Production of Near-Surface Vertical Vorticity by Idealized Downdrafts

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ABSTRACT

This study uses an idealized heat sink to examine the possible roles of the wind profile in modulating the production of surface vertical vorticity by a downdraft. The basic vorticity evolution in these idealized simulations is consistent with previous work: the process is primarily baroclinic and produces near-ground vertical vorticity within the outflow. Sensitivity experiments affirm that the only fundamental requirement for downdrafts to produce surface vertical vorticity is the existence of ambient downdraft-relative flow. Vertical vorticity production increases monotonically as the low-level downdraft-relative flow increases from zero up through intermediate values (in these experiments, 10–15 m s$^{-1}$), followed by a monotonic decrease for greater values. This sensitivity has to do with the degree of cooling acquired by parcels as they pass through the idealized heat sink as well as the degree to which horizontal vorticity vectors subsequently attain an orientation that is normal to isosurfaces of vertical velocity. Although the addition of vertical wind shear is not directly helpful to surface vertical vorticity production in these simulations, increased realism of outflow structure is attained in hodographs with ambient streamwise vorticity. Furthermore, the necessary condition of flow through a region of downdraft forcing would in nature probably require the existence of ambient vertical shear. Therefore, shear in the lower troposphere has a possibly important indirect role in modulating the initial production of near-ground rotation.

1. Introduction

Surface vertical vorticity is a necessary precondition for tornado formation. As reviewed in more detail within our companion paper (Dahl et al. 2014, hereafter DPW14), many recent studies have identified downdrafts and outflow as the source of this initial surface vorticity within supercell thunderstorms (e.g., Rotunno and Klemp 1985; Davies-Jones and Brooks 1993, hereafter DJB93; Walko 1993; Wicker and Wilhelmson 1995; Adelman et al. 1999; Davies-Jones 2000, hereafter DJ00; Straka et al. 2007; Markowski et al. 2008; Dahl et al. 2012; Davies-Jones and Markowski 2013; Markowski and Richardson 2014; DPW14; Markowski et al. 2014; Davies-Jones 2015). Among the aforementioned studies, those that evaluated a vorticity budget along trajectories entering the near-surface vorticity maximum have generally affirmed the conceptual model attributed to DJB93 (subsequently expanded by DJ00). The DJB93 mechanism involves baroclinic generation of horizontal vorticity as air parcels descend, which produces vorticity vectors that are inclined relative to the air parcel’s trajectory. At the base of the downdraft as parcels turn horizontally outward, this inclined vorticity component is then tilted upward into the vertical, producing vertical vorticity at ground level.\(^1\)

Despite this well-established importance of downdrafts to surface vorticity, observations have revealed a robust correlation between the magnitude of the ambient (far field) environmental lower-tropospheric vertical wind shear and the probability of significant tornadoes (Doswell and Evans 2003; Markowski et al. 2003b; 2003d).

\(^1\)In numerical models, this is understood to mean “at or below the bottom physical model level.”
Rasmussen 2003; Thompson et al. 2003; Craven et al. 2004). It has been unclear how the environmental vertical wind profile within the storm’s inflow sector relates to the production of surface vertical vorticity within a storm’s outflow air, which has been demonstrated to precede tornadogenesis. It is possible that the aforementioned baroclinic mechanism is sensitive to the environmental wind profile. It is also possible that the barotropic rearrangement of ambient horizontal vorticity (associated with vertical wind shear) may contribute in some situations; DPW14 noted that “some base-state environments favor substantial crosswise storm-relative vorticity at the height where the parcels originate (p. 3042),” and wondered about the mechanisms determining “…how the baroclinic and barotropic vorticity parts will be oriented near the ground (p. 3042).”

In the present work, we seek to isolate the key processes that answer two separate questions. First, what are the basic requirements for initial surface vorticity production by a downdraft? And second, how is the local production of surface vertical vorticity by a downdraft directly influenced by the wind profile? These questions are not new. The theoretical work of DJ00 investigated analytical solutions for this problem using idealized flows past simple obstacles. He showed that an initial unsheared isentropic flow past a fairing would produce mirror-image centers of cyclonic and anticyclonic vertical vorticity when the isentropes were displaced downward in the downstream direction. In the presence of nonneutral static stability, these local downward displacements of isentropes imply baroclinity, and DJ00 showed that the resultant baroclinically generated vorticity lies within the isentropes and has a vertical component along the sloped sides of the isentropic surfaces. Thus, based on the analytical studies of DJ00, one could conclude that the only fundamental requirement for producing vertical vorticity in a downdraft would be prevailing flow through the zone of baroclinic vorticity generation. Davies-Jones et al. (2001) then further explained that the amount of baroclinically generated vorticity ought to be inversely related to the ambient flow speed through the downdraft.

The treatment of DJ00 involved a number of simplifications; for example, the analytical solutions required highly idealized flow configurations with a primary-secondary flow decomposition, and the production of a cold downdraft in his isentropic framework necessitated the assumption of an ambient environment that was unstably stratified. Our study revisits the problem using a full atmospheric model with a wider range of wind profiles, while still employing an idealized heat sink forcing for the downdraft, which permits closely controlled experiments. In this way, the present work represents a bridge between the analytical solutions of DJ00 and a complex sensitivity study that would produce fundamentally different supercell storms in different wind profiles. First, we affirm that the basic requirement for surface vorticity production is prevailing flow through the zone of downdraft forcing. Then, we assess the influences of realistic vertical wind shear (and thus ambient horizontal vorticity that can be barotropically reoriented). We also expand beyond the local downdraft framework of DJ00 by interpreting the resultant meteorological fields at the surface, and their subsequent evolution as the surface outflow develops and matures.

In section 2, we review the idealized modeling approach for our study. In section 3, we show how our simple downdraft approach produces credible surface vorticity maxima within a variety of controlled wind profiles. The discussion in section 4 then provides additional explanation and context for these results, including their applicability to real supercells and tornadoes. We conclude with a summary in section 5.

2. Methods

Our basic model configuration in most ways mirrors the full-physics supercell simulations of DPW14. We perform 3D simulations using CM1 (Bryan and Fritsch 2002) version 16, with horizontal grid spacing of 250 m and vertical grid spacing that is stretched from 100 m at the surface to 250 m aloft. The lowest scalar model level is 50 m AGL, and the model top is 16.5 km AGL. All other settings are identical to those used by DPW14, except as detailed below.

