The “Owl Horn” Radar Signature in Developing Southern Plains Supercells

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ABSTRACT

During spring 2001 in the Southern Plains, a recurring, hitherto undocumented reflectivity signature that the authors have called the “Owl Horn” signature (because the radar reflectivity pattern resembles the profile of the Great Horned Owl) was observed on a mobile, X-band radar display. The reflectivity signature was always located at the rear side of a developing supercell, spanned the entire rear side of the storm, and was always seen on low-level plan position indicator (PPI) scans. It lasted on the order of only 5–10 min and was not an artifact of the radar.

A study of the Owl Horn signature was undertaken using the Tracking Radar Echoes by Correlation technique (TREC) to estimate the wind field. TREC has previously been applied to clear-air and hurricane environments, and to the internal motions of severe storms, but not to their evolution. The characteristics of the signature are presented, and then, through the application of TREC to the radar reflectivity data (Doppler wind data were not available in 2001) collected during May and June 2001, the horizontal wind field was estimated around and in the Owl Horn signature.

Instances of the Owl Horn in numerical model storm simulations were investigated. The numerical simulations were used to identify conditions under which the signature occurs, the process by which it is created is discussed, and its dependence upon the environmental wind shear is examined. Results indicate that the hodograph shape and magnitude influence the production of the Owl Horn signature. Supercell-magnitude shear is required, and some curvature—particularly low-level curvature—is essential to the production of the feature. The Owl Horn signature is formed when horizontal vorticity is tilted into the vertical by expanding outflow through a positive feedback mechanism with the outflow.

1. Introduction

The flow patterns in severe convective storms have been the subject of numerous research studies in recent years, especially with the advent of mobile Doppler radars (e.g., Wurman et al. 1997; Bluestein and Pazmany 2000; Biggerstaff and Guynes 2000; Pazmany et al. 2003). These radars enable the collection of higher spatial resolution datasets, because the radars can be brought closer to the storms. But despite the higher-resolution datasets, studies of supercells—and of tornadic storms in particular—have typically focused only on the mature stage of the storm, because tornadoes and low-level mesocyclones are not commonly expected to form until at least 30 min–1 h into a storm’s life (e.g., Burgess et al. 1982; Davies-Jones 1986). The result of this focus is that an examination of the earlier stages of the storm’s development is neglected. During
the spring 2001 storm season in the Southern Plains, a group of graduate students and faculty from the University of Oklahoma used a mobile, X-band (3-cm wavelength), non-Doppler radar (Pazmany et al. 2003) developed at the University of Massachusetts, Amherst (UMass) to survey convective storms at both their early and mature stages. On several occasions in the earlier stages of storm development, a curious, recurring signature in radar reflectivity was observed that we have called the "Owl Horn" signature (OHS), owing to its resemblance to the profile of the Great Horned Owl. The OHS was apparent from various viewing angles with respect to the storm, thus demonstrating that it was not an artifact of the radar. The feature is defined by two protrusions in a storm’s low-level reflectivity at the rear side (with respect to storm motion) of an isolated storm (Fig. 1). The OHS appears relatively early in a storm’s development, lasts on the order of 5–10 min, and slowly erodes as the storm evolves.

The UMass radar (Fig. 2) operated with an antenna half-power beamwidth of 1.25°, transmitted 1-μs pulses, and had a range resolution of 150 m. Since the radar lacked Doppler capabilities in 2001, a study of the OHS was undertaken using the Tracking Radar Echoes by Correlation (TREC) technique (Rinehart 1979). TREC was used to compute the low-level wind field in and around the Owl Horn. The OHS is not noted in the literature, and although TREC has previously been applied to clear-air and hurricane environments (Tuttle and Foote 1990; Tuttle and Gall 1999), also absent from the literature is an application of TREC to study the evolution of severe convective storms. Rinehart (1979), however, has studied internal storm motions by applying TREC to severe storms.

Using the Advanced Regional Prediction System (ARPS) model (Xue et al. 2000), the OHS in convective storms has been simulated. The main purposes of this paper are to describe the Owl Horn feature and its structure from both reflectivity data and numerical simulations, to identify the conditions under which it occurs, and ultimately to explain how it forms. The TREC methodology is described in section 2, observational data and analysis are presented in section 3, results and analysis of the control simulation are discussed in section 4, and a summary of sensitivity test simulation results is given in section 5.

2. TREC methodology

The TREC technique is a pattern-recognition procedure applied to radar-reflectivity data using a cross-correlation analysis. Radar-echo data are stored in two-dimensional arrays, and on each iteration, an array is compared to all other arrays of the same size for the subsequent time step to determine which array exhibits the highest correlation with the previous array. The final array with the highest correlation is considered to be the endpoint of a motion vector beginning at the
that characterize convective storms, for which horizontal advection is not dominant. Storm motions were subjectively estimated from Weather Surveillance Radar-1988 Doppler (WSR-88D) data, and were then used as an input parameter in the TREC algorithm to produce a storm-relative wind field.

3. Owl Horn observations and TREC results

Well-defined Owl Horn signatures occurred west of Liberal, Kansas, on 27 May 2001 (Fig. 3a); north of Raton Pass, New Mexico, on 28 May 2001 (Fig. 3b); near Turkey, Texas, on 29 May 2001 (Fig. 3c); and near Woodward, Oklahoma, on 5 June 2001 (Fig. 3d). The storms that exhibited the signature on 28 May, 29 May, and 5 June all developed into supercells that produced funnel clouds or tornadoes. The 27 May storm, however, was overtaken by a very strong outflow boundary from a mesoscale convective system to its north before it could develop further. Of particular interest is that on

![Image](http://journals.ametsoc.org/mwr/article-pdf/133/9/2608/4219436/mwr2992_1.pdf)
29 May, two storms simultaneously exhibited an OHS, and one of these storms later produced the signature a second time—suggesting that the local storm environment may play a role in the feature’s development, and that the process that creates it can be repeated in a storm’s life cycle.

Based on National Weather Service soundings (not shown) from the sites nearest to the observed storms, common to all the storm environments in these cases are supercell shear (>$19 \text{ m s}^{-1}$; Weisman and Klemp 1982) and moderate ($>1500 \text{ J kg}^{-1}$) convective available potential energy (CAPE). Significant turning of the winds with height in the lowest levels is generally present in the wind profiles, suggesting that enhanced low-level rotation can be achieved (Davies-Jones 1984; Weisman and Rotunno 2000).

