15.1 Near-surface vortexgenesis in idealized three-dimensional numerical simulations involving a heat source and a heat sink in a vertically sheared environment

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1. Introduction

Proximity sounding studies show that tornadoes are increasingly likely as the low-level environmental horizontal vorticity increases (e.g., Rasmussen 2003; Thompson et al. 2003; Markowski et al. 2003a; Craven and Brooks 2004). But dual-Doppler wind retrievals strongly suggest that storm-generated (i.e., baroclinic) vorticity is a major, perhaps even dominant contributor to the vorticity field in rear-flank downdraft and hook echo region of supercells, based on the configuration of vortex lines (Straka et al. 2007; Markowski et al. 2008). One major thread of our ongoing research is to better understand the roles of environmental versus storm-generated vorticity. Why is strong low-level vertical shear—especially when combined with high relative humidity in the boundary layer—so favorable for tornadoes in proximity sounding studies? How do vortex line arches interact with environmental vortex lines? Why does low-level cyclonic vorticity typically become so much larger than anticyclonic vorticity? Why do vortex line arches dominate over sagging, downward-depressed vortex lines?

We have begun tackling these questions by way of idealized simulations designed to emulate what we believe to be some of the salient aspects of the processes by which strong vortices are produced at the surface in supercell thunderstorms. The simulations are dry but include the generation of near-surface rotation beneath supercell-like (i.e., helical) updrafts in a way consistent with our present understanding of the importance of a downdraft in environments in which vertical vorticity is initially absent at the surface. A stationary, cylindrical heat source is imposed within a horizontally homogeneous environmental wind field containing vertical shear. The vertical wind profile is described by a semicircular hodograph; i.e., the horizontal vorticity is purely streamwise given the stationary storm motion. The interaction of the heat source and wind field results in a cyclonically spinning updraft with maximum rotation at midlevels. The rotation vanishes at the surface because there is no environmental vertical vorticity to stretch at the surface, and because the tilting of horizontal vorticity by an updraft alone cannot produce vertical vorticity at the surface. Once a steady state is achieved, a heat sink is imposed on the western flank of the updraft at low levels. The heat sink produces baroclinically generated vortex rings that sink and spread beneath the updraft. Under the right conditions, the interaction between these vortex rings and the overlying updraft can result in the formation of strong surface vortices.[1]

The simulations are similar to Walko's (1993) except that Walko's hodograph was straight and passed through the origin (this environment is not really consistent with a stationary heat source), and it is not clear whether baroclinic vorticity was the source of near-surface cyclonic rotation or whether the near-surface cyclonic rotation developed as a result of environmental horizontal vorticity simply being tilted and advected toward ground by the downdraft [i.e., a barotropic redistribution of vorticity like in the simulations by Davies-Jones (2000, 2008) and Markowski et al. (2003b)]. The simulations also share some similarities with those of Straka et al. (2007), although in their simulations there was no environmental vorticity present.

2. Model specifics

The dry version of the Bryan cloud model is used (Bryan and Fritsch 2002). A fifth-order advection scheme is used, which has implicit diffusion. No additional artificial diffusion is included. There are no surface fluxes, Coriolis force, or radiative transfer. The domain is 100 km $\times$ 100 km $\times$ 18 km, with a rigid top and bottom boundary and open lateral boundaries. The horizontal grid spacing is 100 m within a 25 km $\times$ 25 km region centered in the domain, and gradually increases to 3.5 km from the edge of this inner region to the lateral boundaries via the function given by Wilhelmson and Chen (1982). The vertical grid spacing varies from 100 m in the lowest 1 km to 400 m at the top of the domain. Owing to the horizontal and vertical grid spacing, one cannot expect to resolve tornadoes but one can expect that circulations on the "tornado-cyclone" scale are reasonably well-resolved. The large (small) time step is 1 s (0.1 s). The simulations are run for 1 h.

The environmental wind profile is initialized using the formulation for a semicircular hodograph given by McCaul and Weisman (2001) that has a vertically varying wind shear mag-

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[1] The reader is strongly encouraged to view the recorded oral presentation, which contains numerous color animations. This preprint is intended to provide details about the model design that we will not have time to cover in the oral presentation, and the oral presentation is intended to present information in a way that is not easily duplicated on paper. The preprint and archived oral presentation are best viewed as complementary documents.

nitude, such that

$$\bar{v}(z) = \frac{Anz}{H} e^{1-\frac{nz}{H}} \tag{1}$$

$$\bar{u}(z) = \mathrm{sgn}(z - z_0)\{A^2 - [\bar{v}(z)]^2\}^{\frac{1}{2}}, \tag{2}$$

where $\bar{u}$ and $\bar{v}$ are the zonal and meridional environmental wind components, z is the vertical coordinate, A is the hodograph radius, n is the profile "compression parameter," H is the vertical scale, and z_0 is the height where $\bar{v}(z)$ is a maximum. Our simulations used $A = 8$ m s^{-1}, $H = 6$ km, and n was varied from 3 to 6.