We specifically want to know how the wind profile influences the downdraft’s production of surface vertical vorticity, not about storm-scale changes to a parent supercell’s structure in different environments. Therefore, we artificially trigger simple downdrafts by applying idealized forcing. An artificial heat sink is implemented by adding a cooling tendency to the model equation for potential temperature. This forcing is spherical in shape (with a peak amplitude of $-0.05 \text{K s}^{-1}$, varying as the cosine of distance from its center), having a horizontal and vertical radius of 1.4 km (horizontally centered in the domain and extending from the surface to 2.8 km

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$^2$We performed exploratory experiments using different kinds of downdraft forcing (including pulsed cooling, momentum forcing, and initial bubble triggers). We found only modest sensitivities to these treatments [see plots and discussion by Parker and Dahl (2013)]; the largest surface vorticity production was for a steady zone of applied cooling (i.e., the configuration presented in detail here).
The spatial scales and cooling amplitude were selected via trial and error so as to produce downdrafts with vertical profiles, peak velocities, widths, and surface potential temperature perturbations that were comparable to our previous full-physics supercell simulations (e.g., Dahl et al. 2012; DPW14).

Of course, in nature the total amount of cooling that is possible for an air parcel is linked to the availability and size of hydrometeors as well as the parcel’s initial relative humidity. The horizontal footprint and cooling rate of the heat sink can be varied to account for different amounts of potential cooling, although we do not present such results here for brevity. Our model is run dry (no condensation processes are present), and the environmental sounding has constant potential temperature. Therefore, potential temperature perturbations can only arise from the artificial heat sink: there are no phase changes or hydrometeor loading, and vertical motions do not modify parcel buoyancy (because of neutral static stability). Thus, in the present study, the artificial cooling represents a number of different processes that can contribute to producing negative buoyancy in real supercells.

The environment has no initial vertical vorticity, so all vertical vorticity that emerges must be produced by the artificial downdrafts that we instigate. There is no forcing for updraft applied in these simulations, so the eventual concentration of surface vorticity into a deep tornado-like vortex3 is not considered. This approach is not unlike that used in other previous studies (Eskridge and Das 1976; Walko 1993; Markowski et al. 2003a; Straka et al. 2007; Davies-Jones 2008; Markowski and Richardson 2014). The primary difference in this case is an emphasis on the interaction of a wide range of wind profiles with only the downdraft. The downdraft is stationary and all winds presented in this study are downdraft relative [although we also interchangeably refer to them as storm-relative (SR), in keeping with common parlance]. For simplicity, we use the same surface wind direction4 at the base of the hodograph in each experiment presented here. However, these results are insensitive to rotation of the hodographs about the origin. In experiments with vertical wind shear, we exclude any shear above 3km AGL for simplicity; sensitivity tests showed that changes to the winds above that level (which is above the top of the artificial heat sink) had negligible impact on the results.

All simulations were run out and inspected for 20 simulated minutes. In the majority of the simulations, mesocyclonic (i.e., 0.01 s−1) vertical vorticity at the lowest model level emerged within the first 6–8 simulated minutes. The general time period for detailed analysis in this study is the first 12–15 simulated minutes, which largely conforms to the time scales for surface vorticity growth in natural supercells (e.g., Markowski et al. 2012; Kosiba et al. 2013). We focus on the specific question of how the downdrafts produce initial surface vertical vorticity; other processes that may become important to tornadoes later, once a corner flow has been established (e.g., as reviewed by Rotunno 2013; Davies-Jones 2015), are not treated in our study. In each run we launch over 9.6 million forward trajectories, spaced every 50 m in the horizontal and vertical within a box (20 km wide in x and y, 3 km deep) that is centered on the heat sink, and integrated during the native model time steps. For the analyses in sections 3 and 4, we use an approach much like DPW14 and create averaged parcel trajectories by selecting parcels that both acquire vertical vorticity ≥0.002 s−1 while descending through 50 m AGL, and have final values of vertical vorticity ≥0.01 s−1 at an altitude below 100 m.

3 Localized intense stretching of vertical vorticity does occur in zones of near-surface convergence, but the depths of the resultant vortices are rather shallow (akin to what might be called “gustnadoes”).

4 We chose an SR surface wind from the east-northeast that matches the mean direction of the surface winds upstream of the main downdraft in DPW14’s simulation. Interested readers can look ahead to section 4c and Figs. 14 and 15a for more information.

5 Properly, in our simulations this is the lowest physical model level, which is at 50 m AGL.
flow would in nature likely require the presence of ambient wind shear (because nonaxisymmetric flow through the downdraft requires that the mean parcel velocities at some altitude differ from the motion of the downdraft forcing itself). But, for a first controlled experiment we view this configuration as the simplest and most consistent with the analytical model of DJ00.

The unsheared runs with 0 and 5 m s$^{-1}$ of SR flow produce negligible $\zeta_{sfc}$ whereas all stronger SR flow speeds produce $\zeta_{sfc}$ exceeding 0.01 s$^{-1}$ by $t = 8$ min (Fig. 1a). These results show that simple downdraft forcing can produce nontrivial $\zeta_{sfc}$ when there is appreciable downdraft-relative flow, even in the absence of vertical wind shear (just as anticipated by DJ00’s analytical solutions). Beyond the basic requirement of SR flow, it is clear that there is a monotonic increase in $\zeta_{sfc}$ as the SR flow increases from 0 up to 12.5 m s$^{-1}$, followed by a monotonic decrease for greater values (Fig. 1a). First we describe the evolution of the seemingly optimal 12.5 m s$^{-1}$ run in more detail, after which we return to the reasons for the sensitivity to the speed of the SR flow.

As clearly exhibited by the simulation with 12.5 m s$^{-1}$ SR winds (Fig. 2), the heat sink produces a surface outflow with realistic potential temperature perturbations (hereafter $u_{sfc}$). A couplet of positive and negative $\zeta_{sfc}$ emerges from the downdraft (Fig. 2), much as in the studies of Walko (1993), DJ00, Straka et al. (2007), and Markowski et al. (2008). In the remainder of this article, we generally focus on the cyclonic member of this couplet [which is typically more intense and more likely to be associated with tornadogenesis, for the reasons described by Markowski and Richardson (2014)] Although the constancy of wind direction and speed with height leads to symmetry about the mean wind (e.g., as seen in the $\theta'_{sfc}$ and $\zeta_{sfc}$ fields in Fig. 2), the basic orientation and magnitude of features in this idealized experiment are quite similar to what has been repeatedly observed and
simulated (see, e.g., DPW14’s Figs. 3, 5, and 10). A river of $\zeta_{\text{sfc}} > 0.01 \text{s}^{-1}$ emanates from the downdraft over time and is concentrated on the outflow’s left (with respect to the mean wind) edge (Fig. 2). With respect to the common orientation of a parent supercell (again, e.g., DPW14’s Figs. 3, 5, and 10), this river of $\zeta_{\text{sfc}}$ would be moving into the typical position of the rear-flank outflow boundary adjacent to the main updraft.

