The Regulation of Tornado Intensity by Updraft Width

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ABSTRACT

Strong to violent tornadoes cause a disproportionate amount of damage, in part because the width and length of a tornado damage track are correlated to tornado intensity (as now estimated through enhanced Fujita scale ratings). The tendency expressed in the observational record is that the most intense tornadoes are often the widest. Herein the authors explore the simple hypothesis that wide intense tornadoes should form more readily out of wide rotating updrafts. This hypothesis is based on an application of Kelvin’s circulation theorem, which is used to argue that the large circulation associated with a wide intense tornado is more plausibly associated with a wide mesocyclone. Because a mesocyclone is, strictly speaking, a rotating updraft, the mesocyclone width should increase with increasing updraft width. A simple mathematical model that is quantified using observations of mesocyclones supports this hypothesis, as do idealized numerical simulations of supercellular thunderstorms.

1. Introduction

Tornado-occurrence statistics reveal that strong to violent [enhanced Fujita (EF) scale 2–5] tornadoes result in a disproportionate number of fatalities (e.g., Ashley 2007) and cause a disproportionate amount of damage. This is in part because strong to violent tornadoes tend to have relatively longer and wider damage tracks and, thus, implicitly affect larger areas (Brooks 2004). Acknowledging the uncertainties in describing tornado size based on visible and damage characteristics (e.g., Atkins et al. 2014), our interpretation of Brooks (2004) is that a wider tornado is more likely to be strong to violent than is a narrower tornado, though not exclusively so (Wurman and Kosiba 2013). Accordingly, our attention henceforth is on wide intense tornadoes and the characteristics of their parent mesocyclones.

As guided, for example, by the observations presented in Wakimoto et al. (1996), the mesocyclone can be idealized as a columnar vortex that extends from the ground upward through a large fraction of the troposphere. The tornado, which is assumed herein to develop from a contraction of this mesocyclonic parent vortex, can likewise be idealized as a deep columnar vortex. Importantly, such tornado formation requires vertical vorticity near the ground as well as aloft, as is readily shown in axisymmetric model simulations of tornadoes (e.g., Fiedler 1995; Trapp and Davies-Jones 1997; Nolan 2005; Parker 2012).

This framework, the analysis of Brooks (2004), and the correspondence between tornado-vortex core size and updraft-forcing length scale in axisymmetric model simulations by Nolan (2005), motivate a simple hypothesis: Wide intense tornadoes should form more readily out of wide rotating updrafts. This is based on application of Kelvin’s circulation theorem, which is represented as

\[ 2\pi v_T r_T = \Gamma = 2\pi v_M r_M, \]

where \( r_T \) and \( v_T \) (\( r_M \) and \( v_M \) are, respectively, the radius and tangential wind speed of the tornado (mesocyclone), and \( \Gamma \) is circulation. We do not purport to use (1) as a means to predict the ultimate radius (and intensity) to which the mesocyclone is contracted, because this will depend on time-dependent characteristics of the host storm and its 3D environment, as well as nonlinear interactions between the tornadic vortex and boundary layer (e.g., Rotunno 2013). Rather, we use (1), and its conservation of angular momentum equivalence, to argue that the large circulation associated with a wide intense tornado is more plausibly associated with a wide mesocyclone. We then extend this argument to updraft width because it allows us to introduce physical processes relevant to the mesocyclone and thus tornado formation.
Indeed, as demonstrated in the next section, a wide mesocyclone necessarily—and unsurprisingly—forms out of a wide updraft.

2. Theoretical considerations

Starting with a simple model that governs vertical-vorticity generation through the tilting of streamwise horizontal vorticity, we attempt in this section to 1) show that updraft width is relevant in the generation of large circulation and then 2) illustrate using observations how the large circulation associated with a wide intense tornado is more plausibly associated with a wide mesocyclone.

The model is developed by first considering a super-cell thunderstorm during an early stage of its evolution. At such a stage, the storm’s vertical motion field comprises mostly rising air over a large depth of the troposphere. Initial formation of mesocyclical rotation at midlevels (nominally, ~3–7 km above the ground) also occurs at this stage.