The discussion herein focuses on the two best examples of the Owl Horn cases, on 27 May and 5 June 2001. The 28 and 29 May TREC analyses are not described here because the computed results were unreliable. In the former case, the OHS was at too great a distance from the radar (resulting in poor interpolation resolution) to yield accurate computations in TREC. The latter case is omitted because results suggest that the assumption of horizontally dominant advection fails owing to significantly large vertical motions in the storm—evidenced by in situ observations of baseball-to softball-size hailstones.

a. 27 May 2001

The OHS on 27 May was evident between 1643 and 1648 CDT in an isolated cell west of Liberal, Kansas. The storm was probed initially from its east-southeast, but the radar was eventually relocated to a position south-southeast of the storm (at which time the OHS appeared), and the radar remained stationary as the storm moved toward the southeast. During the period when the OHS was present, two distinct shafts of precipitation were apparent (Fig. 4), perhaps suggesting the development of a second updraft on the left side of the storm with respect to its motion. Although the feature was present for roughly 5 min, the best representation of the signature occurred from 1645:17 to 1645:35 CDT. The storm persisted well beyond the onset of the OHS, enduring on radar until roughly 1800 CDT, at which time it was overtaken by a very strong outflow boundary from the north (not shown).

Although the radar was operated at an elevation angle of $8.1^\circ$, this angle is not representative of the actual probing angle. Since the radar truck lacked both a leveling system and dependably flat ground from which to deploy, a bore-sighted camera was used to effect low-level plan position indicator (PPI) scans of the storm by allowing the radar operator to scan just above the tree line. It was found that the actual elevation angle of the data ranged from about $1.5^\circ$–$2.5^\circ$, which corresponds to a sampling elevation of about 500 m at a range of 15 km.

TREC was configured using an array of data points covering a $3 \text{ km} \times 3 \text{ km}$ area and an interpolation resolution of 100 m. Grid points were spaced at 500 m. Storm motion from the time of the best representation of the OHS was subjectively estimated at $11.3 \text{ m s}^{-1}$ toward the southeast (i.e., from $300^\circ$), which is reasonable in light of the northwest flow aloft (not shown). The TREC analyses from 1644:04 to 1644:22 and 1645:17 to 1645:35 CDT (Fig. 5a, top and center panels) are considered first. Flow in the storm was generally rearward with respect to storm motion. The feature of primary interest in the storm-relative wind fields was a wind shift line located at the rear side of the storm with respect to its motion. The line bore 5–10-min temporal continuity, persisting through the TREC analyses, and its ends were generally located near (or collocated with) the Owl Horn appendages in reflectivity.

By 1647:44 CDT (Fig. 5a, bottom panel), the right-flank (with respect to storm motion) Owl Horn persisted while the left-flank signature had weakened significantly in reflectivity. At this time, a secondary maximum in reflectivity (not shown) was observed along the left side of the storm (associated with the second precipitation shaft). This secondary maximum increased progressively in value and began to separate from the storm’s main region of precipitation.

b. 5 June 2001

The 5 June OHS was observed between 1711 and 1717 CDT in a cell near Woodward, Oklahoma, and the
best representation of the signature occurred from 1713:47 to 1715:19 CDT. The organized storm that exhibited the signature developed from clusters of disorganized convective cells. The radar probed the storm initially from the east and remained stationary as the storm moved toward the northeast. The storm was isolated; other significant storms, however, were present in the region. The left-flank feature itself exhibited a relative secondary maximum in reflectivity. This case differs from...
In all four cases, the storms that exhibited the OHS split soon after the signature was observed. No conclusions based on this fact alone can be made about an association between the splitting process and the development of the OHS. However, regions of enhanced reflectivity with distinct maxima developed along the left edges of all four storms as the OHS appeared, and these secondary regions did evolve into discrete left-splitting cells. This topic will be revisited from the perspective of numerical simulations in section 4. Some unanswered questions remain: the TREC-derived wind shift line behaved as a boundary in its evolution, but what does it represent? (It may be speculated that the line denotes an outflow boundary or a boundary between outflows.) Is the Owl Horn process confined to the lowest levels of a storm, or does it span the entire depth? Moreover, at what stage of maturity is a storm when the OHS is in progress? (Since the radar lacked Doppler capabilities, and volume scans were not collected, it was uncertain whether storms exhibiting the OHS had yet developed into supercells.) Numerical simulations were required to answer these questions.

4. Numerical simulations

The nonhydrostatic ARPS model with 1 km × 1 km × 500 m resolution (Xue et al. 1995, 2000) was used with a composite Del City, Oklahoma, sounding from 20 May 1977 (a smoothed version of that used by Klemp et al. 1981) to initialize a control simulation over a 64 km × 64 km × 16 km domain. Ice microphysics (Lin et al. 1983) was used to better mirror atmospheric processes. The thermodynamic and dynamic properties of the sounding are sufficient for supercellular development (e.g., Adlerman et al. 1999): nearly 2700 J kg⁻¹ of CAPE, and 36 m s⁻¹ shear over the 0–6-km layer. Simulation times herein are such that 2100 UTC corresponds to the beginning of the model run (i.e., at 0000 s), and results were displayed each minute from 2125 (1500 s) to 2200 UTC (3600 s). Observational evidence suggests that the OHS lasts on the order of several minutes. Therefore, an examination of its structure at each minute helps to discern the larger-scale storm features responsible for its creation. The simulation results are shown as horizontal plots of reflectivity, perturbation potential temperature, vertical motion w, vertical vorticity, and horizontal storm-relative wind vectors, as well as vertical cross-section plots of these quantities in and around the OHS.

Simulations run with warm rain microphysics also produced the OHS, but the feature was less pronounced.
a. Discussion of simulation results

Radar reflectivity was calculated using the formulation of Ferrier et al. (1995). At 2134 UTC (2040 s), prior to the onset of the OHS, the reflectivity field at 250 m AGL (Fig. 6) is characterized by an elliptical shape exhibiting concave curvature on its southeast side, and an orientation along a southwest–northeast axis with a single 59-dBZ reflectivity core. In a plot of perturbation potential temperature at 250 m AGL (Fig. 7) for the same time, two distinct surface cold pools are present. The coldest temperatures are located in the right-side cold pool (with respect to a storm motion from 192°), reaching as low as 5.1 K below ambient temperature. As time passes, the coldest air remains confined to narrow, elongated protruding bands, instead of spreading as a single mass. Moreover, the shape of the contours within the northern portion of these elongated bands

Fig. 6. Four-panel sequence of reflectivity plots at 250 m AGL for the control Del City, OK, simulation. Times shown are at (top left) 2134 UTC (2040 s), (top right) 2138 UTC (2280 s), (bottom left) 2140 UTC (2400 s), and (bottom right) 2147 UTC (2820 s). The OHS is denoted by the arrows at 2140 UTC. Distance in km. Color scale denotes reflectivity factor in dBZ.
bears a striking resemblance to the shape of the TREC-derived wind shift lines, supporting the earlier speculation that the TREC-derived lines may represent outflow boundaries.