We have verified that only SR flow is needed for the production of $\zeta_{\text{sfc}}$ by the heat sink. How well does the mechanism of vorticity production in this idealized scenario compare to the baroclinic (or DJB93) mechanism reviewed in section 1? The inviscid Boussinesq equation for a parcel’s vorticity vector ($\omega$), neglecting planetary rotation, is

$$\frac{D \omega}{Dt} = \omega \cdot \nabla u + \nabla \times B k,$$  (1)

wherein $u$ is the wind vector and $B$ is buoyancy. The first term on the right-hand side includes tilting and stretching (i.e., barotropic rearrangement of vorticity) and the second term on the right-hand side is baroclinic generation of horizontal vorticity. In the runs with unsheared initial conditions there is no ambient vorticity that can be rearranged, meaning that the barotropic vorticity component (e.g., as defined by Davies-Jones 2006 and DPW14) is zero initially and hence for all time. Thus, it is clear from (1) that in the unsheared runs, the origins of nonzero vorticity must be baroclinic. Indeed, as is shown for an average of 553 downdraft parcels that satisfy the vorticity criteria explained in section 2, all vorticity components remain zero through roughly 300 s, when the air first encounters the heat sink (Fig. 3a). Thereafter, we find that the evolution of the vorticity components is almost identical to what was described in detail by DPW14.

Weak baroclinic generation of horizontal vorticity begins during the period of weak ascent at the edge of the heat sink from $t = 300$ to 360 s, and then significant baroclinic generation occurs once more cooling has accumulated and the air begins its descent in the...
downdraft (Figs. 3a,b,d). Thus, all parcel vertical vorticity originates from reoriented horizontal vorticity that was produced baroclinically in this case (there is no ambient vorticity). Because of the parcel’s position relative to the heat sink’s center (Figs. 3b,c), primarily crosswise horizontal vorticity (hereafter $v_c$) is generated baroclinically (Figs. 3a,b), which amplifies the total vorticity vector and also inclines it upward relative to the trajectory (Figs. 3b,d). From $t = 540$ s onward, this inclined vorticity vector becomes increasingly streamwise (Figs. 3a,c). Initially this is due to the “river bend effect” (i.e., the turn toward the south), which engenders a crosswise-to-streamwise exchange that seems to be common to trajectories in many studies (Adlerman et al. 1999; Markowski and Richardson 2014; Schenkman et al. 2014; DPW14). The streamwise vorticity component (hereafter $v_s$) continues to grow thereafter due to horizontal stretching and continued baroclinic generation as the parcel flows southward in the pool of outflow (Figs 3a,b,d).

Meanwhile, as the trajectory bottoms out, the inclined vorticity vector is tilted upward, producing positive vertical vorticity as the parcel descends below 200 m AGL (Figs. 3a,b,d); this is therefore another reconfirmation of the mechanism explained by DJB93 and DPW14. Along the averaged trajectory, the vertical vorticity component continues to amplify rapidly (especially seen after 660 s in Fig. 3a) as the parcel departs the downdraft and moves into a zone of strong convergence along the outflow edge (with convergence values...
peaking at 0.024 s\(^{-1}\) at \(t = 12\) min). Although not shown, there are mirror-image companion trajectories that produce the opposite-signed \(\zeta\) extremum seen on the outflow’s right edge in Fig. 2.

The baroclinic vorticity creation mechanism we describe is fully consistent with the DJB93 trajectory perspective. Not surprisingly (since vorticity vectors are tangent to vortex lines), this mechanism is interchangeably consistent with the perspective of baroclinically generated vortex line “loops” that are reoriented at downdraft edges as proposed by DJ00, Straka et al. (2007), and Markowski et al. (2008). Looking shortly after the horizontal vorticity components become nonzero (cf. Fig. 3a) it is clear that the averaged parcel is attached to a horizontally oriented vortex line loop (Figs. 4a,b). By \(t = 10.5\) min (Fig. 4c) there is obvious downward displacement of the upstream (northeastern) side of the loop, just as described by DJ00, Straka et al. (2007), and Markowski et al. (2008). This is where the descending parcel is found to be acquiring positive vertical vorticity via the DJB93 mechanism. By \(t = 12\) min (Fig. 4d), the averaged parcel has descended to the lowest model level, and its large positive \(\zeta_{sfc}\) is attached to one of the “vortex line arches” described by Straka et al. (2007) and Markowski et al. (2008). The only difference is that our present simulations have no imposed updraft, and hence the loops are not lifted up on their downstream edges. Rather, the loops are only noticeably reoriented downward (by the downdraft) on their upstream edges, extending toward the surface in a manner that is reminiscent of the surface mesovortex generation mechanism described for squall lines by Trapp and Weisman (2003).

Having established that the familiar baroclinic mechanism is operating in these simulations, we now return to the question of the differences among the runs with varying SR flow. Compared to the 12.5 m s\(^{-1}\) run, simulations with both stronger and weaker SR flow produce smaller \(\zeta_{sfc}\) values (Fig. 1a). In the limit of no SR flow, an axisymmetric and comparatively cold

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**Fig. 4.** Replication of the averaged parcel trajectory and vorticity vectors from Fig. 3 as viewed from the south-southeast at (a) \(t = 7.5\), (b) \(t = 9\), (c) \(t = 10.5\), and (d) \(t = 12\) min. The trajectory is depicted from its starting point through the valid time for each panel, with contemporaneous surface vertical vorticity shaded (s\(^{-1}\)), and the vortex line passing through the instantaneous parcel position plotted in magenta.
outflow is produced with negligible $\xi_{\text{sfc}}$ (Fig. 5a). For the stronger SR flows, the resulting outflows are plume shaped and less cold, with elongated zones of small $\xi_{\text{sfc}}$ along their left and right edges (Fig. 5d). The intermediate values of SR flow (e.g., Figs. 2 and 5b,c) represent transitions between these two extremes, with maximized $\xi_{\text{sfc}}$ production due to two opposing effects, which we explore next.

On the one hand, increasing the SR flow reduces the residence time of parcels in the heat sink, thereby decreasing both downdraft (Figs. 6a,c,e) and cold pool (Figs. 6b,d,f) strength. In turn, this diminished cooling lessens the magnitude of baroclinic horizontal vorticity production (e.g., vectors in Figs. 6b,d,f). This effect would generally favor greater $\xi_{\text{tot}}$ under weaker SR flow. This importance of long parcel residence times within a zone of baroclinic horizontal vorticity generation has been known at least since the full supercell simulations of Rotunno and Klemp (1985). The decline in a downdraft’s vertical vorticity production under vanishing SR flow is also consistent with the analytical model of DJ00, as reviewed by Davies-Jones et al. (2001).