Midlevel mesocyclogenesis is modeled well by an equation governing the vertical component of vorticity, linearized about a base state with a vertically dependent zonal wind \( U(z) \):

\[
\frac{\partial \zeta'}{\partial t} + (U - c_x) \frac{\partial \zeta'}{\partial x} + (-c_y) \frac{\partial \zeta'}{\partial y} = \frac{dU}{dz} \frac{\partial w'}{\partial y},
\]

(2)

where primes represent deviations from the base state, \( \zeta' \) is the deviation vertical vorticity, \( V' = (u', v', w') \) is the deviation velocity vector, \( c = (c_x, c_y) \) is the storm motion vector, and all other variables have their traditional meanings (e.g., Trapp 2013, 201–213). Absent from (2) is the nonlinear process of vortex stretching: because stretching has a nonzero contribution to the vertical vorticity tendency only after vertical vorticity itself is nonzero, its neglect in our examination of mesocyclogenesis is acceptable. Retained in (2) is the process of tilting, expressed in terms of the vertical gradient of the base-state wind (herein, environmental shear) and the storm’s vertical motion field. And implicit in (2) is a consideration of the problem in a frame of reference fixed to the moving storm.

Following Lilly (1986), we assume steadiness in this reference frame, and then evaluate (2) at the height (critical level) in which \( U - c_x = 0 \). The reduced equation can be integrated to yield

\[
\zeta'(x, t) = \left( \frac{dU/dz}{-c_y} \right) w'(x, t).
\]

(3)

The physical interpretation of (3) is that tilting of environmental horizontal vorticity is exactly balanced by storm-relative advection of vertical vorticity (see also Klemp 1987). The result of this balance is a perfect spatial correlation between the vertical velocity and vertical vorticity fields; in an environment of westerly shear \( (dU/dz > 0) \) and with a “right moving” propagation such that \( c_y < 0 \), the balance yields a cyclonically rotating updraft. Given the reasonable assumptions that both the storm motion vector and environmental wind shear are uniform horizontally across the storm, we see that \( w' \) is the only variable in (3) that contributes to the spatial structure of \( \zeta' \); this contribution can also be deduced from (2) alone. The important implication is that the radius of the updraft determines the radius of the midlevel mesocyclonic vortex.

To illustrate this size relationship between updraft and midlevel mesocyclone, we use a simple top-hat profile to represent the updraft core:

\[
w'(r) = \begin{cases} W, & r \leq R \\ 0, & r > R \end{cases}
\]

(4)

where \( r \) is the radius from the centerline of the updraft core, \( W \) is maximum updraft-core speed, and \( R \) is the radial extent of the updraft core. It is trivial to show using (3) that the corresponding solution for \( \zeta' \) is

\[
\zeta'(r) = \begin{cases} Z, & r \leq R \\ 0, & r > R \end{cases}
\]

(5)

where \( Z = (W/|c_y|)dU/dz \) is the maximum mesocyclone core intensity. Equation (5) represents a core of constant vertical vorticity \( Z \) surrounded by irrotational flow and, thus, describes a Rankine vortex profile in tangential velocity with a radius of maximum winds at \( r = R \). Accordingly, (5) shows that radial extent \( R \) of the updraft core is also the radial extent of the mesocyclone core; note that other updraft-core profiles, such as one that is jetlike, yield similar relationships, and thus further reinforces this point.

The circulation \( \Gamma \) associated with the mesocyclonic vortex can be expressed as

\[
\Gamma = \int_0^{2\pi} \int_0^R \zeta' r dr d\theta,
\]

(6)

which makes use of Stokes theorem and assumes integration over a circular area with radius \( R \); such area is considered here to be a cross section through a cylindrical rotating updraft. After substitution of \( \zeta' \) from (5) into (6), and then integration of the result, we have

\[
\Gamma = \pi Z R^2;
\]

(7)
under the inviscid and barotropic flow assumptions required by Kelvin’s circulation theorem, fluid parcels will conserve this circulation. Given the quadratic dependence of $\Gamma$ on $R$, (7) illustrates the relevance of updraft width on the generation of circulation and, thus, satisfies our first objective of this section.

