# The Relationship between Spatial Variations in the Structure of Convective Bursts and Tropical Cyclone Intensification as Determined by Airborne Doppler Radar

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#### ABSTRACT

The relationship between radial and azimuthal variations in the composite characteristics of convective bursts (CBs), that is, regions of the most intense upward motion in tropical cyclones (TCs), and TC intensity change is examined using NOAA P-3 tail Doppler radar. Aircraft passes collected over a 13-yr period are examined in a coordinate system rotated relative to the deep-layer vertical wind shear vector and normalized by the low-level radius of maximum winds (RMW). The characteristics of CBs are investigated to determine how the radial and azimuthal variations of their structures are related to hurricane intensity change. In general, CBs have elevated reflectivity just below the updraft axis, enhanced tangential wind below and radially outward of the updraft, enhanced vorticity near the updraft, and divergent radial flow at the top of the updraft. When examining CB structure by shear-relative quadrant, the downshear-right (upshear left) region has updrafts at the lowest (highest) altitudes and weakest (strongest) magnitudes. When further stratifying by intensity change, the greatest differences are seen upshear. Intensifying storms have updrafts on the upshear side at a higher altitude and stronger magnitude than steady-state storms. This distribution provides a greater projection of diabatic heating onto the azimuthal mean, resulting in a more efficient vortex spinup. For variations based on radial location, CBs located inside the RMW show stronger updrafts at a higher altitude for intensifying storms. Stronger and deeper updrafts inside the RMW can spin up the vortex through greater angular momentum convergence and a more efficient vortex response to the diabatic heating.

#### 1. Introduction

Improving tropical cyclone (TC) intensity forecasts continues to be a challenging yet critical area of research (i.e., Rogers et al. 2006; DeMaria et al. 2005). Of particular concern is understanding the conditions that favor TC intensification, including rapid intensification [RI; defined as an increase of 30 kt (1 kt =  $0.5144 \text{ m s}^{-1}$ ) in 10-m peak winds over a 24h period; Kaplan and DeMaria 2003]. This area of research requires a consideration of a vast spectrum of spatial and temporal scales to understand a TC's dynamical and thermodynamical structure and evolution during its life cycle (Marks and Shay 1998).

Precipitation, and convection in particular, has received extensive coverage in the literature for its potential role in TC intensity change due to its role in heating the inner core of the vortex (e.g., Malkus and Riehl 1960; Gentry et al. 1970; Heymsfield et al. 2001; Kelley et al. 2004; Tao and Jiang 2015). Intensification as a result of convection has been hypothesized to be governed by a variety of processes. One set of studies has emphasized the response of the vortex to diabatic heating from convection via gradient adjustment, with a more efficient vortex response occurring when diabatic heating occurs in the higher inertial stability environment inside the radius of maximum wind (RMW; e.g., Shapiro and Willoughby 1982; Schubert and Hack 1982; Nolan et al. 2007; Vigh and Schubert 2009; Pendergrass and Willoughby 2009). Another set of studies describes low-level intensification more directly in terms of the radial advection of absolute angular momentum relative to the diabatic heat source. In the planetary boundary layer (PBL), enhanced spinup of the vortex winds occurs as parcels move inward to smaller radii at a

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rate faster than the rate of angular momentum loss through the frictional torque (Smith et al. 2009; Montgomery et al. 2014; Sanger et al. 2014; Smith and Montgomery 2016). It should be noted that each mechanism considers a radial distribution of diabatic heating that is peaked inside the RMW as more favorable for intensification.

Other studies have proposed that deep convection, termed convective bursts (CBs), occurring near the RMW can cause intensification by producing peripheral subsidence warming at high altitudes (i.e., above 12 km) that extends into the eye (Heymsfield et al. 2001; Guimond et al. 2010; Zhang and Chen 2012; Chen and Zhang 2013; Guimond et al. 2016). All of the aforementioned dynamical responses to CBs are in conjunction with the positive effect of diabatic heating on the vortex.

In addition to processes tied to the kinematic fields, other research has focused on the role of thermodynamic modification of the PBL in facilitating TC intensity change. Convective downdrafts outside the RMW transport low equivalent potential temperature ( $\theta_e$ ) air to the PBL, reducing the local buoyancy of inflowing air and modifying both the radial and azimuthal distribution of deep convection (Riemer et al. 2010; Molinari et al. 2013; Zhang et al. 2013; Zawislak et al. 2016; Rogers et al. 2016; Nguyen et al. 2017). Such a flushing of the boundary layer can inhibit convection and TC intensification unless the PBL can recover via surface enthalpy fluxes, turbulent mixing, or advection of high- $\theta_e$  air.