The initial environmental potential temperature field ($\bar{\theta}$) is horizontally homogeneous. At the surface, $\bar{\theta} = 300$ K; $\bar{\theta}$ increases with height at 1 K km^{-1} in the lowest 10 km AGL, and 10 K km^{-1} above 10 km AGL.

A heat source (S_w) and heat sink (S_c) were added to the model's potential temperature tendency equation. The heat source and sink are given by

$$S_\phi = S_{\phi 0}\ R(x, y)\ Z(z), \tag{3}$$

where ϕ is either w or c to indicate a heat source or sink, respectively, $S_{\phi 0}$ is the heat source/sink amplitude, and

$$R(x, y) = \begin{cases} 1 - \frac{r^2}{R_\phi^2}, & r \le R_\phi \\ 0, & \text{otherwise} \end{cases} \tag{4}$$

$$Z(z) = \begin{cases} 1 - \frac{(z - z_\phi)^2}{Z_\phi^2}, & z \le Z_\phi \\ 0, & \text{otherwise} \end{cases} \tag{5}$$

where R_ϕ is the radius of the heat source/sink, Z_ϕ is the half-depth of the heat source/sink, $r^2 = (x - x_\phi)^2 + (y - y_\phi)^2$, and the heat source/sink is centered at (x_ϕ, y_ϕ, z_ϕ).

The heat source has an amplitude of S_{w0} = 0.048 K s^{-1}, is centered at (x_w, y_w, z_w) = (50, 50, 5.25) km, and has dimensions given by $R_w = 3$ km and Z_w = 4.75 km. The heat source is present throughout the simulation. It produces an approximately steady, cyclonically rotating updraft by 900 s, at which time the updraft has a maximum potential temperature excess of 7.2 K, a maximum vertical velocity of 42 m s^{-1}, and a maximum vertical vorticity of 0.017 s^{-1}.

The heat sink is activated at 900 s at low levels and to the west of the heat source in order to emulate an RFD. It is present for the remainder of the simulation. Its zonal position is x_c = 49 km and its meridional position (y_c) is varied from 45–55 km in different experiments; it is centered at the surface (i.e., $z_c =$ 0 km). Its dimensions are given by $R_c = 1$ km and Z_c = 3 km. Its amplitude (S_{c0}) is varied from 0.008–0.064 K s^{-1}.

3. Overview of results

When the heat sink is strong (e.g., S_{c0} = 0.064 K s^{-1}, which results in a large, rapidly advancing cold pool with potential temperature deficits as large as 12 K at the surface), the parcels originating in the heat sink, as well as the vortex lines generated baroclinically within the temperature gradient along the periphery of the heat sink, simply undercut the updraft and fail to be lifted; only weak vertical vorticity arises at the surface in this case. If the heat sink is too weak (e.g., S_{c0} = 0.008 K s^{-1}, which results in maximum surface potential temperature deficits of only 1–2 K), the baroclinic vorticity generation is small and/or parcels that have acquired baroclinic vorticity are unable to spread beneath the updraft from the rear and be lifted by it; only weak vertical vorticity arises at the surface in this case as well. For intermediate heat sink strengths (e.g., S_{c0} = 0.016 K s^{-1}, which results in maximum potential temperature deficits of 2–5 K at the surface), significant baroclinic vorticity is generated, yet parcels originating within the heat sink's outflow are able to be forcibly lifted in spite of their negative buoyancy [the upward-directed, dynamic vertical perturbation pressure gradient force (VPPGF) is sufficiently strong relative to the negative buoyancy in this situation]. Strong surface vortices develop in these simulations, and the wind field kinematically resembles that of a supercell near the time of strong low-level rotation (e.g., an occluded "gust front" structure develops, including an occlusion downdraft, and vortex lines form arches).

The generation of a strong vortex in the simulations can be controlled not only by the strength of the heat sink, but also by the strength of the low-level wind shear. Increasing the shear increases the strength of the low-level updraft because the upward-directed, VPPGF is increased. For a given heat sink strength, as the low-level shear increases, it becomes increasingly likely that the air parcels and vortex lines that originate in the heat sink's outflow will be able to be lifted, thereby increasing the likelihood of the formation of a strong vortex at the surface. We believe these simulations provide a plausible explanation for why tornadic supercells are favored in environments containing large low-level wind shear, in addition to environments that limit cold pool production (e.g., environments that have large boundary layer relative humidity).