The essence of the preceding analysis, including (2), applies to mesocyclonic rotation at cloud base. However, it is appropriate to ask whether updraft width connects in any way to a “near ground” mesocyclone that owes its existence to the tilting of horizontal baroclinic vorticity within downdrafts (e.g., Davies-Jones et al. 2001). As supported by numerical simulations presented in section 3, our hierarchical answer begins with the argument that the horizontal length scale of the primary downdraft should relate to the horizontal length scale of the primary updraft. This is because the area of precipitation that drives this downdraft is inherited from the updraft in which the precipitation is formed; observations in a range of convective storms show a parity in updraft and downdraft size (as well as intensity; e.g., Lucas et al. 1994). The length scale of the near-ground mesocyclone is thereafter imparted by the length scale of this downdraft, wherein vertical vorticity is generated via tilting of horizontal streamwise baroclinic vorticity; because the baroclinity is derived from the characteristics of the downdraft and associated outflow, even the magnitude of the vertical vorticity should depend in some way on the downdraft and thus updraft. Finally, the convergent low-level inflow, which contracts the mesocyclone, has a scale that is imposed by the scale of the updraft (see also Nolan 2005). For example, in a Ward-type tornado simulator (Ward 1972), the inflow begins to turn into the vertical within some radius $r_0$ of the updraft-core centerline (e.g., Fig. 1 in Davies-Jones 1973). It is within this “convergent area” that a tornadic vortex forms; it follows based on the circulation conservation arguments made herein that the diameter of the vortex is necessarily limited by $r_0$ and the associated convergent area. We note that the physical analog of $r_0$ is updraft width.

By this reasoning, we conclude that the radial extent of the near-ground mesocyclone is at least constrained by the radial extent $R$ of the updraft, even though some of the details of the processes inherent in (5) and thus (7) are different at low levels (see also Rotunno et al. 2017).

The second objective of this section can therefore now be pursued, beginning with a quantification of (7) to show that its inherent processes equate to a reasonable value of circulation for a tornadic mesocyclone. For this purpose, we are guided by the Musil et al. (1986) measurements and let $R = 3$ km and $W = 20$ m s$^{-1}$, and then use somewhat arbitrary yet still relevant supercell values of $c_p = -5$ m s$^{-1}$ and $dU/dz = 5 \times 10^{-3}$ s$^{-1}$. These yield a $\Gamma$ value of $5.6 \times 10^5$ m$^2$ s$^{-1}$, which is considered to be large but not unprecedented circulation about a mature mesocyclone (Wakimoto et al. 2004). Among the possible combinations of tornado characteristics that [through (1)] equate to this circulation is $r_T = 800$ m and $u_T = 112$ m s$^{-1}$ (i.e., a “mile wide” tornado with EF5 wind speeds), such as observed with the Mulhall, Oklahoma, tornado on 3 May 1999 (Wurman et al. 2007).

Working in reverse, we seek the characteristics of a mature mesocyclone that can explain this particular circulation,

$$2\pi v_M r_M = \Gamma = 5.6 \times 10^5 \text{ m}^2 \text{ s}^{-1}. \quad (8)$$

For example, letting $r_M = R = 3$ km in (8) gives $v_M = 30$ m s$^{-1}$. To provide some perspective of these and other values, we consult frequency distributions of tornadic-mesocyclone radius and rotational velocity (Figs. 1a,b), extracted from the mesocyclone dataset described by Trapp et al. (2005). The data in Figs. 1a and 1b represent the peak, low-level attributes of 219 tornadic mesocyclones that were detected at distances from 10 to 225 km using the mesocyclone detection algorithm (Stumpf et al. 1998); the mesocyclones were associated with tornadoes that spanned the full range of the Fujita scale. Values of $r_M = 3$ km and $u_M = 30$ m s$^{-1}$ exceed both the median and mean of their respective distributions ($2.75$ km, $19.5$ m s$^{-1}$; $2.85$ km, $21.0$ m s$^{-1}$); $v_M = 30$ m s$^{-1}$ is also within the range of (presumed) mesocyclonic rotational velocities associated with EF5 tornadoes (Kingfield and LaDue 2015). A smaller radius of $R = 1.5$ km in (8) requires $v_M = 59$ m s$^{-1}$, which is in the upper tail of our tornadic-mesocyclone distribution (Fig. 1b) but well outside the distribution of nontornadic mesocyclones derived from the same mesocyclone dataset of Trapp et al. (2005) (Fig. 1c). Thus, although 59 m s$^{-1}$ is perhaps a plausible rotational velocity of a preternadic mesocyclone, it is equally likely that such a large $v_M$ actually reflects the intensity of the tornado it hosts (e.g., Thot et al. 2013). Further decreases in $r_M$ require implausibly high mesocyclonic rotational velocities and also approach the $r_T$ value of large tornadoes. On the other hand, an increase in $r_M$ to 5 km, which is in the upper tail of the mesocyclone size distribution

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1 We equate this to recent usages of “low-level mesocyclone” (e.g., Coffer and Parker 2017).