On the other hand, in the limit of zero SR flow (Fig. 6a), negligible tilting of the baroclinically generated horizontal vorticity occurs, as we explain next. Emergence of vertical vorticity is shown formally by the vertical component of (1):

$$\frac{D\xi}{Dt} = \mathbf{\omega}_h \cdot \mathbf{V}_h + \frac{\partial \mathbf{w}}{\partial z},$$

wherein the first term on the right-hand side is tilting of horizontal vorticity and the second term is stretching of vertical vorticity. For our initial conditions with no ambient vertical vorticity, it is clear that $\xi$ may only arise from tilting. In turn, tilting requires that $\mathbf{\omega}_h$ is not normal to $\mathbf{V}_h$ (which is to say that $\mathbf{\omega}_h$ cannot be parallel to the $w$ contours). As shown by Figs. 6b,d,f, the angle between $\mathbf{\omega}_h$ and the $w$ contours, and thus the potential for tilting, increases with increasing SR flow (i.e., the flow through
the downdraft); we discuss the reasons for this in more detail in section 4a.

In summary, the two opposing effects described above combine to yield optimum \( \zeta_{sfc} \) production for intermediate SR flow magnitudes. For weaker SR flow, there is ample cooling and horizontal vorticity generation, but improper orientation of \( \mathbf{v}_h \) for tilting (Figs. 6a,b). For stronger SR flow, there is an optimal orientation of \( \mathbf{v}_h \) for tilting, but only minimal cooling and horizontal vorticity generation (Figs. 6e,f). For our particular heat sink’s shape and cooling rate, it appears that SR flow of 12.5 m s\(^{-1}\) optimizes the combined contributions of these two effects (Figs. 6c,d). We use this 12.5 m s\(^{-1}\) flow as a baseline for the remainder of the experiments.

b. Streamwise profiles with varied shear

From the previous experiments, it is clear that when a downdraft’s horizontal movement is allowed to be independent of the wind field (i.e., SR flow exists), then vertical wind shear is not a fundamental requirement for downdraft production of \( \zeta_{sfc} \). However, the symmetry of the resulting outflow and vorticity maxima does not seem to be especially realistic. In addition, neither supercells, nor nonzero downdraft-relative flow, are likely to occur in nature without the presence of at least some vertical wind shear. Thus, we still wish to address the degree to which shear influences downdraft production of \( \zeta_{sfc} \).

For this next set of experiments, we use the seemingly ideal 12.5 m s\(^{-1}\) SR wind magnitude at every level and ask whether the addition of shear to the profile helps or hinders \( \zeta_{sfc} \) production. To address this question, we construct hodographs with purely streamwise vorticity but vary the vertical wind shear from 0 to 0.0125 s\(^{-1}\), in effect creating segments of a full circular hodograph having radius 12.5 m s\(^{-1}\) (Fig. 7a).

The streamwise runs with the least vertical shear produce the highest \( \zeta_{sfc} \) (Fig. 1b). Small amounts of shear (up through about 0.0025 s\(^{-1}\)) do not change the results much. In contrast, for shear \( \geq 0.005 \) s\(^{-1}\), the production of \( \zeta_{sfc} \) decreases monotonically with increasing shear (Fig. 1b). In terms of the evolution of surface fields, the lower shear runs (Figs. 8a,b) are largely similar to the previous unsheared run with
identical SR flow (Fig. 2d). The $\zeta_{sfc}$ and $\theta_{sfc}$ fields become increasingly asymmetric with increasing shear, with preference for a left-flank river of cyclonic vorticity that is stronger than the right-flank river of anticyclonic vorticity (Figs. 8b–d, with “left” and “right” defined with respect to the ambient SR flow). As such, the runs with shear look increasingly realistic, and bear a much closer resemblance to the full physics simulations of DPW14. Despite this increased realism and preference for cyclonic vorticity, the overall peak values of vertical vorticity are diminished with increasing shear. For a closer look at the reasons for this, we present the details of the 0.0075 m s$^{-1}$ run, which is beyond the apparent tipping point where $\zeta_{sfc}$ begins to fall off rapidly (Fig. 1b).

The basic evolution of vorticity along an averaged trajectory for the 0.0075 m s$^{-1}$ run (for 378 downdraft parcels that satisfy the vorticity criteria in section 2) is largely similar to that described for the 12.5 m s$^{-1}$ unsheared run (cf. Figs. 3 and 9), and again conforms to the DJB93 mechanism. Here we highlight the primary differences from the unsheared runs.

First, in contrast to the unsheared runs, the averaged trajectory begins with ambient streamwise horizontal vorticity due to the vertical shear (Fig. 9a). As the parcel begins to descend, downward tilting of the original streamwise vorticity causes the parcel’s vertical vorticity to become negative (e.g., Davies-Jones 1984; DBJ93). This is evident from the appearance of negative vertical vorticity from $t = 240$ to $360$ s (Fig. 9a), as well as the increasing downward inclination of the vorticity vectors (Figs. 9b,d). Second, even though the ambient SR wind speed is constant with height (Fig. 7a), the accumulated cooling in the sheared runs decreases with increasing vertical wind shear (Fig. 8). This decrease occurs because the downdraft’s nonzero $\zeta$ aloft (explained above) is associated with rotational wind perturbations (interested readers can look ahead to Fig. 13 for examples of this); ultimately, the perturbed wind field reduces the average parcel residence times within the heat sink’s interior zone of strongest cooling. From the perspective of the averaged trajectory, the impact of the lesser accumulated cooling is to decrease the accumulated baroclinic generation of horizontal vorticity; this is most easily seen in the time series of horizontal vorticity values (cf. Figs. 3a and 9a) and the vertical cross-sectional views of the vectors (cf. Figs. 3d and 9d). As a result, there is less horizontal vorticity available for tilting into the vertical at the base of the downdraft.

We return to the question of how shear influences the outflow structure and distribution of cooling later in section 4b, after a wider variety of sheared runs have been presented.
The streamwise experiments show that ambient shear is neither necessary nor directly helpful\textsuperscript{6} to downdraft production of $\zeta_{\text{sfc}}$. We next present several alternative ways of implementing vertical shear in order to assess whether the presence of ambient crosswise vorticity is relevant to the results.

c. Profiles with varying shear angle

In our final baseline experiments, we fix the surface SR wind magnitude\textsuperscript{7} at the seemingly ideal value of 12.5 m s$^{-1}$, and fix the magnitude of the vertical shear (i.e., magnitude of the horizontal vorticity) at 0.0075 s$^{-1}$. The angle between the horizontal vorticity vectors and the wind vectors is then varied (allowing the relative magnitudes of the streamwise versus crosswise vorticity to vary), using two different approaches. In the first approach, we use straight-line hodographs, and vary the orientation of the hodograph with respect to the wind direction (Fig. 7b). The reference profile ("S0") has ambient horizontal vorticity that is purely crosswise. We then test straight-line hodographs of the same length that have been rotated 15°, 30°, and 45° to the left (counterclockwise) of this reference profile ("S15," "S30," and "S45"); the increasing angles provide an increasingly streamwise orientation of the initial horizontal vorticity. In the second approach, we use the 0.0075 s$^{-1}$ streamwise run as a baseline, and then test curved hodographs in which the angle between the horizontal vorticity vector and wind vector at each level varies among $+30^\circ$, $+15^\circ$, $-15^\circ$, and $-30^\circ$, which modifies the concavity of the hodograph (Fig. 7c).