The azimuthal distribution of convection has also received attention for its importance in TC intensification. Numerical simulations have shown that an instantaneous localized heat source (simulating latent heating associated with convection) can lead to intensification by contributing to a contraction of the maximum tangential winds radially and an expansion of these winds vertically (Persing et al. 2013; Smith et al. 2017). Alternatively, an emphasis on the azimuthal coverage of convection was advanced by Nolan and Grasso (2003) and Nolan et al. (2007), who noted that a greater azimuthal coverage of convection projects a greater amplitude of the heating onto the azimuthal-mean diabatic heating, which is a condition favorable for intensification. This relationship between precipitation symmetry and TC intensification has been documented using passive microwave satellite data (Alvey et al. 2015) and satellite-based radar (e.g., Kieper and Jiang 2012; Jiang and Ramirez 2013).

Attention in the literature has been focused on what processes govern the azimuthal distribution of convection. A significant influence is the response of the storm to deep-layer vertical wind shear (Marks et al. 1992; Black et al. 2002). In separate simulations of Hurricane Bonnie (1998), Rogers et al. (2003) and Braun et al. (2006) found the shear produced a wavenumber-1 asymmetry in vertical motion with upward motion in the direction of the vortex tilt. In trajectories of mesoscale air motions calculated from airborne Doppler analyses, Marks et al. (1992) were able to track an updraft that initiated at 1.1-km altitude on the downshear side of Hurricane Norbert (1984). Later studies suggested that the downshear-left (DSL) quadrant was the initiation quadrant because of the high occurrence of deep convection (Reasor et al. 2009; Molinari and Vollaro 2010; Nguyen and Molinari 2012). Reasor et al. (2013) and DeHart et al. (2014) showed that in a sheared storm, the secondary circulation has a deep layer of inflow in the downshear quadrants. The inflow is maximized in the downshear-right (DSR) quadrant, with convergence just inside the RMW. This leads to the downshear quadrants, but especially the DSR quadrant, being a preferential location for the initiation of deep convection. A case study of Hurricane Karl (2010) used airborne analyses to confirm that convection was initiated in both downshear quadrants due to convergence from low-level, counterrotating mesovortices embedded in the flow and the turbulent transport of positive buoyancy anomalies from the eye to the eyewall (Guimond et al. 2016).

After the convection is formed in this region and travels downwind, it grows in size and strength. In the upshear-left (USL) quadrant, there is generally a deep layer of inflow aloft that descends in the eyewall region (Reasor et al. 2013). This implies that as an updraft continues to travel downwind, it begins to encounter subsidence and drying in the USL quadrant and starts losing its size and structure. Such a structure and evolution was shown in the analysis of precipitation in Hurricane Erin (2001) in Halverson et al. (2006; see their Fig. 12).

The azimuthal distribution of CBs can have a significant relationship with TC intensification. In an analysis of aircraft observations of Hurricane Edouard (2014), Rogers et al. (2016) showed that CBs were located on the upshear side of the storm during the time that Edouard was intensifying. Two days later, when Edouard was beginning to weaken, there was a smaller number of CBs that were mostly confined to the downshear side of the storm. Such a relationship between the upshear distribution of CBs and TC intensification is consistent with the efficiency arguments relating the azimuthal-mean projection of diabatic heating to TC intensification, as articulated above and in Nolan et al. (2007). Zawislak et al. (2016) and Rogers et al. (2016) emphasized differences in the local environment of convection to explain the evolution of Edouard's CBs. In particular, sea surface cooling in the upshear-right (USR) and DSR quadrants prevented Edouard's boundary layer from recovering via surface enthalpy fluxes from downdraft cooling that occurred left of shear. The combination of both increased static stability and dry air upshear prevented significant

upshear propagation of CBs during the time Edouard was weakening. Using ensemble simulations of Hurricane Edouard, Leighton et al. (2018) noted that ensemble members that rapidly intensified showed a larger percentage of CBs USL, while those members that did not intensify had CBs that remained confined to the DSR quadrant. They attributed this difference to stronger zonal shear that effectively prevented the upshear movement of CBs for the nonintensifying ensemble members. Using a similar methodology, Munsell et al. (2017) found a similar result and attributed the differences between early- and late-RI-onset members to differences in the precession of the vortex and an associated increase in the upshear distribution of convection.