2 We consider the lowest few hundred meters above the ground to qualify as near ground (e.g., Trapp et al. 2005; but also see Rotunno et al. 2017).
(see also Bellon and Zawadzki 2003; Wakimoto et al. 2004), requires $v_M = 18 \text{ m s}^{-1}$, which is in the lower tail of the tornadic (and nontornadic) mesocyclone distribution, yet both are quite plausible.

This exercise suggests that a large-$r_M$, small-$y_M$ product is more likely to explain a large value of mesocyclonic circulation than does a small-$r_M$, large-$y_M$ product. Circulations even larger than that in (8) would bolster this argument, which is based on the range of observed mesocyclone characteristics documented here and in the literature. However, we do acknowledge the existence of a range of intermediate values of $r_M$ and $y_M$ that can also be combined to yield large circulation. Thus, in section 3 we seek more definitive support of our hypothesis using numerical simulations of supercellular thunderstorms.

### 3. Supporting numerical model simulations

The simplifications made in the preceding sections include the neglect of baroclinity, friction, and turbulent mixing as well as of the time-dependent characteristics of the host storm and its environment. They also include the assumption of conservation of circulation itself, which has uncertainty in real tornado–mesocyclone systems (e.g., Wakimoto et al. 2003; Atkins et al. 2012). In this section we attempt to remove these simplifications by using the full nonlinear equations within the context of idealized numerical simulations of supercellular thunderstorms.

The simulations are conducted with CM1 (version 18), a cloud-resolving model described in Bryan and Morrison (2012). Details of the model configuration are provided in Table 1. In brief, the homogenous initial and boundary conditions are characterized by thermodynamic and wind
profiles (Fig. 2) that are known to promote tornadic supercellular structure in dominant right-moving (RM) storms (see Fig. 3). Defining the thermodynamic profile (Fig. 2a) are (mixed layer) lifting condensation level (LCL) and convective available potential energy (CAPE) values of 1000 m and 1500 J kg\(^{-1}\), respectively; these are approximately equal to the median LCL and CAPE values in spring-season (March–May) tornado environments within the Great Plains region of the United States (e.g., Thompson et al. 2003; Grams et al. 2012). As described below, the wind profile is characterized by a quarter-circle hodograph (Fig. 2b) with a radius that is varied to result in a vertical wind shear range observed in association with tornadic RM supercells.

Convection is initiated in the 3D model domain with a “warm bubble” as defined in Table 1 (see also Weisman and Rotunno 2000), and the subsequent storm(s) is simulated over a period of 4 h, which allows it to reach a peak intensity. A constant horizontal motion vector (8 m s\(^{-1}\), 0 m s\(^{-1}\)) is subtracted from the wind profile to allow the convective storms to remain in the model domain over this time.

Because the purpose of these simulations is to examine the hypothesized updraft width–tornado intensity relationship, the model experimentation involves only the environmental vertical wind shear, which is the initial/boundary condition that exerts the largest influence on supercellular-updraft width (Kirkpatrick et al. 2009). Adapting the methodology of Kirkpatrick et al. (2009), we vary the radius of the quarter-circle hodograph over the 0–2-km layer from 6 to 10 m s\(^{-1}\) (U0), in increments of 0.5 m s\(^{-1}\). With each increment, the length of the hodograph curve is increased over the 2–6-km layer to preserve the hodograph shape; the winds above 6 km are held constant at the 6-km value (see Fig. 2b). The 0–6-km bulk wind shear corresponding to these hodograph variations ranges from 21 to 36 m s\(^{-1}\), which span the 10th–90th-percentile values of bulk shear in spring-season (March–May) tornadic environments analyzed by Grams et al. (2012); considering the results of Thompson et al. (2003), a similar comment can be made for the 0–3-km storm-relative helicity (SRH) (as computed here using actual storm motion), which ranges from 115 to 450 m\(^2\) s\(^{-2}\).