Perhaps surprisingly, given the identical surface winds and shear magnitudes, the highest and lowest $\zeta_{\text{sfc}}$ values among these experiments deviate by more than a factor

\textsuperscript{6}By "directly helpful" we are referring to the immediate production of vertical vorticity at the base of the downdraft. In section 4c, we discuss a number of indirect effects of vertical shear that are likely to be important in real storms.

\textsuperscript{7}We also ran this "shear angle" experiment with a fixed SR wind speed of 12.5 m s$^{-1}$ at 1.5 km AGL (instead of the surface) for comparison. The basic findings were unchanged in these profiles, but the overall spread in $\zeta_{\text{sfc}}$ values among simulations was much less. We, therefore, chose to present the runs with a fixed surface wind in order to explore their greater variability.
of 2 (Figs. 1c,d). A comparison among the hodographs (Figs. 7b,c) reveals that the simulations ultimately producing the highest $\zeta_{sfc}$ (S45, S30, and C15R) have deeper layers that fall within or near the seemingly optimal 10–15 m s$^{-1}$ SR flow range (imposed circles in Fig. 7). In other words, when the SR flow speed is allowed to vary with height (unlike in the unsheared or purely streamwise experiments), the benefits of intermediate SR flow through a deep layer emerge. We had originally hypothesized (section 1) that the presence of ambient crosswise vorticity might be beneficial to vorticity generation in the downdraft, but this notion is not supported by the present results. The runs with the greatest crosswise initial vorticity (S0, C30L, C30R) are among the lowest in terms of $\zeta_{sfc}$ at $t = 15$ min (Figs. 1c,d); and, based upon inspection of averaged trajectories (not shown), there is little evidence that the initial magnitude and orientation of $\omega_c$ substantially influences the downdraft production of $\zeta_{sfc}$.

Although SR flow again emerges as the most important to $\zeta_{sfc}$, this is not to say that the shear and its orientation are irrelevant; the patterns of surface outflow and vorticity vary strongly among the present hodographs. For the environments possessing at least moderate amounts of ambient streamwise vorticity (S30, S45, and all of the curved profiles) the familiar surface patterns from section 3b reemerge (Figs. 10c,d and 11). These runs have asymmetric rivers of vertical vorticity emanating from the downdraft, with magnitudes of positive $\zeta_{sfc}$ on their left flanks that exceed the magnitudes of negative $\zeta_{sfc}$ on their right flanks. It is clear from S30 and S45 (Figs. 10c,d) that hodograph curvature is not needed to induce the asymmetries. Notably, for the asymmetries to emerge in straight hodographs, the motion of the heat sink must lie off of the hodograph (unlike S0); in nature this situation would be akin to storm splitting in a straight-shear environment (e.g., Rotunno and Klemp 1982). We will show in section 4b

FIG. 9. As in Fig. 3, but for the streamwise experiment with shear $= 0.0075$ s$^{-1}$. A total of 378 parcels met the averaging criteria.
that the simulated asymmetry is strongly correlated to the magnitude of the streamwise vorticity in the ambient environment, regardless of hodograph shape.

In short, across a wide range of hodographs, the primary importance of the SR flow to $\zeta_{\text{sfc}}$ is upheld. The tendency for increased realism (e.g., asymmetric outflows and vorticity rivers) with increasingly streamwise ambient vorticity is clearly demonstrated by the straight hodographs (Fig. 10), but it can also be seen among the curved hodographs (cf. Fig. 8c vs Figs. 11a,d). Beyond these considerations, the inclusion of crosswise ambient vorticity does not appear to play a major role in producing $\zeta_{\text{sfc}}$ within our experiments.

4. Discussion

a. Fundamental requirements for reorienting baroclinically generated horizontal vorticity

Our analysis in section 3 highlights the need for non-zero SR flow to facilitate tilting of the baroclinically generated horizontal vorticity. Why is SR flow a requirement? We may expand the horizontal vorticity to rewrite the tilting term in (2) as

$$T = \left( \kappa \times \mathbf{V}_h^w + \kappa \times \frac{\partial \mathbf{V}_h}{\partial z} \right) \cdot \mathbf{V}_h^w = \kappa \cdot (\omega_{h1} \times \omega_{h2})$$

where $\mathbf{v}_h$ is the horizontal velocity vector and $\kappa$ is the vertical unit vector. This shows that horizontal vorticity consists of a component due to horizontal gradients of $w$ ($\omega_{h1}$) and a component due to vertical wind shear ($\omega_{h2}$). Only $\omega_{h2}$ (the “vertical shear” part) can be re-orientated into the vertical (produce nonzero $T$), since $\omega_{h1} = \kappa \times \mathbf{V}_h^w$ is normal to $\mathbf{V}_h^w$ (i.e., $\omega_{h1}$ is parallel to the $w$ contours). It is clear that tilting requires $\omega_{h2}$ to be both nonzero and also nonparallel to the horizontal $w$ contours (i.e., $\omega_{h1}$).
A parcel within horizontal buoyancy gradients acquires vorticity only in the form of $v_{h1}$ because buoyancy acts directly on the vertical velocity component (Fig. 12). As described above, this initial vorticity fundamentally cannot be tilted into $z$. Following (3), tilting into the vertical can occur only in the region where some of the horizontal vorticity takes the form of $v_{h2}$. As the parcel approaches the ground, its descending motion is converted to horizontal motion by a pressure maximum at the base of the downdraft (Figs. 12b,e). This horizontal outflow from the downdraft contains vertical shear (or $v_{h2}$), as seen in Figs. 12c,f. It can be shown that $v_{h2}$ results from conversion of $v_{h1}$ by horizontal gradients of the vertical perturbation pressure gradient force (Figs. 12c,f), while the total horizontal vorticity is unaffected.\(^8\) Satisfying the additional requirement that $v_{h2}$ is not parallel to $v_{h1}$ depends upon the magnitude of the SR flow. As shown in Fig. 6, stronger flow through the downdraft produces a larger angle between $v_{h}$ and the $w$ contours (i.e., between $v_{h1}$ and $v_{h2}$); this is because the vertically sheared outflow increasingly deviates from axisymmetry under stronger SR flow (as discussed in section 3a). If ambient shear is present, the above argument still applies to the baroclinically generated part of the horizontal vorticity, but in addition the ambient vorticity (i.e., initial $v_{h2}$) can also be barotropically rearranged by the downdraft (e.g., DPW14).