In another detailed case study, Nguyen et al. (2017) used aircraft observations to document differences in the distribution of precipitation upshear for Hurricanes Bertha and Cristobal (2014), finding that the greater amount of precipitation upshear for Bertha (a rapid intensifier) was due to a moister upshear environment, with less subsidence, than Cristobal (a slow intensifier). A related explanation for these different azimuthal distributions of CBs invokes tropical cyclone-relative environmental helicity (TCREH; Onderlinde and Nolan 2014, 2016). Onderlinde and Nolan (2016) found that the difference in intensification rate between TCs embedded in positive versus negative TCREH primarily results from the position of convection and associated latent heat fluxes relative to the wind shear vector. Trajectories for positive versus negative TCREH simulations showed that parcels in the positive TCREH (clockwise rotation of winds with height) case experienced a favorable vortex tilt that allowed for both longer-lasting convection and a greater recovery of PBL equivalent potential temperature downwind of convection through latent heat flux near the TC core. Air parcels that experience larger fluxes are more frequently ingested into the TC core, and convection is more readily advected upshear, resulting in intensification.

The distribution of convection of varying depths (e.g., shallow, moderate, and deep) has been the focus of many satellite studies relating convection to TC intensification. With a much larger database than the airborne Doppler, Zagrodnik and Jiang (2014) showed that a higher percentage of reflectivity greater than 20 dBZ occurred above 10 km in the upshear quadrants during RI. Additionally, a significant relationship between an azimuthal ring of precipitation in 37-GHz microwave imagery (highlighting weak-to-moderate updrafts) and RI onset was identified in Jiang (2012) and Kieper and Jiang (2012). Tao and Jiang (2015) found that the presence of widespread shallow convection (20-dBZ echo top below 6 km) is the best indicator of RI. They argued that the presence of deep

convection is a response to the strengthening vortex instead of a precursor to intensification.

As the above discussion illustrates, the TC intensity response to convection is sensitive to the radial and azimuthal location of convection, as well as its depth. What has not been examined in a composite study is how the characteristics of the convection, as measured by the strength of the peak updraft, altitude of the peak updraft, and echo top height, vary as a function of radial and azimuthal location and how those variations are related to TC intensity change. An improved understanding of these relationships can yield insight into the convective-scale and mesoscale processes responsible for intensity change.

The work presented here will examine these characteristics by comparing composites of the structure of convection from tail Doppler radar data on the National Oceanic and Atmospheric Administration (NOAA) WP-3D aircraft for missions from 1997 to 2010. Such a composite approach follows previous work examining composites for vertical velocity and reflectivity from airborne Doppler radar (e.g., Black et al. 1996; Heymsfield et al. 2010; DeHart et al. 2014) and reflectivity from the TRMM Precipitation Radar (Jiang 2012; Tao and Jiang 2015; Tao et al. 2017). What is unique about this study, though, is that it examines the composite structure of this convection from airborne Doppler radar in a framework that considers both TC intensity change and vertical shear. Similar to Heymsfield et al. (2010), the focus of these composites will be on the strongest updrafts (termed CBs here), since these features are the easiest to detect in the radar and have been shown to exhibit variations in their amount and distribution for different TC intensity change categories (Rogers et al. 2013, 2015, 2016). This is not to say, of course, that differences in the structure and distribution of more moderate updrafts are not important in TC intensity change, but initial efforts are aimed at examining these extreme portions of the vertical velocity distribution. Such a large database of observations allows for a robust statistical analysis of CB structure that highlights the cohesive structures that are present in various stratifications, including as a function of shear-relative quadrant, radial location, and the intensity change of the TC.