As can be inferred from Fig. 3, the midlevel (\(z = 6.25 \text{ km}\)) updraft core area is highly sensitive to U0: for the particular thermodynamic profile shown in Fig. 2a, the core area exhibits a 69.75-km\(^2\) (or 319\%) increase over the experimental range of U0 (Fig. 4). Note that area is used here instead of width, because the updrafts are not exactly circular as assumed in the preceding theoretical analysis. Following Musil et al. (1986),
updraft core area is quantified by the number of contiguous grid points within the storm that exceed a vertical velocity threshold of $20 \text{ m s}^{-1}$ at a height of 6.25 km. Three related areas are also quantified: downdraft core area, midlevel mesocyclone area, and near-ground mesocyclone area. A vertical velocity threshold of $-8 \text{ m s}^{-1}$ at a height of 3.25 km is used to describe downdraft core area; a vertical vorticity threshold $10^{-2} \text{ s}^{-1}$ at a height of 6.25 km and a height 250 m is used to describe the midlevel and near-ground mesocyclone areas, respectively. These admittedly subjective thresholds on vertical velocity and vertical vorticity are based on observed (e.g., Brandes 1978) and simulated (e.g., Nowotarski et al. 2011) tornadic supercells. Finally, we also quantify the near-ground vertical-vorticity maximum at a height of 250 m. This is used as a proxy for tornadic-vortex intensity, with the full recognition that a tornadic vortex is not well resolved on a model grid with 500-m spacings. The analysis is valid over a time interval that begins immediately after initial storm splitting and ends 5 min prior to the time of peak near-ground vertical vorticity.

Figure 5 reveals robust relationships between midlevel ($z = 6.25 \text{ km}$) updraft core area and midlevel mesocyclone area (Fig. 5a), midlevel mesocyclone area and near-ground ($z = 0.25 \text{ km}$) mesocyclone area (Fig. 5b), and then midlevel updraft area and near-ground vertical vorticity (Fig. 5c). We note in Fig. 5a that the correspondence between updraft area and mesocyclone area is not one to one, as would only be expected in the special case of an environment with a fully circular hodograph (e.g., Weisman and Rotunno 2000 and references therein). Figure 5b supports the conclusion made

![Graph showing relationship between updraft area and hodograph radius](image)

**Fig. 4.** Summary of CM1 experimentation: scatterplot showing sensitivity of peak midlevel ($z = 6.25 \text{ km}$) updraft area ($\text{km}^2$) to hodograph radius ($\text{m s}^{-1}$).

![Image showing simulated radar reflectivity factor](image)

**Fig. 3.** Plan view of simulated radar reflectivity factor (color fill; dBZ) at $z = 1.25 \text{ km}$, overlaid by the $20 \text{ m s}^{-1}$ vertical velocity contour at $z = 6.25 \text{ km}$ (solid) and the $8 \text{ m s}^{-1}$ vertical velocity contour at $z = 3.25 \text{ km}$ (dashed), for the CM1 experiment with hodograph radius of (a) 6, (b) 8, and (c) 10 m s$^{-1}$. Views show model data at $t =$ (a), (b) 2 and (c) 3 h. Only a portion of the total domain is shown.

3 The value of near-ground mesocyclone area is evaluated at the time of peak mesocyclone area at cloud base ($z = 1.25 \text{ km}$).
in section 2 that the scale of the near-ground mesocyclone should reflect the scale of the updraft and, in turn, the scale of the midlevel mesocyclone. This was based in part on the reasoning that the downdraft, which imparts its scale to the near-ground mesocyclone, should have a comparable length scale as the primary updraft; this is confirmed (in terms of area) by Fig. 5d.

We claim that Fig. 5 supports the hypothesized relationship between near-ground vortex intensity (via maximum vertical vorticity) and updraft core width (via updraft area). This relationship is expected to be relevant for tornadogenesis arising from a contraction of deep mesocyclonic-scale vertical vorticity; in cases of tornadogenesis arising from vertical vorticity not considered to be mesocyclonic (e.g., Dahl et al. 2014; Coffer and Parker 2017; Orf et al. 2017), the role of updraft core width is less clear. This relationship will be constrained in part by the limits imposed on the updraft size by buoyancy pressure forcing (Yuter and Houze 1995; see also Morrison 2016) and will be modulated by the details of the thermodynamic environment (e.g., the LCL and CAPE) as well as of the environmental shear. For example, environmental shear \( \frac{dU}{dz} \) directly affects the magnitude of midlevel mesocyclonic vertical vorticity \( \zeta' \), as predicted by Eqs. (2) and (3), and also indirectly affects the spatial distribution of \( w' \) and thus \( \zeta' \), as demonstrated in Fig. 4 (and as can be explained through a consideration of linear and nonlinear dynamic pressure forcing; G. R. Marion et al. 2017, in preparation; see also Marion 2017). Accordingly, we are mindful of the interdependencies that arise as we attempt to extend our simple ideas to complex situations.