\(8\) Ultimately, this follows from the perturbation pressure-gradient force being irrotational in the Boussinesq limit. Note that during the conversion, additional $v_{h1}$ may still be generated baroclinically, as is apparent in Fig. 12.

b. Impact of streamwise vorticity upon outflow symmetry

Another noteworthy aspect of our results is that the cyclonic river of $\zeta_{sfc}$ along the outflows’ left flanks strengthens with distance from the downdraft due to the sustained gust front convergence there (e.g., Figs. 3 and 9). Certainly, it is not surprising that convergence amplifies the surface vertical vorticity. These zones of
vertical vorticity that amplify via convergence along the outflow edges are quite reminiscent of the vortex sheets produced in a full-supercell simulation by Markowski et al. (2014). What is interesting is that the degree of this postdowndraft amplification varies widely among the different wind profiles we have tested.

As described in section 3, the unsheared runs and purely crosswise straight-line hodograph produced symmetrical outflows (Figs. 5 and 10a). Every simulation possessing ambient streamwise vorticity produced an asymmetric outflow (Figs. 8, 10b–d, and 11). As reviewed in section 3b, the existence of ambient streamwise vorticity leads to the production of negative vertical vorticity above the surface in the downdraft through tilting. Thus, the sheared environments with positive streamwise vorticity (which is the case for all of the sheared runs here except for S0) have anticyclonic wind perturbations within their downdrafts, typically peaking in amplitude at or just below 1 km AGL. This is exemplified by the S45 case in Fig. 13 (where it is compared to an unsheared run and the purely crosswise run). We computed the averaged vertical vorticity enclosed by the $\omega_{h2}$ contour at $t = 5$ min for $z = 1$ km. The correlation (across all simulations) between this downdraft vertical vorticity and the ambient environmental streamwise vorticity is $-0.96$

This net circulation in the downdrafts leads to asymmetries in the cold outflow at lower levels. Although it is a topic that quickly becomes tangential to the goals of the present article, under nonzero SR flow, the coldest air within the midlevels of the downdrafts (e.g., at 1 km AGL, which is just below the height of maximal cooling in the heat sink) is found on the downdrafts’ upstream sides (bottom row of Fig. 13). In brief, this occurs because as parcels move laterally through the heat sink over time, they first acquire negative buoyancy, after which they gradually acquire downward velocities, after which they gradually acquire downward displacements.

**Fig. 12.** Downdraft vertical cross sections depicting components of the horizontal vorticity and related fields, for the (a)–(c) SR = 0 and (d)–(f) SR = 12.5 m s$^{-1}$ unsheared runs. The cross sections pass through the downdraft centerline and are parallel to the ambient SR flow. The $x^*$ coordinate is distance along the cross section from the downdraft center point, and in all panels the wind vectors (m s$^{-1}$, scaled as shown) represent the velocity components in the plane of the vertical cross section. The shaded vorticity values (s$^{-1}$) are the horizontal component lying normal to the cross section (i.e., in the $y^*$ direction), and are positive signed for vectors pointing into the page (i.e., toward positive $y^*$). (a),(d) The total horizontal vorticity ($\omega_{h1}$) and potential temperature perturbations (contoured at $\pm 5, \pm 3$, and $\pm 1$ K). (b),(e) The “w gradient” part of the horizontal vorticity ($\omega_{h1}$) and perturbation pressure (contoured every 50 Pa, with thick zero contour). (c),(f) The “vertical shear” part of the horizontal vorticity ($\omega_{h2}$) and the conversion term between the components of $\omega_{h1}$ and $\omega_{h2}$ that are normal to the cross section $[-(1/\rho_0)(\partial \Omega/\partial x^*)(\partial p/\partial z)]$, contoured every 0.0001 s$^{-2}$, with negative values dashed and zero contour omitted.
Progressive time lags between $Dw/Dt$, $w$, and $\delta z$. Given this placement of the coldest air on the upstream side of the downdraft, a net anticyclonic circulation within the midlevel downdraft preferentially advects the pocket of coldest air clockwise toward the south (Fig. 13f, $x = -1 \text{ to } 0 \text{ km, } y = -1 \text{ to } 0 \text{ km}$). In contrast, for the unsheared cases (where very little perturbation flow is induced; Fig. 13b) or purely crosswise cases (where a symmetric couplet is induced; Fig. 13d), this colder upstream air is not preferentially advected across the central axis of the downdraft.

This explains why all of the runs with positive ambient streamwise vorticity have outflows with the coldest air on their southeastern (left) flanks, which in turn explains why the river of cyclonic vorticity that is found there surpasses the magnitude of the anticyclonic river of vorticity on their northern (right) flanks. There is more baroclinic generation of horizontal vorticity and more amplification of $\zeta_{sfc}$ by convergence on their left flanks.

As a statistical test, we constructed a line passing through the origin (heat sink axis) and the surface cold pool's centroid for each run, and then compared the magnitudes of both $\theta'_{sfc}$ and $\zeta_{sfc}$ to the left versus right of that line (normalizing the differences by the peak values found in each run). The ambient streamwise vorticity is strongly correlated with cold pool asymmetry, both in terms of $|\theta'_{sfc}|_{\text{left}} - |\theta'_{sfc}|_{\text{right}}$ (for which $r = 0.73$) and in terms of $|\zeta_{sfc}|_{\text{left}} - |\zeta_{sfc}|_{\text{right}}$ (for which $r = 0.83$). We hasten to add that the simulations with the largest final values of $\zeta_{sfc}$ are not generally those with the largest streamwise vorticity (or largest asymmetries). The computed correlation between ambient streamwise vorticity and peak $\zeta_{sfc}$ is only 0.16, and the purely streamwise simulations in section 3b produced weaker $\zeta_{sfc}$ as the ambient shear was increased. In our experiments, the primary factor in determining the magnitude