The upward vertical velocities at 250 m AGL (Fig. 8) are also primarily confined to elongated bands flanking the outer edges of the cold-air protrusions in perturbation potential temperature. On the inner side of the cold-air protrusions, weak descending motion is found. The vertical motion bands along the outer edges of the cold pools resemble those produced in coarser supercell simulations (e.g., Klemp and Wilhelmson 1978a,b; Weisman and Klemp 1982) and are likely a consequence of air being lifted over the cold outflow as it

Fig. 7. Four-panel sequence of 250 m AGL perturbation potential temperature plots for the control Del City simulation. Times are as given in Fig. 6. Distance in km. Perturbation potential temperature is in K.
expands outward. However, a contribution to the upward motion from pressure perturbation gradients induced by winds that turn with height could also be expected.

Horizontal storm-relative winds (Fig. 9, vectors) flow from front to back with respect to storm motion. The strongest winds are focused in the elongated bands of outflow. A surprising result is shown in plots of vertical vorticity $\zeta$ (Fig. 9, color contours): an elongated pair of couplet bands of counter-rotating vortices is present at the back side of the storm with respect to storm motion, also flanking the cold-air protrusions. Cross sections through the Owl Horn region of the storm (Fig. 10, top and center panels) show that the cold-air protrusions are roughly 1.0–1.5 km deep, and that the right-flank protrusion is generally deeper than its left-flank counterpart. Moreover, the couplets of vertical vorticity are also confined to the lowest 1.0–1.5 km.
The OHS first appears at 2138 UTC (2280 s) in the reflectivity field at 250 m AGL (Fig. 6), and the appendages are collocated with the coldest air contained in the cold-air protrusions (Figs. 6 and 7). The feature lasts for 5 min, and as it dissipates, the storm splits and a new cell moves off to the left of the parent storm. Farther aloft in the storm \((Z = 5 \text{ km}, \text{not shown})\) a midlevel couplet in vertical vorticity, typical of supercell structure, is present (Bluestein 1993). The presence of a rotating updraft suggests that the storm has become a supercell by the time the OHS appears. Just prior to and during the appearance of the signature in the low levels, a strengthening second midlevel cyclonic/anticyclonic vorticity couplet develops. This fea-

\textbf{Fig. 9.} Four-panel sequence of 250 m AGL vertical vorticity (color contours) and horizontal storm-relative winds (vectors) for the control Del City simulation. Times are as given in Fig. 6. Distance in km. Winds are in m s\(^{-1}\). Vertical vorticity is in \(10^{-5}\text{ s}^{-1}\).
Fig. 10. Cross sections along lines depicted in reflectivity image at lower right: (top) vertical vorticity (Y = 19.5 km, line A); (middle) perturbation potential temperature (Y = 22.5 km, line B); and (bottom) reflectivity (Y = 32.5 km, line C) in the x-z plane at 2140 UTC (2400 s). Vertical vorticity is in $10^{-5}$ s$^{-1}$, perturbation potential temperature is in K, and reflectivity is in dBZ.
ture is commonly associated with a splitting storm, where a second couplet develops as a storm splits owing to a downward-directed midlevel pressure gradient (Bluestein and Sohl 1979; Klemp 1987; Bluestein 1993). Thus, evidence of a storm split comes in the mid- to upper levels just prior to and during the appearance of the OHS at the surface. Also in agreement with storm observations discussed in section 3, a vertical cross-section plot of reflectivity (Fig. 10, lower-left panel) through $Y = 32.5$ km at 2140 UTC (2400 s) shows two distinct shafts of precipitation reaching the surface. The second, younger, and less vertically extensive shaft eventually becomes the discrete second cell that splits off the simulated parent storm soon after the OHS develops.

It has been shown in simulations that the OHS in reflectivity is collocated with protrusions of colder outflow air. The colder air is channeled into narrow protrusions and advected rearward (along with radar targets) from the storm by flow amplified by the vertical vorticity couplets that flank the protrusions. It is apparent, then, that explaining the formation of the OHS is less about identifying the process that brings about the reflectivity signature, and more about accounting for not only the appearance of the vorticity couplets themselves but also the elongation of the bands in the perturbation potential temperature field. To this end, trajectory calculations were made using the ARPS results as input to existing code (Adlerman et al. 1999). The code uses a three-step predictor/corrector method and a trilinear interpolation scheme. Because the OHS is confined to the low levels of the storm (in particular, it seems strongest in reflectivity plots at 250 m), trajectory plots were prepared at 750, 500, and 250 m, for points composing an ensemble of boxes in and around the OHS.

**b. Discussion of trajectory and vorticity analyses**

Consider the vorticity equations in seminatural coordinates:

\[
\frac{D\omega}{Dt} = \omega \frac{D\psi}{Dt} + \omega \cdot \nabla V_H + \frac{\partial B}{\partial n}, \tag{1}
\]

\[
\frac{D\omega_n}{Dt} = -\omega \frac{D\psi}{Dt} + \omega \cdot \nabla (V_H \nabla \psi) - \frac{\partial B}{\partial s}, \tag{2}
\]

\[
\frac{D\zeta}{Dt} = \omega_n \cdot \nabla w + \xi \frac{\partial w}{\partial z}, \tag{3}
\]

where \((s, n, k)\) represent orthonormal basis vectors (Lilly 1982), \(D/Dt\) is the derivative following parcel motion, \(\omega\) is the three-dimensional vorticity vector, vector wind \(\mathbf{V} = (V_n, 0, w), \xi\) is the vertical vorticity, \(B\) represents buoyancy, and \(\Psi = \tan^{-1}(\psi/u)\) is the parcel streamline. These equations represent the changes in total vorticity in the streamwise \(\psi\) [along the flow, Eq. (1)], crosswise \(\zeta\) [normal to and to the left of the flow, Eq. (2)] and vertical \(\omega\) [Eq. (3)] directions. On the right-hand side of the first two equations, the first term is interpreted physically as the exchange of streamwise and crosswise vorticity in the plane of motion. The second terms represent the stretching and tilting of streamwise and crosswise vorticity in the plane of motion, and the third terms represent the baroclinic generation of streamwise and crosswise vorticity. In the third equation, the first term on the right-hand side represents the tilting of horizontal vorticity into the vertical, and the second represents the amplification of vertical vorticity by convergence/stretching. Trajectories for ensembles of parcels (spaced at 500-m intervals) were computed, that passed through the regions of cyclonic and anticyclonic vertical vorticity associated with the cold-air protrusions.