It is appropriate to note that the environmental shear–tornado pathway described here provides a means to reconcile the observed relationship between SRH and tornado damage (EF scale), especially for an SRH that is maximized over low levels (e.g., Thompson et al. 2012). Indeed, it provides an alternative, albeit dynamically consistent explanation to that of Markowski and Richardson (2014), who argue that the role of low-level

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**Fig. 5.** Supercell thunderstorm characteristics from CM1 experimentation over range of hodograph radii (m s\(^{-1}\)): scatterplot showing relationships between (a) midlevel \((z = 6.25 \text{ km})\) updraft area (km\(^2\)) and mid-level mesocyclone area (km\(^2\)), (b) mid-level mesocyclone area (km\(^2\)) and near-ground \((z = 0.25 \text{ km})\) mesocyclone area (km\(^2\)), (c) mid-level updraft area (km\(^2\)) and near-ground vertical vorticity (s\(^{-1}\)), and (d) mid-level updraft area (km\(^2\)) and downdraft area (km\(^2\)) at \(z = 3.25 \text{ km}\). The values of updraft, downdraft, and mid-level mesocyclone areas represent peaks over an evaluation period between the time of storm split and the time of maximum near-ground vertical vorticity; the value of near-ground mesocyclone area is evaluated at the time of peak mesocyclone area at cloud base \((z = 1.25 \text{ km})\).
environmental shear is effectively to lower the base of the midlevel mesocyclone, which then dynamically enhances low-level vertical accelerations.

Apropos to this mention of enhanced vertical accelerations, one of the reviewers questioned whether updraft intensity, rather than width, was the more relevant characteristic here. Our immediate response to this question is that updraft intensity alone does not explain the relatively large width of many intense tornadoes: per our note in section 2, tornado width is necessarily limited by updraft width. However, we do acknowledge the theoretical basis (e.g., Fiedler and Rotunno 1986) and model demonstration (Nolan et al. 2017) of a scaling between the maximum tangential winds of the tornado and the maximum buoyantly forced updraft speed \( (-\sqrt{2} \text{CAPE}) \); a specific form of this scaling is known as the “thermodynamic speed limit,” which is readily exceeded in certain tornado morphologies (e.g., Fiedler and Rotunno 1986; Fiedler 1994). With this in mind, our model experiments can be used to explore this question further. In Fig. 6a, we find that the simulated supercells have peak updraft intensities (prior to the time of peak near-ground vertical vorticity) that exhibit relatively little sensitivity to \( U_0 \). The most intense updrafts here are generally the widest updrafts, but as can be deduced from Figs. 4 and 6a, the linear relationship between these two characteristics is not particularly strong \( (R^2 = 0.31) \). An upper limit on intensity is presumably imposed by the environmental thermodynamics: the simulated intensities approach the value of \( 55 \text{ m s}^{-1} \) \( (-\sqrt{2} \text{CAPE}) \), especially as the updraft area increases and thus the dilution of updraft buoyancy by entrainment decreases. In light of the fact that peak near-ground vertical vorticity relates strongly to \( U_0 \) (Fig. 5), it should not be surprising from Fig. 6a that the model simulations exhibit a weak relationship between peak updraft intensity and peak near-ground vertical vorticity over the range of \( U_0 \) (Fig. 6b).

Because the model experiments were conducted with relatively coarse grid lengths and a free-slip lower boundary condition, the results in Fig. 6b are not related to the end-wall vortex dynamics invoked by Fiedler and Rotunno (1986) but, rather, to convective-storm dynamics as manifest by the \( U_0 \)-dependent increases in mesocyclocnic circulation available for contraction. Our planned future experiments with varying thermodynamic profiles will shed further light on CAPE dependencies, but in the meanwhile we can consider the observations of CAPE in environments of tornadic storms. As revealed by the analysis of Smith et al. (2015), CAPE is not a good discriminator of EF scale [e.g., see Fig. 9 in Smith et al. (2015)]. Other studies in which CAPE is evaluated across convective modes, hazard types, and severity have reached similar conclusions (e.g., Thompson et al. 2012; Grams et al. 2012; references therein). It is noteworthy that even for hail size, CAPE fails as a robust discriminator (Johnson and Sugden 2014). Vertical wind shear and attendant updraft area, on the other hand, have been shown recently by Dennis and Kumjian (2017) to be key controls on hail growth in simulated supercells. Similarly, new research by Warren et al. (2017) relates precipitation intensity and supercell morphology to updraft area, as influenced by upper-layer wind shear. Both of these findings corroborate our emphasis on updraft area.