In the oral presentation, we will present animations of the low-level wind fields, vortex lines, and trajectories derived from a subset of the simulations. For example, a comparison of a simulation with strong environmental shear at low levels and a weak cold pool, and a simulation with weak environmental shear at low levels and a strong cold pool, reveals that the former develops much stronger, deeper cyclonic vortices at the surface than the latter, consistent with the summary above. We will also discuss some interesting aspects of the early evolution of the vorticity field, shortly after the heat sink is activated. In the early stages of evolution (i.e., when the cold pool is weak), the development of a downdraft on the rear flank of the updraft results in an elevated vorticity couplet joined by sagging (U-shaped) vortex lines. The realism of this evolution is unknown. We don't usually observe storms in storm intercept projects this early in the evolution of storms (usually chasers don't commit to a storm until significant precipitation/outflow is present). Once the cold pool becomes significant, the low-level vertical vorticity field is dominated by vortex lines that have been altered substantially (or generated) by baroclinity, and subsequently lifted. This is similar to the formation of bookend vortices in mesoscale convective systems (Weisman and Davis 1998).

We also found that the symmetry of the amplification of the near-surface vorticity extrema is sensitive to the location of the heat sink relative to the updraft. The asymmetry appears to result from differences in low-level stretching (i.e., which branch of an arching vortex line is beneath the strongest part of the overlying updraft) and differences in how the arching vortex lines interact with the vortex lines in the overlying updraft, which originate in the environment (note that these two factors are probably partly related). For example, in a simulation in which the heat sink is located farther north than in the simulations that produce strong cyclonic vortices at the surface, the dominant vortices at the surface are anticyclonic, although they are not as strong as the strongest cyclonic vortices in the simulations in which the heat sink location favors cyclonic vortices at the surface. Moreover, the vortex line arches never evolve into a deep column of approximately vertical vortex lines, despite the fact that parcels from the cold pool are able to be lifted as they are in the experiments that result in strong cyclonic vortices (in the experiments that produce strong cyclonic vortices, the vortex line arches evolve into approximately vertical vortex lines that extend to the updraft summit). More will be said about these asymmetries in the oral presentation.

4. Tentative conclusions and concluding remarks

The strongest, deepest vortices form when cold pool air and associated baroclinic vorticity can be "processed" by the overlying updraft. This is most likely to occur when the low-level updraft is strong and cold pools are relatively weak (but not too weak lest they cannot overcome the front-to-rear updraft-relative winds at low levels and therefore fail to spread beneath the updraft). The strength of the low-level updraft increases with increasing low-level environmental shear. One interesting question to ponder is whether the role of environmental low-level shear is simply to enhance the low-level updraft.

The symmetry of the amplification of near-surface vorticity extrema is sensitive to the location of the heat sink relative to the updraft. It is tempting to consider whether, in actual storms, a cyclonic midlevel mesocyclone indirectly tends to favor the amplification of the cyclonic member of the near-surface vorticity couplet by guiding precipitation around the updraft so that the "heat sink" finds itself in a favorable location. The heat sink location also would be sensitive to the environmental wind field and microphysical attributes of the storm. We still have an awfully limited conceptual model of the three-dimensional buoyancy field and baroclinic vorticity generation within supercells, and are almost totally ignorant of the buoyancy fields above the ground. The degree to which cold pool vortex lines are "lifted" also seems to depend on interactions between the vertical vorticity of the midlevel mesocyclone and the vertical vorticity (both cyclonic and anticyclonic) of the upward-growing arches. In addition to influencing the VPPGF, the environmental vorticity (both its magnitude and orientation) also affects the vertical vorticity within the overlying updraft, and therefore the potential interaction with baroclinic vorticity that is drawn upward from below.

Regarding future work, we are trying to assess the role of shear instability and how our understanding of shear instability can be merged with the concept of the lifting of baroclinic vortex lines (in all of the simulations—and in more realistic three-dimensional, high-resolution supercell simulations as well—the evolution of the near-surface vorticity field resembles the breakdown of a vortex sheet into vortices). We also would like to use the same model set-up to explore a wider range of environmental wind profiles (e.g., ones with significant crosswise vorticity). How is the lifting of cold pool air affected, and how is the interaction of arching vortex lines with the environmental vortex lines in the overlying updraft affected? This remains a work-in-progress, and as of now we are really just scratching the surface. Even in these "simple" model experiments there is a humbling degree of complexity.

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