It is apparent in backward trajectories along the right-flank cold-air protrusion (Fig. 11) that parcels originated well to the east of the storm, in an environment unaltered by the storm. Parcels in the region of cyclonic vertical vorticity were traced to nearly 350 m AGL (Fig. 13a). They advanced toward the storm, ascended slowly until they neared the cold-air protrusion, when they ascended more rapidly to 750 m. Parcels in the region of anticyclonic vertical vorticity were traced to nearly 200 m AGL (Fig. 13c), ascended to nearly 950 m, and flowed over the cold-air protrusion before descending on the inner side. It is noted in forward trajectories (Fig. 12) that the parcels progressed farther westward with time; parcels from the cyclonic vorticity side continued their vertical ascent across and over the cold-air protrusions and subsequently descended on the anticyclonic vortex side of the protrusion (Fig. 13c). Therefore, it is clear from the trajectory plots that environmental air flowed up and over the right-flank cold-air protrusion.

Parcels along the left-flank cold-air protrusion (Fig. 11) behaved in the same manner, initially exhibiting westward motion. Parcels in the region of anticyclonic vertical vorticity were traced to nearly 100 m AGL (Fig. 13b). They advanced westward (away from the storm), ascended slowly as the cold-air protrusion spread westward, and were lifted more rapidly to 750 m AGL as the outflow overtook them. Parcels in the region of cyclonic vertical vorticity (Fig. 13d) were also traced to nearly 100 m AGL. They too initially moved westward and flowed over the cold-air protrusions as the expand-
ing outflow overtook them. It is seen in forward trajectories (Fig. 12) that the parcels continued their westward motion through the end of the simulation. From the trajectories it is thus verified that environmental air was lifted over the cold air as it expanded outward.

Consider a vorticity analysis for the backward trajectory of a parcel terminating at the point (36 km, 21 km, 750 m) at 2140 UTC (2400 s) in the simulation (denoted by a star along the right flank in Fig. 11a). This parcel was located in the region of cyclonic vertical vorticity associated with the right-flank Owl Horn. Time plots of the vorticity components (Fig. 14a) show that the parcel began with almost no vertical vorticity, and with almost all its vorticity in the streamwise direction. As the parcel moved westward toward the storm, its vertical vorticity slowly increased through tilting (Fig. 14c) as it

![Image](http://journals.ametsoc.org/mwr/article-pdf/133/9/2608/4219436/mwr2992_1.pdf)
neared the gradient of vertical velocity that flanks the cold-air protrusion. When the parcel approached the cold-air protrusion, the ascending motion ahead of the expanding outflow served to amplify the tilted vorticity through stretching. Then as the parcel neared the peak of its trajectory, tilting began to lessen. Since crosswise horizontal vorticity did not cross gradients of vertical motion here and would only have served to weaken the streamwise vorticity if tilted into the streamwise direction, it may be concluded that the cyclonic vortex associated with the right-flank Owl Horn was the result of tilting into the vertical of horizontal vorticity in the streamwise direction.

The origin of the streamwise vorticity must still be identified. Consider first the components of the streamwise and crosswise vorticity equations (Figs. 14b,d). In all vorticity calculations, the exchange term was neglected since it is relevant primarily in highly curved flows, and most parcel trajectories here (those not in the immediate vicinity of the updraft/downdraft interface) did not exhibit significant curvature. From a plot of the streamwise terms (Fig. 14b) it is apparent that

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**Fig. 12.** Same as Fig. 11, but for forward trajectories: (top) 2140 UTC (2400 s) and (bottom) 2200 UTC (3600 s).
neither tilting nor baroclinic generation initially effected changes in the streamwise horizontal vorticity, and stretching only amplified pre-existing streamwise vorticity. As the parcel neared the outflow, the contribution in both the streamwise and crosswise directions to the baroclinically generated vorticity increased (Figs. 14b,d). However, as noted earlier, the crosswise component of the vorticity could not be tilted vertically, so it is the streamwise vorticity that is of primary relevance along the right flank. But, since the model was initial-

**Fig. 13.** Plots of parcel elevation (km) vs time (s) for (a) parcel (36 km, 21 km, 750 m) in the region of cyclonic vorticity associated with the right-flank Owl Horn; (b) parcel (23 km, 24 km, 750 m) in the region of cyclonic vorticity associated with the left-flank Owl Horn; (c) parcel (32.5 km, 20.5 km, 750 m) in the region of anticyclonic vorticity associated with the right-flank Owl Horn; (d) parcel (21 km, 26 km, 750 m) in the region of anticyclonic vorticity associated with the left-flank Owl Horn.
ized with no pre-existing horizontal vorticity beyond that resulting from environmental vertical wind shear, and streamwise stretching, tilting, and baroclinic generation contributed very little to the total changes in streamwise vorticity, the streamwise (horizontal) vorticity initially experienced by parcels east of the storm could only have come from the environmental vertical wind shear.

A parcel at (23 km, 24 km, 750 m) on the cyclonic side of the left-flank appendage (denoted by a star in Fig. 14. For parcel (36 km, 21 km, 750 m) located in the region of cyclonic vorticity associated with the right-flank Owl Horn at 2140 UTC (2400 s), plots of (a) the parcel's total vorticity and components of total vorticity (10^{-3} s^{-1}), (b) terms in the streamwise vorticity tendency equation (10^{-3} s^{-2}), (c) terms in the vertical vorticity tendency equation (10^{-3} s^{-2}), (d) terms in the crosswise vorticity tendency equation (10^{-3} s^{-2}).
along the left-flank in Fig. 11a) originated in a region with positive streamwise vorticity and almost no crosswise vorticity, as it moved slowly westward. Owing to its origin within a weak gradient of vertical motion opposing the horizontal vorticity vector, the parcel initially had a negative component of vertical vorticity from tilting (Figs. 15a,c). The streamwise vorticity began to decrease (Fig. 15a), rapidly becoming negative.
while the crosswise vorticity briefly became positive. This change in sign is likely a reflection of the parcel’s change of direction of motion as it was overtaken by the outflow boundary and lifted, before restoring its original sense of motion. When the parcel reached its peak at about 2250 s, the vertical vorticity changed sign as the vertical tilting term contributed to increasing positive vertical vorticity (or less negative vertical vorticity). Ultimately, the streamwise vorticity increased again as the parcel leveled out at its peak. Given that the parcel’s motion is not completely perpendicular to the left-flank outflow boundary (especially when the parcel is overtaken by the boundary), it is plausible that a component of the crosswise vorticity along the left-flank could also have been tilted vertically.