4. An application

This research was originally motivated by a challenge to rethink how operational data might be used to anticipate high-impact tornadoes. Although we and others have demonstrated the utility of operational Doppler radar data for estimation of potential tornado intensity (Toth et al. 2013; Kingfield and LaDue 2015; Smith et al. 2015), the ideas espoused herein also suggest the utility of satellite remote sensing for this purpose.

In Fig. 7 we show GOES-West infrared brightness temperature (BT) from the 31 May 2013 El Reno, Oklahoma, tornadic storm. This storm yielded the widest documented tornado to date \( (4.2 \text{-km damage-path width}) \) and, although associated with a damage rating of EF3, the tornado had (mobile radar) measured tornadic wind speeds of \( 130–150 \text{ m s}^{-1} \), which is well within the range of EF5 (Wurman et al. 2014; Bluestein et al. 2015; Wakimoto et al. 2016). For comparison, we show BT from the 21 May 2011 tornadic storm event near Ada, Oklahoma; the three tornadoes associated with this storm were rated EF0–EF1 and had damage-path widths of at most \( 800 \text{ m} \) \( \text{SPC} \ 2017 \). We suggest that the area encompassed by satellite overshooting tops (OTs; e.g., Bedka et al. 2010) likely relates to midlevel updraft area and thus could provide information about the potential tornado intensity. In addition to a significantly wider damage path and apparent intensity, the El Reno storm appears to have a significantly larger OT area (especially as defined by the BT minimum near \( 35.5^\circ \text{N} \), \( \text{98}^\circ \text{W} \)) relative to the effectively nonexistent OT area with the Ada storm (Fig. 7).

The veracity of an OT area–tornado intensity relationship depends on how well OT area can be defined and then related to updraft area. Our ongoing work with OT-area quantification is making successful use of advanced image-processing techniques. Some evidence of a relationship between OT area and updraft area is found in the polarimetric radar analyses presented by Homeyer and Kumjian (2015). The supercell
simulations presented in section 3 also provide us a ready means to quantitatively evaluate such a relationship. For reference, Fig. 8 shows a representative plan view of simulated radar reflectivity, and the corresponding cloud-top temperature, just prior to the time this particular simulation exhibited its peak near-ground vertical vorticity. “Cloud top” is determined as the highest level in each grid column that total condensate mixing ratio exceeds 0.1 g kg\(^{-1}\). Contiguous grid points with cloud-top temperature less than a threshold of 210 K (which is \(\sim 5\) K less than the tropopause temperature; see Fig. 2a) define the OT area. The peak in the OT-area time series, as modified by a five-point running mean to account for gravity wave transience, defines the peak OT area for each experiment. Figure 9 reveals a robust, nearly one-to-one relationship between peak OT area and peak (midlevel) updraft area, thus implying through Fig. 5c a strong relationship between peak OT area and peak near-ground vertical vorticity.

A comprehensive analysis will be required to fully establish the veracity of such a relationship. The availability of GOES-16 data with its higher spatial and temporal resolution will help facilitate this analysis, as will other numerical simulations. In addition, dual-polarization radar measurements of the area of elevated specific differential phase (\(K_{DP}\)) regions below 0°C could also yield information about updraft width (van Lier-Walqui et al. 2016).

5. Discussion and concluding remarks

The 31 May 2013 El Reno, Oklahoma, tornado highlighted in section 4 was not only wide and intense, but at times also possessed subsidiary or “suction” vortices (Wurman et al. 2014; Bluestein et al. 2015; Wakimoto et al. 2016). Subsidiary vortices can cause
spots or short swaths of enhanced damage embedded in a broader damage track (e.g., Wakimoto et al. 2016, their Fig. 1). Anecdotal evidence from mobile radar data suggests that not all intense, large-core tornadoes have subsidiary vortices (K. Kosiba and J. Wurman 2017, personal communication). Nevertheless, it is appropriate to ask whether the updraft-width hypothesis—which is devoid of the dynamics that specifically address the formation of such multivortex tornadoes (e.g., Rotunno 2013)—is at least consistent with their existence (and nonexistence).

Consider that the experimentally (e.g., Ward 1972; Davies-Jones 1973; Baker and Church 1979; Church et al. 1979) and numerically simulated (e.g., Rotunno 1979; Fiedler 1995; Lewellen et al. 2000; Nolan 2005) range of tornado structural behavior, including the metamorphosis of a single-celled vortex to a vortex with a two-celled structure and subsidiary vortices, depends on an imposed swirl ratio $S$ as well as on the Reynolds number ($\text{Re}$) and associated boundary layer processes (Rotunno 2013). Implicit in this behavioral range is an increase in tornadic-vortex core radius and intensity with an increase in $S$ (and a modulation by $\text{Re}$).