FIG. 13. Snapshots of fields at $z = 1 \text{ km AGL, } t = 5 \text{ min, supporting the emergence of symmetrical vs asymmetrical outflows. (a),(c),(e) Vertical vorticity is shaded (s$^{-1}$), vertical velocity is contoured at $-10, -6, \text{ and } -2 \text{ m s}^{-1}$, the ambient SR flow is depicted with a black vector (m s$^{-1}$, scaled as shown), and the ambient horizontal vorticity is depicted with a purple vector (s$^{-1}$, scaled as shown). (b),(d),(f) Potential temperature perturbation is shaded (K), the vertical velocity contours are repeated from (a),(c),(e), and the perturbation horizontal wind vectors are plotted (m s$^{-1}$, scaled as shown). (a),(b) The unsheared run with SR flow = 12.5 m s$^{-1}$; (c),(d) the experiment with a straight, purely crosswise profile (S0); and (e),(f) the experiment with a straight hodograph rotated 45° to the left (S45). This S45 environment is the only one possessing ambient streamwise vorticity. The position of the artificial heat sink is annotated in all panels with a thin cyan contour, and the “center line” discussed in the text is shown as a thick black line (which passes through the center of the heat sink and the grid point with the strongest downward velocity).
of $\zeta_{\text{sfc}}$ appears to be the SR flow, but the inclusion of streamwise vorticity produces more realistic outflows with the kinds of predominant cyclonic rivers of $\zeta_{\text{sfc}}$ that were studied by DPW14. We next try to connect our findings more closely to those of DPW14 by using a profile from their simulation within our idealized model.

c. Applicability to real supercells and tornadoes

1) HODOGRAPH FROM A SIMULATED SUPERCELL

The simulations presented in section 3 were largely designed to be instructive. It is fair to ask whether these results are pertinent to real supercells. Using our idealized heat sink, we incorporate a wind profile from the Del City, Oklahoma, supercell simulation of DPW14. We use a time- and space-averaged wind profile from the adjacent forward flank region of DPW14’s simulation (the box shown in Fig. 14), which is the region from which high vorticity downdraft parcels originate in the DPW14 study (e.g., their Fig. 3). The resulting 0–3-km hodograph from that part of the storm is rather similar to the ambient base state’s 0–3-km hodograph (Fig. 15a), just slightly backed (presumably due to the parent storm’s cyclonic rotation). This is somewhat reassuring because it implies that the environmental SR winds and streamwise vorticity are not a bad first guess for what is directly upstream of the downdraft within a supercell.

The DPW14 hodograph is most similar to the streamwise 0.005 s$^{-1}$ and C15R runs, and it produces similar trends in $\zeta_{\text{sfc}}$ (albeit higher by $\approx$20%). But, in short, the more realistic initial profile produces credible values of surface vorticity (values in the vorticity rivers of DPW14 were generally $1–3 \times 10^{-2}$ s$^{-1}$, as seen in their Figs. 4 and 5). Given the rather similar initial hodographs, it is probably not surprising that the evolving cold pool in the DPW14 experiment (Fig. 16) also closely resembles that produced in the streamwise 0.005 s$^{-1}$ experiment (Fig. 8b). It exhibits a realistic asymmetric cold pool (Fig. 16), with a predominant river of cyclonic vorticity on its eastern flank that flows southward toward (what would be) the supercell’s rear flank (very similar to DPW14’s Fig. 4), with a

![Fig. 14. Time-averaged fields from the primary simulation of DPW14: pseudoradar reflectivity at 1 km AGL (shaded, dBZ), surface wind vectors (m s$^{-1}$, scaled as shown), and vertical velocity at 1 km AGL (contoured in black at −4, −2, 2, and 4 m s$^{-1}$, negative values dashed). The fields are averaged over the period from 4200 to 5400 s (70–90 min; i.e., the range of times analyzed in detail by DPW14). The white box in the northeastern part of the storm indicates the area over which the winds were horizontally averaged to produce the near-downdraft wind profile used in the present study (see hodograph in Fig. 15a).](image-url)
corresponding zone of comparatively cooler air on its eastward-facing edge (very similar to DPW14’s Fig. 10).

Perhaps the most important comparison is the evolution of vorticity vectors along the downdraft trajectories, which was the primary focus of DPW14’s study. Using an average of 258 parcels that satisfy the criteria in section 2, we find that the basic topology of the trajectory (Fig. 17) is similar to those studied by DPW14.

Fig. 15. Hodograph diagram and history of maximal surface vertical vorticity (s$^{-1}$) vs time for the DPW14 near-downdraft run. (a) As in Fig. 7, but showing the original base state wind profile of the DPW14 simulations (magenta) and the averaged near-downdraft profile (taken from the box in Fig. 14). (b) As in Fig. 1, but for the DPW14 near-downdraft experiment. Note: the range plotted in (b) differs from that in Fig. 1.

Fig. 16. As in Fig. 2, but for the run using the DPW14 near-downdraft profile.
(cf. their Figs. 7a and 8a). Parcels flowing west-southwestward turn due southward while descending (Fig. 17b), the horizontal component of vorticity grows large and points primarily southward by the nadir of their descent (Figs. 17b,c), and emergence of positive vertical vorticity through tilting occurs quite close to the ground (near 250 m AGL in the present experiments, Figs. 17a,d; closer to 150 m AGL in DPW14).

Although we acknowledge the extreme simplicity of the present treatment, it appears that the downdraft production of surface vertical vorticity in a supercell can be reasonably idealized in terms of the interaction between the prevailing flow and the downdraft forcing.

2) IMPLICATIONS FOR TORNADGENESIS

Our experiments have specifically focused on the initial production of $\zeta_{sfc}$, which is a necessary but not sufficient prerequisite for tornadogenesis. All things being equal, the initial production of larger values of $\zeta_{sfc}$ would seem to be a more favorable precondition for tornadogenesis, although it is certainly possible that real tornadoes require only minimal nonzero $\zeta_{sfc}$ to be available and stretched. However, in order to connect the present results to natural tornadogenesis, one must address the likelihood that high vorticity surface parcels will actually be lifted out of the cold pool and stretched into a tornado. Our experiments do not include a parent updraft, which limits the direct assessment of this final part of tornadogenesis. In the present framework, we find that the river of cyclonic vorticity is located along what would be the “updraft facing” side of the outflow, suggesting the potential for this vorticity to be stretched by the parent updraft of a real storm. Of course, it is possible that a natural storm with a realistic rotating updraft (and associated pressure perturbations) might have trajectory shapes and vorticity river orientations that differ from what we have produced. We find it encouraging that our surface outflows and averaged parcel
trajectories strongly resemble those that occur in our full-physics supercell simulations (e.g., DPW14).