## 2. Data and methodology

#### a. Datasets used

This study used the tail Doppler radar database from the NOAA WP-3D aircraft from 1997 to 2010 that has been analyzed in several other studies (Reasor et al. 2013; Rogers et al. 2013; DeHart et al. 2014; Hazelton et al. 2015). The data came from 28 intensive observing periods (IOPs), of which 14 were from intensifying [IN; increase in storm intensity at a rate greater than or equal to  $20 \text{ kt} (24 \text{ h})^{-1}$  during 12 h following the IOP] storms, and 14 were from steady-state [SS; change in storm intensity at a rate less than or equal to  $10 \text{ kt} (24 \text{ h})^{-1}$  during the 12h following the IOP] storms. Each IOP lasted for  $\sim$ 4h and contained between two and six swaths, where a swath is defined as an eye penetration and a downwind leg. Each swath was analyzed on a Cartesian grid with a horizontal resolution of 2km and a vertical resolution of 0.5 km. This resolution is coarser than some other studies using airborne Doppler radar: for example, the vertical incidence composites from Black et al. (1996), which had an along-track spacing of ~750 m, and the NASA ER-2 Doppler radar composites from Heymsfield et al. (2010), which had an effective resolution of a few hundred meters at 10-km altitude to 500 m at the surface. However, it is felt that many of the gross characteristics of the strongest and largest updrafts are effectively captured here, as discussed in previous composite studies using the P-3 tail Doppler radar (e.g., Rogers et al. 2012, 2013; DeHart et al. 2014). A comparison of the error characteristics for tangential, radial, and vertical wind for the automated analyses used here with flight-level data was performed in Rogers et al. (2012; see their Table 2). For vertical wind, this comparison showed that the automated analyses had a root-mean-square (RMS) error of  $\sim 1.6 \,\mathrm{m\,s}^{-1}$ and a bias of  $\sim 0.1 \,\mathrm{m \, s^{-1}}$ . The RMS and bias values for these automated analyses are comparable with those from manually edited analyses for the flights in Hurricane Guillermo (1997). For a full description of the properties of the dataset and how it was developed, please refer to these papers.

The dataset was limited to TCs of hurricane strength to ensure a well-developed primary and secondary circulation. Additionally, to ensure that each TC had the thermodynamic potential to intensify, only storms that were at least 25 kt below their maximum potential intensity in the Statistical Hurricane Intensity Prediction Scheme (SHIPS) database (DeMaria and Kaplan 1999) were included. For each swath, the storm motion from the National Hurricane Center (NHC) best track data was subtracted from the analysis to get storm-relative motions.

The horizontal dimension for each swath was scaled by the RMW at 2km to create a common reference frame, and each storm was centered on the 2-km vortex center. No attempt was made to account for the slope of the RMW with height, though this may introduce some uncertainty in terms of radial location with respect to the local RMW (Hazelton et al. 2015; Rogers et al. 2015). However, adjusting for the local RMW creates more uncertainty, given occasionally sparse data coverage, particularly at higher altitudes.

For each IOP, data from each swath were also combined to create a merged analysis. With a greater horizontal coverage than individual swaths, these merged analyses are useful for resolving the vortex-scale circulations of the TC on longer time scales than localized convection (Rogers et al. 2013, 2015, 2016). The merged analyses were used to compute an azimuthal mean at each altitude and radial band of the radar-derived fields analyzed in this study (e.g., vertical velocity, radial wind, tangential wind, reflectivity, divergence, and vorticity). To analyze features on the scale of individual CBs that evolve over time scales smaller than the typical IOP, the individual swaths were used. This methodology is similar to that used in Rogers et al. (2013).

#### b. Identifying CBs

While previous studies focused on the top 1% of upper-level (at or above 8km) updrafts (Rogers et al. 2013, 2015, 2016), it was preferable in this study to consider convection at all altitudes to potentially sample convection at different stages of its life cycle. As a result, the entire vertical column, extending from 0.5- to 16-km altitude, was considered in identifying CBs. The peak vertical velocity for each column was computed, and from this distribution, the top 1%, calculated to be  $6.04 \,\mathrm{m \, s^{-1}}$ , was used as the vertical velocity threshold for identifying CBs. To test the sensitivity in this threshold, multiple vertical velocity distributions were created by changing the lower- and upper-altitude cutoffs. In all cases, the deviation from the  $6.04 \,\mathrm{m \, s^{-1}}$  cutoff was within 10%. For the 6.04 m s<sup>-1</sup> threshold used in this study, the maximum occurrence of CBs was within 8-12-km altitude, with a secondary peak between 2- and 4-km altitude, indicating that the results are not sensitive to small changes in the CB identification threshold. In addition to the vertical velocity threshold, the average reflectivity between 0.5 and 14 km simultaneously had to be greater than 10 dBZ for a CB to be identified at a given horizontal location. The latter criterion ensured that there were an adequate number of scatterers in a sufficiently deep vertical column at the location of the CB.

Individual grid points were considered as separate structures to be incorporated into the statistics. An attempt was made to group adjacent points on a swath that met the vertical velocity and reflectivity criteria and to identify them as one CB. This connectivity would eliminate oversampling of the larger updrafts embedded within the TC. However, this became unfeasible as the sample sizes of the CBs diminished, and the ability for a statistically robust analysis was frequently lost. Even so, when the points were grouped, the composite mean features had nearly the exact same physical characteristics as the composite fields shown in this study. An attempt was also made to have a vertical connectivity constraint to prevent oversampling of the same updraft, but due to the complex nature of the convective features and coarse horizontal resolution of the analyses, no physically feasible algorithm was found. For the purposes of this study, it was determined that because nearly similar results were obtained and a robust statistical analysis was warranted, not grouping connected points would still lead to an adequate understanding of how strong updrafts interact with the overall storm. Thus, it is more appropriate to think of the composite results as composites of locations with strong vertical velocities rather than as independent convective cells.