The parcel at (32.5 km, 20.5 km, 750 m) on the inner (anticyclonic) side of the right-flank cold-air protrusion (denoted by a star along the right-flank appendage in Fig. 11b) originated in a region with almost exclusively positive streamwise vorticity. It is suggested in plots of height, vorticity, and vorticity changes for this parcel (Figs. 13c, 16a–d) that the parcel undergoes a process akin to that experienced by the left-flank cyclonic parcel examined above, which was lifted over the cold-air protrusion and descended on the inner side.

The parcel at (21 km, 26 km, 750 m) on the outer (anticyclonic) side of the left-flank cold-air protrusion (denoted by a star along the left-flank in Fig. 11b) appears to have undergone the same vorticity generation processes (Figs. 17a–d) as its outer right-flank counterpart. It must be noted, though, that a significant portion of the tilted (vertical) vorticity here also appears to have come from the crosswise component of horizontal vorticity, since the streamwise and crosswise vorticity both decrease simultaneously (Fig. 17a). Based on the comparable magnitudes of the streamwise and crosswise vorticity after these concurrent decreases, it seems likely that the rapid changes in sign can be attributed to temporary changes in the parcel’s motion (and consequent reorientation of the vorticity vector) while the parcel is overtaken by and lifted over the left-flank boundary.

It is seen in trajectories of parcels contained in the cold-air protrusions (not shown) that parcels on both the right and left flanks originated in the primary downdraft, which is somewhat unexpected. Only outflow parcels very close to the precipitation shaft of the developing left-splitting cell originated in the left-splitting storm’s downdraft. Perhaps the relative proximity of the right-flank Owl Horn to the primary (and stronger) downdraft/precipitation shaft explains why the right-flank appendage is generally observed to be the stronger/longer lived of the two, particularly in environments that favor right-moving cells.

5. Sensitivity test simulations

To test the sensitivity of the OHS to the wind profile in which a storm is initiated, several more simulations were conducted, using the same thermodynamic profile (CAPE of nearly 2700 J kg⁻¹) with a suite of supercell and nonsupercell hodographs previously used by Adlerman and Droegemeier (2002). Simulations with varied values of CAPE were not conducted, primarily because the Owl Horn process is believed to be a result of dynamic rather than thermodynamic processes. It is possible that varying the CAPE might alter the range of hodographs that support Owl Horn formation, but doing so is beyond the scope of this study. Hodographs included: a full 360° circle, a ¼ circle, a ½ circle over 0–10 km, a ½ circle over 0–6 km, a ¼ circle over 0–3 km becoming unidirectional above 3 km, and a straight line (intended to parallel the simulations of Weisman and Klemp 1982). Discussion of the full-circle hodograph will be omitted, since it is a rare occurrence. For the circular hodographs, the radius of the circle in each hodograph varied in 5 m s⁻¹ intervals from 10 to 35 m s⁻¹ over 0–10 km (and also 0–6 km in the case of the ½ circle hodograph), and tails varied from 0 to 60 m s⁻¹ over 3–9 km for the ¼ circle hodograph. Adlerman and Droegemeier (2002) showed (Fig. 18) that, after a rotation of 15°, the hodograph for the 20 May composite sounding can be very well represented by a semicircular hodograph of radius 19 m s⁻¹ over 0–10 km. Therefore, for control purposes, 19 m s⁻¹ was used throughout our extended simulations in place of 20 m s⁻¹.

A summary of the results of the sensitivity test simulations is given in Table 1. Where wind shear magnitude is concerned, supercell shear (Weisman and Klemp 1982) appears essential for the production of an OHS. However, within the spectrum of supercell shear, only weaker magnitudes of shear produced Owl Horn signatures in the simulations, whereas storms in environments with very strong shear did not—that is, the majority of the hodograph lengths fell generally into the range of 20–40 m s⁻¹ over 0–6 km, depending on the hodograph shape. As the shear magnitude was increased, the left-splitting storm and the left-flank Owl Horn became indistinguishable—in fact, for higher-shear cases, the left-splitting cell became the left-flank Owl Horn itself.

Hodograph curvature is also necessary, given that simulated storms in an environment with a straight-line
TABLE 1. Summary of results from numerical simulation sensitivity tests. For each hodograph shape used in the simulations, the shear magnitude(s) that produced a storm that exhibited the OHS is given. A positive case was defined as a storm that produces significant protrusions in both reflectivity and cold-air outflow at the rear side of the storm, given the apparent dependence on the cold-air protrusions for a true Owl Horn to develop. It is noted that too much shear seems detrimental to Owl Horn production: hodographs with circle radii of 30 m s$^{-1}$ and greater did not produce Owl Horns during the simulations.

<table>
<thead>
<tr>
<th>Hodograph shape</th>
<th>Circle radius/tail magnitude that produced OHS (in m s$^{-1}$)</th>
<th>Hodograph length(s) 0–6 km (in m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>½ circle 0–10 km</td>
<td>15, 19, 25</td>
<td>28, 35, 47</td>
</tr>
<tr>
<td>½ circle 0–6 km</td>
<td>*15, *19, *25</td>
<td>28, 35, 47</td>
</tr>
<tr>
<td>¾ circle 0–3 km,</td>
<td>10, tail 10 m s$^{-1}$</td>
<td>21</td>
</tr>
<tr>
<td>tail 3–9 km</td>
<td>15, tail 0, 10, and 20 m s$^{-1}$</td>
<td>24, 29, 34</td>
</tr>
<tr>
<td>¾ circle 0–10 km</td>
<td>10, 15, 19</td>
<td>28, 42, 54</td>
</tr>
<tr>
<td>Straightline</td>
<td>None</td>
<td></td>
</tr>
</tbody>
</table>

* Magnitudes represent a corresponding radius: the radii actually used in the ½ circle 0–6-km simulations are chosen to create the same mean shear (0–6-km hodograph lengths) as in the ½ circle 0–10-km simulations with the given radius.

hodograph did not produce an OHS for any magnitude of shear while storms in environments with a curved hodograph did produce the signature for the same mean shear. The principal difference between the curved hodograph cases and the straight-line hodograph cases was in the configuration of the vorticity couplets: banded couplets flanked the cold-air protrusions in the curved hodograph cases, but in the straight-line hodograph cases, significant cold-air protrusions did not appear, and only one vorticity couplet at the rear side of the storm was evident.