It has been repeatedly observed (e.g., Markowski et al. 2002; Grzych et al. 2007) that significantly tornadic storms are associated with less cold outflows. Markowski et al. (2008) suggested that this may be an example of a “Goldilocks’s problem,” in which the outflow must experience some cooling in order to facilitate sufficient baroclinic generation of horizontal vorticity during descent, but must not be so cold that the outflow parcels with large vorticity cannot be lifted and stretched. This has an interesting parallel in the present experiments, where intermediate values of SR flow produce intermediate amounts of cooling and maximize the production of \( \zeta_{\text{sfc}} \) (section 3a). In the present simulations with the largest \( \zeta_{\text{sfc}} \), the coldest \( \theta'_{\text{sfc}} \) values are still largely in the \(-3\) to \(-4\)-K range, which would not be exceptionally cold in the context of previous studies (e.g., Markowski and Richardson 2010, their Fig. 10.14). Of course, in real supercells there is a wide range of thermodynamic and microphysical sensitivities that could also influence outflow temperatures, and it would also be of interest to try to isolate such influences from the direct effect of SR flow illuminated in the present experiments.

At first sight, our results seem to be at odds with the notion that large lower-tropospheric vertical wind shear and SRH are beneficial for tornadoes (Doswell and Evans 2003; Markowski et al. 2003b; Rasmussen 2003; Thompson et al. 2003; Craven et al. 2004). It is likely that their importance to tornadoes hinges on processes not included in our study. For example, SRH is a good predictor for supercells (e.g., Davies-Jones 1984), whereas our experimental design does not include the rotating parent supercellular updraft and its possible feedbacks upon the surface vorticity evolution (e.g., locations and magnitudes of evaporative cooling and upward accelerations by dynamic perturbation vertical pressure gradients). Indeed, Markowski and Richardson (2014) showed that the dynamic lifting of vorticity-rich outflow air into a supercell’s updraft is strongly promoted by large SRH values. Furthermore, in a typical curved environmental hodograph, SRH is closely linked to the magnitude of the SR flow. It is encouraging that the optimal values for SR flow in the present experiment are above the minimal value of 10 m s\(^{-1}\) that Droegemeier et al. (1993) found to be necessary for supercells in a curved hodograph (a result that they linked to SRH). The parent storm’s rotation, and the orientation of the SR flow, also influence the horizontal distribution of hydrometeors (e.g., as reviewed by Markowski 2002; Dawson et al. 2015), which could modulate the location and intensity of the vorticity-producing downdrafts.

Some recent studies (e.g., Schenkman et al. 2014) have suggested that friction may be a substantial contributor to the vorticity budget along trajectories entering a developing surface vortex. Although we have not reported on them in detail, we reran the majority of our experiments using a simple no-slip bottom boundary condition. We found that the no-slip runs were almost identical to our free-slip runs (which are presented throughout this paper) through roughly 10 min, after which they began to lag behind in terms of peak \( \zeta_{\text{sfc}} \). This slight lagging was associated with the gradual slowing of the low-level outflow winds, which weakened the gust-front convergence where \( \zeta_{\text{sfc}} \) is typically amplified in these simulations. In general, surface drag has little influence on the initial appearance of surface vertical vorticity in our experiments. This is certainly not to discount the possibility that surface drag is important in some sense to tornadic storms. But, this supplemental result does give us more confidence that baroclinic generation is the key source for the onset of surface vertical vorticity at the base of a downdraft.

Ultimately, an understanding of the full interplay among the environmental wind profile, the structure of the parent supercell, and the details of the surface vorticity evolution will likely require numerous full-physics supercell sensitivity simulations at very high resolution. Such large projects may become computationally feasible in the coming years.

5. Conclusions

The study was motivated by the question of how the environmental vertical wind profile relates to the likelihood of intense surface vortices in supercells. Here we examined the possible role of the wind profile in modulating the initial production of vertical vorticity by a downdraft. Our approach was to avoid the problem of simulating different supercells in different environments, and instead to look for direct attribution of differences associated with how the ambient wind profile interacts with a simple downdraft forced by a heat sink. From these experiments, we have attempted to understand the basic sensitivities of surface vorticity production.

Our simple downdraft (absent any updraft forcing) produces credible vertical vorticity values at the lowest model level. As in many previous studies (reviewed in section 1), the primary mechanism in these experiments is the baroclinic generation of horizontal vorticity that is subsequently tilted during descent. In unsheared runs, there is a monotonic increase in surface vorticity production as the SR flow increases from 0 up to 12.5 m s\(^{-1}\), followed by a monotonic decrease for greater values. This sensitivity has to do with the degree of cooling...
acquired by parcels as they pass through the idealized heat sink (net cooling decreases for increasing SR flow), as well as the degree to which horizontal vorticity vectors subsequently attain a component that is normal to the contours of vertical velocity (the normal component increases for increasing SR flow). As implied by the analytical solutions of DJ00, it is clear that vertical wind shear is not necessary for the production of surface vorticity by a downdraft. The only fundamental requirement for downdrafts to produce surface vertical vorticity is the existence of downdraft-relative flow. Perhaps this explains why the vast majority (perhaps all) observed surface-based (i.e., not elevated) supercells have been found to have nonzero vorticity at the ground.

Notably, nonzero downdraft-relative flow is unlikely to occur in an unsheared environment. Thus, there may be an indirect linkage between the vertical wind shear and the appropriate magnitude of downdraft-relative flow. In natural tornadogenesis it is likely that lower-tropospheric vertical wind shear is also important in producing the upward dynamic accelerations needed to lift high vorticity surface outflow parcels back into the parent storm’s updraft, as explained by Markowski and Richardson (2014). Such an effect was not possible in the present downdraft-only study, which only emphasized the primary production of surface vertical vorticity.

Although the addition of shear is not directly helpful to surface vorticity production in these simulations, increased realism is attained in clockwise turning hodographs as the shear magnitude increases. These increasingly realistic elements include asymmetry, with colder outflow and enhanced surface vertical vorticity values (relative to those found in the anticyclonic vorticity center) concentrated on the “updraft facing” (southeastern) side of the outflow. In nature this configuration would presumably be favorable for collocating large surface vorticity in a zone of deep stretching, which is a precondition for tornadogenesis. However, beyond its direct linkages to the low-level SR flow speed, we do not find much other evidence that the initial orientation of ambient vorticity (and possibility for barotropic reorientation thereof) significantly influences the final values of surface vertical vorticity produced by a downdraft.

For some time, researchers have considered the generation of near-surface vorticity to be one of the greatest challenges to understanding tornadogenesis (e.g., Davies-Jones et al. 2001; Markowski and Richardson 2009). In light of recent publications and the present study, a relatively consistent picture emerges. Our study emphasizes that surface vertical vorticity emerges in the outflow primarily due to tilting of horizontal vorticity that was generated baroclinically during the parcels’ descent. We affirm that the only fundamental requirement for this process is nonzero downdraft-relative flow. Such a requirement seems rather easy to satisfy in the moderately and strongly sheared environments that typically support severe storms.

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REFERENCES


