{ km}$ in Fig. 17), where a weak heat burst is present.

We noted the occurrence of near-severe surface winds in advance of the observed cells late in the life cycle of the 6 May 2010 case (section 2). Interestingly, the simulated supercell produces severe surface winds out ahead of the surface cold pool (Fig. 18). Two parcels linked to this phenomenon were singled out for further investigation. Each parcel follows an up–down
flow branch, originating in the lowest 100 m AGL and ascending to greater than 400 m AGL. These parcels both descend in the hook echo, suggesting that evaporation may have enhanced their negative buoyancy. Upon descending to the ground, a sharp increase in ground-relative wind speed is experienced to above severe limits, reaching values as high as 40 m s$^{-1}$ (Fig. 18). For both the gray and black parcels in Fig. 18, the air is accelerated while traveling close to the ground because of the dynamic high that formed where the downdraft meets the surface ($y = -9$ km), implying continued horizontal pressure gradient forcing. The pressure gradient between a dynamic surface high in the downward branch of the up–down trajectory and a dynamic surface low, due to a rotorlike circulation, is the cause of this acceleration to the east (Fig. 18b). The

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These parcels actually descend below the lowest model level ($z = 50$ m).
FIG. 15. Vertical cross sections through the (a) short and (b) long axes of the supercell updraft of buoyancy with storm-relative wind vectors, buoyant pressure perturbations with buoyant acceleration vectors, and dynamic pressure perturbations with dynamic acceleration vectors 120 min into the simulation. Here, $w > 5 \text{ m/s}$ is shaded in gray and vertical accelerations $\geq 0.05 \text{ m/s}^2$ are hatched in red. Buoyancy is contoured every 0.10 m s$^{-2}$ and the pressure perturbations are contoured every 50 Pa with negative values dashed.

W > 5 m/s shaded on all plots

ACCB, ACCD in m/s$^2$
Vertical component > 0.05 m/s$^2$ hatched in red

BUOY contoured every 0.10 m/s$^2$
P$^B$, P$^D$ contoured every 50 Pa (negative values dashed)

$t = 120 \text{ min, } y = -8 \text{ km}$

W > 5 m/s shaded on all plots

ACCB, ACCD in m/s$^2$
Vertical component > 0.05 m/s$^2$ hatched in red

BUOY contoured every 0.10 m/s$^2$
P$^B$, P$^D$ contoured every 50 Pa (negative values dashed)

$t = 120 \text{ min, } x = 17 \text{ km}$
footprint and magnitude of the surface winds seem to correspond well with the observed case.

c. Simulation summary

A numerical model simulation was run using a blend of soundings from the 6 May 2010 VORTEX2 case to further investigate the mechanisms behind the structure, maintenance, and severe surface wind production of the storm. Updraft air was found to originate from above 1 km AGL and was mostly driven upward by dynamic lifting from the supercell itself. Parcels originating below ~1.2 km AGL were initially lifted dynamically before becoming negatively buoyant, causing the air to descend; these are known as up–down parcels. This flow branch was the main low-level downdraft mechanism. Downdrafts originating in the midlevels were not able to penetrate through the stable layer. It therefore appears that the surface cold pool was fueled by the evaporation of precipitation beneath the inversion, which is supported by both the ambient surface $\theta_e$ values and the lack of midlevel downdraft parcels below 1 km in the simulation. Meanwhile, up–down trajectories that impinged on the warm, inversion layer assisted in the development of an elevated cold pool atop the inversion. In the descending branch of the up–down parcels, pools of negative vertical vorticity due to vortex tilting and stretching contorted the tip of the hook echo into an anticyclonic bend. Up–down parcels from the lowest 100 m AGL were also the driver behind simulated severe surface winds, a result of strong low-level downdrafts reaching the surface and an acceleration of the winds near the ground by a horizontal pressure gradient.