6. Summary and conclusions

A radar-reflectivity feature, detected in developing supercells by a mobile, X-band radar and called the “Owl Horn” signature (OHS), has been analyzed. Characterized by weak-echo protrusions in the reflectivity field along the rear side of these storms, the feature is confined to the lowest levels of the storm and lasts only 5–10 min, making its detection very difficult.

Simulations and analysis have provided insight into the development of the OHS. It was shown that the coldest low-level outflow often began in two discrete or nearly discrete pools at the outer edges of the surface cold pool associated with the convective storm and developed into elongated protruding bands. Banded low-level vertical vorticity couplets (not related to the couplets generated through tilting in the midlevels by the main updraft) were generated at the rear side of the storm, flanking the cold-air protrusions in the outflow as horizontal vorticity (environmental vorticity and vorticity generated baroclinically as parcels approached the outflow) was tilted into the vertical when ambient air flowed over the cold-air protrusions. As the cold air was channeled farther rearward into long, narrow protrusions, the OHS developed in the reflectivity pattern when precipitation was advected rearward by the enhanced winds from the vertical vorticity couplets. Ultimately, as the cold pool expanded, the reflectivity pattern dissipated primarily because the enhanced winds on the vorticity couplet interfaces no longer had a direct influence on the low-level trajectories of the precipitation particles in the downdraft.

It was shown that the cold-air protrusions (outflow heads) were initially responsible for the appearance of the vorticity couplets (although the extended protrusions were, effectively, a result of the vorticity couplets—they worked as a positive feedback mechanism); however, observations and simulations also indicated that not every splitting storm produces an OHS. A plan-view conceptual model of the structure of the OHS is given in Fig. 19. Figure 20 illustrates in the $x$–$z$ plane the method by which the vorticity couplets that flank the outflow protrusions are generated, and Fig. 21 illustrates the same process for the case of the unidirectional wind profile. It is clear from these figures, then, that the production of the OHS is dependent on the initial vertical configuration of the outflow: the outer edge of the cold pool must be somewhat deeper than the region just inside (i.e., there must be an outflow head), and/or the coldest outflow air must be at the outer edge of the cold pool. If this condition is not met, then an internal vortex couplet cannot form at the rear side of the storm, since parcels will not descend after rising up and over the leading edge of the cold air, and vorticity will not be tilted into the opposing vertical direction. Thus, only a single vortex couplet will form at the rear side of the storm.

However, the configuration of the outflow is generally dictated by both the wind shear profile and the strength of the cold pool, although the dependence of the OHS process on the latter has not been examined here. If there is too little vertical wind shear, outflow...
spreads more uniformly, and the cold-air protrusions and nonuniform depths cannot develop. If there is too much shear, precipitation is advected so far downstream that either the outflow cannot configure itself at the rear side of the storm, or the storm moves too quickly to allow such a configuration to occur. It is nevertheless possible that Owl Horn signatures may develop on higher-shear storms later in their life cycles.

Fig. 16. Same as Fig. 14, but for parcel (32.5 km, 20.5 km, 750 m) located in the region of anticyclonic vorticity associated with the right-flank Owl Horn.
once they have become better organized. Since storms in high-shear environments often require more time to become established (Weisman and Klemp 1982), it is reasonable that, if the higher-shear simulation cases were extended temporally, an OHS might appear, despite a lack of evidence of the feature during the 2-h simulations. It is a safe conclusion that the simulations successfully discerned the spectrum of hodograph

Fig. 17. Same as Fig. 14, but for parcel (21 km, 26 km, 750 m) located in the region of anticyclonic vorticity associated with the left-flank Owl Horn.
shapes, which produce an OHS, but temporal extension or higher-resolution model simulations could perhaps better define an upper limit on necessary hodograph magnitude.

The low-level winds also influence the motion of the outflow. Winds more perpendicular to the boundary impede its progress in the direction opposing the low-level flow, and allow for the outer edge of the boundary to acquire a greater depth by a damming effect. Whereas when low-level flow is parallel to the side-flank outflow boundaries (as in the case of the unidirectional wind profile), no resistance to the outflow’s expansion is encountered, and the outflow can spread out with a more uniform depth. Curvature of the hodograph (particularly in the low levels) is required, then, because it brings about a low-level wind component opposing the motion of the outflow boundary, and it reorients the horizontal streamwise vorticity vector from a direction more parallel to the outflow boundary to one with a component across the outflow boundary (thus amplifying cross-boundary flow).

The presence of outflow heads (Xu 1992) along the side flanks of storms is essential to the development of the Owl Horn signature, but it is not clear in this research precisely how the heads develop. A certain range of vertical shear directed normal to and outward from the outflow boundary enhances the lifting along a density current and the formation of a deep head (Rotunno et al. 1988; Xu 1992). Since the flanks of the outflow boundaries are oriented approximately meridionally and the environmental hodograph is indicative of low-level shear mainly parallel to the right and left flanks of the outflow, it is not obvious why heads appear on both flanks (Fig. 10, middle panel).

Although the shear profiles in simulations of Xu (1992) were only two-dimensional, supercritical heads (much akin to the outflow head seen in cross sections of perturbation potential temperature in the ARPS simulations) were observed in the density current evolution. Studies of outflow–storm interaction have generally focused on the role the outflow boundary plays in sustaining squall lines and mesoscale convective systems (e.g., Weisman 1993). Nevertheless, the authors are not aware of any study that examines the effects of small-

![Figure 18](http://journals.ametsoc.org/mwr/article-pdf/133/9/2608/4219436/mwr2992_1.pdf)
scale changes to the outflow configuration on the dynamic storm processes of supercells.

Through numerical simulations, the process by which the OHS develops has been identified, and its dependence on the environmental wind shear has been examined. Every storm, both real and simulated, which produced an OHS exhibited several signs of a forthcoming split concurrent with the appearance of the signature: a second precipitation shaft developing on the left side of the storm with respect to its motion in both observed and simulated storms, a second vertical vorticity couplet in the midlevel simulated wind field, and a second maximum in reflectivity on the left side of the simulated storm with respect to its motion direction that becomes more separated with height. Soon after the OHS appears, the storm split occurs in both observed and simulated cases. While not every splitting storm exhibits an OHS, every storm that exhibited an OHS underwent a splitting process soon after. Based on this study, it is suggested that the OHS is an indication of a strengthening supercell and an imminent storm split.