As defined in terms of a Ward-type tornado-vortex simulation chamber (Ward 1972), $S$ can be written as

$$S = \frac{r_0 \Gamma}{2Q}, \quad (9)$$

where $\Gamma$ represents the circulation imposed on inflowing air at the outer radius of the chamber, $Q$ is the volume flow rate through the chamber, and $r_0$ is the radius of the zone of convection (Davies-Jones 1973). The analogous parameter in a Fiedler-type numerical simulation (Fiedler 1994) is

$$S = \frac{l_r \Omega}{W_{\text{conv}}}, \quad (10)$$

where $l_r$ is the half-width of the buoyancy-like forcing of vertical accelerations, $\Omega$ is the constant ambient vertical vorticity, and $W_{\text{conv}}$ is the vertical velocity scale based on the buoyancy forcing (e.g., Nolan et al. 2017). It is noteworthy that in these and related forms of swirl ratio, a representation of updraft width appears in the numerator [i.e., $r_0$ in (9) and $l_r$ in (10)].

A typical experimental means of increasing $S$ en route to the emergence of subsidiary vortices within a broadened, two-celled vortex is to increase $\Gamma$ or $\Omega$. 

![Fig. 8. Plan view of (a) simulated radar reflectivity (dBZ) at $z = 1.25 \text{ km}$ and (b) cloud-top temperature (K), for the CM1 experiment in which $U_0 = 10 \text{ m s}^{-1}$. The black contours indicate vertical velocity equal to $20 \text{ m s}^{-1}$ at $z = 6.25 \text{ km}$. Both panels show model data at $t = 3 \text{ h}$. Only a portion of the total domain is shown in (b).](image-url)

![Fig. 9. As in Fig. 5, but for a scatterplot showing the relationship between peak overshooting-top area (km$^2$) (see text) and peak midlevel ($z = 6.25 \text{ km}$) updraft area (km$^2$).](image-url)
Thus, as recently demonstrated by Nolan et al. (2017), a range of tornado-like vortex sizes and intensities can obtained with a fixed updraft width ($r_0$ or $l_r$). However, increases in $S$ can also be achieved by progressively increasing $r_0$ or $l_r$ and fixing the other variables in (9) or (10). Based on laboratory study results, in which the tornado-vortex core $r_c$ dependency on $S$ is presented as $r_c/r_0$ (e.g., Baker and Church 1979, their Fig. 2), we expect that the actual $r_c$ for a given $S$ in these $r_0$ or $l_r$ experiments to be larger than $r_c$ in $\Gamma$ or $\Omega$ experiments. Indeed, this is precisely what is shown in Figs. 3a and 3b of Nolan (2005), wherein for a comparable $S$ (and Re), the core radius (and tangential velocity) is larger when $l_r$ is increased than when $\Omega$ is increased. We use this to argue that an intense single-celled vortex with a relatively wide core is a reasonable outcome, just as is a relatively narrow-core two-celled vortex with subsidiary vortices. These two outcomes depend, respectively, on whether $S$ is primarily controlled by $r_0$ (or $l_r$) or by $\Gamma$ (or $\Omega$) and suggest that the widest tornadoic vortices, in an absolute sense and regardless of morphology, arise when updraft width is large with all else being the same.

In a tornado chamber, it is straightforward to isolate the response of one variable like updraft width or ambient vorticity. In numerically simulated supercells or in the real atmosphere, this task is less straightforward because of how the convective components like the updraft and downdraft are coupled both internally and to the environment. Indeed, as noted in section 3, environmental shear affects not only updraft and thus mesocyclone width but also mesocyclone intensity; an analogous statement can be made about the effects of the environmental thermodynamics. With this acknowledgment of the complexity of the general problem of identifying controls on tornado intensity, we end here with the caveat that our basic conclusion regarding updraft width to intensity.

To help us further understand these complexities and the need for this caveat, we intend to consider a broader set of thermodynamic and wind profiles within idealized and nonidealized simulations run at higher resolution. We will also continue our analyses of updraft width itself, which involves the forcing due to buoyancy and dynamic pressure, to learn more about the environmental controls on width as well as how these are coupled to the downdraft and cold pool.

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