4. Conclusions and future work

The 6 May 2010 storm observed during VORTEX2 is perhaps the first coordinated dataset ever collected for an elevated supercell. The supercell formed to the north of a warm front in an environment characterized by strong vertical wind shear and zero surface-based CAPE. Despite the lack of surface instability, MUCAPE approaching 1000 J kg$^{-1}$ sustained the supercell throughout its life cycle. The storm was steady over the course of the radar sampling, moving to the east-northeast without much
change in structure. An anticyclonic hook echo was prominent with this supercell and was collocated with anticyclonic rotation in the midlevels (Fig. 6). Cyclonic midlevel rotation existed in the column above the weak-echo region in the single-Doppler analysis, implying the existence of a rotating updraft and justifying the classification of this storm as a supercell. Only a weak surface cold pool was evident, and it was apparently devoid of air originating from midlevels.

To explore some of the relevant dynamical processes that were difficult to observe or not directly observed, we simulated the storm using an idealized numerical model with a blend of near-storm soundings taken on 6 May 2010. The simulated supercell was similar in structure to the observed storm, having a less well-defined structure near the surface that became more supercellular-looking farther aloft (Fig. 8). The simulated storm also included an anticyclonic hook echo that was the result of a pool of negative vorticity in the rear-flank region. Only air parcels originating above the near-surface layer fueled the supercell updraft, with most of the inflow air coming from above ~1.2 km AGL. This air was lifted dynamically by the supercell itself; buoyant accelerations were negligible in the low levels. Below ~1.2 km AGL, the majority of air followed an up–down trajectory, being lifted dynamically (to <4 km AGL on average) until it was strongly negatively buoyant and then quickly descending to the surface. This flow branch was the main low-level downdraft mechanism in the storm, since parcels originating in the midlevels were unable to penetrate the stable layer. The surface cold pool was consequently fueled by the evaporation of precipitation below the stable layer. Up–down parcels, halted in their ascent by the warm inversion, created an elevated cold pool on top of the inversion layer. Air descending in these up–down trajectories created pools of negative vertical vorticity that bent the hook echo in a counterclockwise manner. In the lowest 100 m, the up–down parcels were also the driver behind the simulated surface winds that reached severe wind speed thresholds, descending to the surface rapidly and being accelerated by a dynamic perturbation pressure gradient near the ground. This is similar to the mechanism discussed by Schmidt and Cotton (1989) and Bernardet and Cotton (1998), both of whom found that strong surface winds in a mesoscale convective system stemmed from up–down parcels and were enhanced by a horizontal pressure gradient associated with a mesohigh–mesolow couplet near the ground.

Interestingly, our results differ somewhat from those of Nowotarski et al. (2011). The majority of the elevated
supercells in their study, simulated over statically stable boundary layers with various levels of stability, were still fed by near-surface parcels and had midlevel downdrafts that were able to reach the surface. Only in cases of extreme stability were their simulated supercells truly decoupled from the surface. Although the reasons for the differences between their findings and ours are not totally clear, we speculate that much of the difference hinges on our use of an observed sounding [as opposed to the rather moist Weisman and Klemp (1982) sounding in their study]. The present sounding (e.g., Fig. 4a) has zero surface-based CAPE, such that near-ground air simply cannot acquire positive buoyancy and fuel updrafts.

As with any case study, it is unknown how representative this storm is of all elevated supercells (nor how well the simulation replicates the processes of the real-world storm). More cases would obviously be ideal. It is also unknown whether frontally forced
evolved storms are fundamentally different from nocturnal elevated storms, nor how such storms might respond to spatial or temporal changes in inversion strength. Our simulation also suggests some important roles for up–down trajectories and possible gravity waves, but such phenomena are likely quite difficult to observe. In terms of societal impact, the seemingly unusual mechanics of severe surface winds in our simulated elevated supercell deserve further exploration. Are there specific environmental factors, outside of the stable inversion, that could make an elevated supercell environment more conducive to severe wind production? And, are storms such as this truly incapable of producing tornadoes? Although not discussed in detail here, our simulated elevated supercell occasionally produced short-lived near-surface vortices. We presume these to be spurious, but a more advanced understanding of tornadoes in supercells that appear to be elevated would be worth pursuing. The long-range goal of this line of research is to advance the knowledge of how storms respond to their environments across the full range of observed lower-tropospheric stability.

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