Several Owl Horn echoes were observed with the UMass radar (newly upgraded to include Doppler capabilities) during the 2004 storm season, including storms on 12 and 29 May, but unforeseen circumstances prevented Doppler data collection. In addition, the signature was observed with the Shared Mobile Atmospheric Research and Teaching (SMART) Radar (Biggerstaff and Guynes 2000) on 15 May 2003 in a storm near Lela, Texas. But of great interest was the documentation of two instances of the OHS on the WSR-88D radar at Amarillo, Texas. The feature was observed on 17 and 21 June 2004 in close proximity to the radar. It was hitherto thought that the WSR-88D network would be incapable of resolving the OHS owing to too infrequent scanning of the low levels of the atmosphere (where the signature is apparent) and its too coarse spatial resolution [although the Open-system Radar Data Acquisition (ORDA), which boasts 250 m and 0.5° resolution (Saffle et al. 2003), has a better chance of resolving the feature more routinely]. But in light of these recent observations, the forecasting utility of the OHS is apparent, since it can alert operational forecasters to the intensity of a storm [in some instances, well ahead of time—in the case of WFO Amarillo, a full 45 min before baseball-size hail ravaged western Amarillo, and almost 20 min before the first of eight tornadoes was reported on 21 June (SPC storm data)]. In all of these cases, storms were observed to

![Fig. 19. A conceptual diagram of the Owl Horn, including the relevant variables. Color contours are the reflectivity field at 250 m AGL. The shaded region identifies the cold-air protrusions. The outer and inner unshaded bounded areas (black contours) are both regions of vertical motion and regions of cyclonic (+) and anticyclonic (−) vertical vorticity. Updrafts are represented by U. Stippled arrows illustrate the general storm-relative low-level flow. Color scale denotes reflectivity factor in dBZ. The line labeled A identifies the location of the x–z cross section in Fig. 20.](http://journals.ametsoc.org/mwr/article-pdf/133/9/2608/4219436/mwr2992_1.pdf)
Fig. 20. A conceptual model in the $x$-$z$ plane (vertical cross section along line A denoted in Fig. 19) to explain how the couplets of vertical vorticity $\zeta$ are generated by rearward protrusions of relatively colder air in the outflow. (a) Outflow (filled gray region) initially moves outward, with weak upward motion along the outer edges and weak downward motion on the inside (stippled arrows). Parcels are initially moving in the sense of the vorticity vectors (solid arrows) since nearly all their vorticity is in the streamwise direction ($\omega_z$). (b) As parcels approach the oncoming outflow, they encounter a gradient of vertical motion, which tilts the horizontal component of their streamwise vorticity vertically. Stretching also acts on the tilted vorticity to enhance it. (c) As parcels reach the peak of the cold-air protrusions, their vertical motion levels out and the vertical vorticity weakens as tilting begins acting to generate less positive vertical vorticity. (d) As parcels begin their descent, they encounter a gradient of downward motion, which tilts their horizontal vorticity into the vertical again, but in the opposite direction. Thus, the vertical vorticity changes sign. (e) The process continues as the outflow expands farther. (f) Vertical vorticity couplets [cyclonic (C) and anticyclonic (A)] are generated and work in tandem to amplify the flow along their interface so that the cold-air protrusions and radar targets may be advected farther rearward, creating the OHS in the outflow and reflectivity fields.

Fig. 21. Same as Fig. 20, but to explain how the vorticity couplets are generated along the side flanks when outflow heads are not present, as in the case of the unidirectional wind profile. Parcels in this case are moving into the page and parallel to the outflow boundary, as the flow is unidirectional. The horizontal vorticity vector is toward the left, perpendicular to the parcel motion and completely in the crosswise direction. (a) Outflow (filled gray region) initially moves outward, with weak upward motion (stippled arrows) along the outer edges. Since there is no resistance to its outward motion, the cold pool spreads along the side-flanks with a more uniform depth, and neither an outflow head nor substantial downward motion develops over the cold pool. (b) As the oncoming outflow approaches the parcels, they encounter a gradient of vertical motion, which tilts the horizontal component of their vorticity vertically. Stretching also acts on the tilted vorticity to enhance it. (c) As parcels are lifted, their trajectories begin to level off and tilting diminishes. (d) As the parcels level off, their vertical vorticity approaches zero. (e) Unlike in the presence of outflow heads, the parcels never encounter a region of downward motion to tilt the horizontal vorticity in the opposite direction. (f) Therefore, no second opposing vorticity maximum develops on the inner side of the outflow boundary, and only one vorticity couplet [cyclonic (C) and anticyclonic (A)] is established. It may be noted that while the outflow has a component of motion out of the page, the environmental winds oppose its motion. Moreover, any parcels lifted by this part of the outflow boundary do not experience tilting of their vorticity because the vorticity vector is perpendicular to the environmental winds and therefore parallel to the oncoming boundary.
split soon after the signature appeared, and all went on to produce documented tornadoes. Although no evidence beyond the observational has yet been established, perhaps there is an association between Owl Horn production and greater likelihood of tornadic circulations developing in the storm at a later time owing to enhanced low-level cyclonic vertical vorticity generation along the right-flank gust front.

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APPENDIX

Scale Analysis of the Radar-Reflectivity Equation

Consider a conservation equation for a reflectivity pattern $R$:

$$ \frac{\partial R}{\partial t} + u \frac{\partial R}{\partial x} + v \frac{\partial R}{\partial y} + w \frac{\partial R}{\partial z} = \text{Condensation & Evaporation} + \kappa \nabla^2 R, \quad (A1) $$

where, on the left-hand side of Eq. (A1), term 1 is the local time rate of change of the reflectivity pattern, terms 2 and 3 are the horizontal advection of the pattern, term 4 is the vertical advection of the pattern; and on the right-hand side, term 1 represents reflectivity sources (condensation) and sinks (evaporation) omitting other microphysical processes, and term 2 is the pattern’s turbulent diffusion. For TREC to function properly, it must be assumed that radar targets are predominantly moved around by the quasi-horizontal (planar) winds; that is, that terms 1, 2, and 3 on the left-hand side dominate Eq. (A1). (Term 1 is included in the list because it must necessarily be of the same order of magnitude as the advection terms, since it is expanded from the Lagrangian form of the equation.) A scale analysis may be used for two cases of hypothetical data obtained with the UMass X-band radar: the first taken in the central part of a convective storm, and the second at the back side of the storm, in the vicinity of the OHS. The source and sink terms may be estimated by examination of two reflectivity plots of a storm to see how the reflectivity magnitude changes with time at a point within the storm.

a. Worst case: Case 1

Horizontal winds in the storm are on the order of $U = V \sim 10 \text{ m s}^{-1}$; maximum vertical velocity is around $W \sim 50 \text{ m s}^{-1}$; horizontal scale $L_x = L_y \sim 5000 \text{ m}$; vertical scale $L_z \sim 5000 \text{ m}$; $\delta_x R = \delta_y R \sim 50 \text{ dBZ}$; $\delta_z R \sim 50 \text{ dBZ}$; $\kappa \sim 100 \text{ m}^2 \text{s}^{-1}$ (Sasamori 1970; Stull 1988). Observe that term 1 on the left-hand side, scaled as $\delta R / T$, is large for small values of $T$. Therefore, to determine the maximum possible contribution of the time rate of change term, $T$ is scaled by $T = \min[(L_x/U), (L_y/W)]$, where the former term provides the horizontal advective time scale, and the latter provides the vertical advective time scale. The choice in scale for $T$ then dictates the $\delta R$ scale, which is chosen to correspond to either the horizontal or vertical advection scale for $R$. For the condensation and evaporation term, changes are considered for a point in the core of a storm:

$$ \frac{\partial R}{\partial t} \sim \delta R / T = 50 \text{ dBZ} \min \left( \frac{5000}{10}, \frac{5000}{50} \right) \text{s} $$

$$ u \frac{\partial R}{\partial x} + v \frac{\partial R}{\partial y} \sim U(\delta_x R / L_x) = 10 \text{ m s}^{-1} (50 \text{ dBZ}/5000 \text{ m}) $$

$$ = 0.10 \text{ dBZ s}^{-1} $$

$$ \kappa \nabla^2 R \sim \kappa (\delta_x R / L_x^2) = (100 \text{ m}^2 \text{s}^{-1}) $$

$$ \times (50 \text{ dBZ}/25 \times 10^6 \text{ m}^2) $$

$$ = 2 \times 10^{-4} \text{ dBZ s}^{-1} $$

$$ w \frac{\partial R}{\partial z} \sim W(\delta_z R / L_z) = (50 \text{ m s}^{-1}) $$

$$ \times (50 \text{ dBZ}/5000 \text{ m}) $$

Condensation and Evaporation $\sim 30 \text{ dBZ}/600 \text{ s} = 0.05 \text{ dBZ s}^{-1}$.

b. More realistic: Case 2

Horizontal winds in the Owl Horn are on the same scale as in the rest of the storm, namely, $U = V \sim 10 \text{ m s}^{-1}$; the vertical wind speeds in the “Owl Horn” are much smaller than in the updraft core of the storm, so $W \sim 1 \text{ to } 10 \text{ m s}^{-1}$; the reflectivity gradient is much weaker than in the main part of the storm, with $\delta_x R = \delta_y R \sim 0.50 \text{ dBZ}$, $\delta_z R \sim 0.10 \text{ dBZ}$, $\kappa \sim 10 \text{ m}^2 \text{s}^{-1}$.
\( \delta R \sim 10 \text{ dBZ} \) over a horizontal distance of \( L_x = L_y \sim 1000 \text{ m} \); also, \( \delta R \sim 10 \text{ dBZ} \) over a vertical depth of \( L_z \sim 1000 \text{ m} \); \( T \sim \min[(L_x/U), (L_y/W)] \), as earlier; the \( \delta R \) scale in the time rate of change term is chosen to correspond with the \( T \) scale, and \( \kappa \sim 100 \text{ m}^2 \text{ s}^{-1} \). At the rear side of a storm, the reflectivity values are much weaker and do not change as much over time, so the condensation and evaporation are virtually negligible and are omitted here:

\[
\frac{\partial R}{\partial t} \sim \frac{\delta R}{T} = 10 \text{ dBZ} \min \left( \frac{1000}{100 \text{ s}}, \frac{1000}{T} \right) \text{ s}^{-1}
\]

\[
= 10 \text{ dBZ} \text{ s}^{-1} \text{ s}^{-1}
\]

\[
u \frac{\partial R}{\partial x}, v \frac{\partial R}{\partial y} \sim U(\delta_x R / L_x) = 10 \text{ m s}^{-1} (10 \text{ dBZ} / 1000 \text{ m})
\]

\[
= 0.10 \text{ dBZ} \text{ s}^{-1}
\]

\[
\kappa V^2 R - \kappa (\delta_x R / L_x^2) = (100 \text{ m}^2 \text{ s}^{-1})
\]

\[
\times (10 \text{ dBZ} / 1000 \text{ m})^2
\]

\[
= 1 \times 10^{-3} \text{ dBZ} \text{ s}^{-1}
\]

\[
\frac{\partial R}{\partial z} \sim W(\delta_z R / L_z) = (1 \text{ m s}^{-1}) (10 \text{ dBZ} / 1000 \text{ m})
\]

\[
= 0.01 \text{ dBZ} \text{ s}^{-1}. \quad \text{A1}
\]

The scale analysis for case 2 indicates that for a relatively nonturbulent dataset (in which vertical motions are weak relative to the horizontal motions) the assumption of horizontally dominant advective motion is a decent one, since only if the scale of the vertical winds approaches the scale of the horizontal winds does any term in the equation approach the magnitude of the horizontal advection terms. However, in case 1, we see that in the region of a storm’s updraft, an assumption of horizontally dominant advective motion probably fails. Since updrafts and downdrafts can move radar targets much more quickly than horizontal winds, new reflectivity patterns can be created when particles penetrate a specified elevation plane from above and below. Therefore, care must be taken when using TREC results obtained in a convective storm’s inner environment. At best, the TREC analysis may be used for general information about motions in the Owl Horn region of the storm with the knowledge that the results are reasonably accurate.

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\[A1\] For the larger end of the \( W \) scale (\( \sim 10 \text{ m s}^{-1} \)), this term approaches 0.10 dBZ s\(^{-1}\), on scale with the horizontal advection terms, but in the lowest levels, upward vertical velocities of this magnitude are not expected.

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**REFERENCES**


Saffir, R. E., R. C. Elvander, and M. J. Istok, 2003: NEXRAD product improvement—Expanding science horizons. Pre-