Figure 1 shows an example of CBs identified for the 1341 UTC center pass during the Hurricane Earl mission on 30 August 2010. Once identified, the CBs were grouped into a shear-relative framework based on the 850–200-hPa deep-layer shear from the SHIPS database (DeMaria et al. 2005). The shear values in this dataset ranged between 1.5 and 19 kt, with an average value of 11.75 kt. The average shear was 11.2 and 12.3 kt for IN and SS storms, respectively, indicating similar average environmental conditions between the two groups. An attempt was made to further stratify the dataset by shear magnitude, but the sample sizes quickly became too small for a statistically robust analysis.

# c. Statistical distributions and structural characteristics of CBs

Two types of analyses are performed on the composite datasets. The first set of analyses focuses on the statistical distributions of the strength and height of the peak updraft and height of 15-dBZ echo tops (analysis of 20and 12-dBZ echo tops was also performed, but those results were essentially identical to the 15-dBZ threshold). These parameters serve as proxies for the intensity and structure of the CBs and for how these properties vary as a function of shear-relative azimuth, radius relative to the RMW,  $r^*$  (i.e.,  $r^* = R/RMW$ ), and intensity change of the TC. To compare the robustness of the differences in the means of the magnitude and height of the peak vertical velocity in the column, as well as maximum height of the 15-dBZ echo top, of CBs for different TC intensity change categories, Student's t tests were performed at standard confidence intervals (i.e., 90%, 95%, and 99%). However, these tests need to be interpreted with caution since the distributions of these parameters are often nonnormal. Additionally, since individual grid points were used to identify CBs and, in most cases, multiple grid points constitute a convective core, not all the samples in the significance tests are truly independent. To account for these important caveats, boxplots (highlighting medians, the 25th and 75th percentiles, and extreme values of the distributions) were compared and used as the primary statistical analysis tool.



FIG. 1. An example of the CB identification algorithm for the 1341 UTC swath of Hurricane Earl on 30 Aug 2010. Shading is maximum vertical velocity in the vertical column. The 2-km RMW is identified as a black circle, and a shear-relative quadrant framework is superimposed. Black dots are columns that are identified as CBs. Contour interval is  $0.5 \text{ m s}^{-1}$ .

In addition to the statistical comparisons mentioned above, vertical cross sections extending 20 km radially inward and outward of an identified CB were constructed to examine the radius-height variation of the CB structure. For CBs that were within 20 km of the storm center, the cross section was cut off at the center of the storm. For each cross section (recall that they are created for each vertical column meeting the vertical velocity criteria), the three-dimensional kinematic fields were averaged 4 km azimuthally upwind and downwind to obtain the general signatures of the CB while ensuring the best data coverage. Figure 2 shows a schematic of this averaging technique. Then, for each radial cross section, the azimuthal mean from that IOP's merged analysis was subtracted from the total measured field to define a perturbation value. Finally, cross sections from multiple CBs were then averaged within this CB-relative coordinate system to construct a composite of CB structure. Only coordinates where greater than 30% of the individual CB cross sections contained data were included in the composite cross sections. Additionally, to avoid one TC dominating the composite structure, we required that data from at least three IOPs be averaged into each coordinate in the composite cross sections. This type of analysis allows for a determination of the local mesoscale imprint of a CB on the background circulation of the storm.

Sensitivity tests were performed to examine the robustness of this approach, given the variable range of data coverage in the IOPs. For multiple IOPs that contained greater than two swaths, azimuthal means of vertical velocity and reflectivity were computed using



FIG. 2. A schematic showing how the CB radial cross sections were computed. The inner and outer rings represent the R/RMW = 0.75 and 1.25, defined as the eyewall boundaries. The "×" represents the location of a CB. The length of the rectangle represents the radial bands where the cross section is taken from, while the width represents the averaging area for each radial band.

subsets of the swaths available. For all these cases, using any combination of at least two swaths gave a reasonable representation of the azimuthal mean calculated from the merged analysis (as determined by the location and strength of dynamical features, such as the eyewall). Therefore, it was determined that for IOPs where only two swaths were available, the azimuthal mean fields were representative of the values that would have been computed if more data were available, and the perturbation analysis remained valid. There are still identified limitations to the interpretation of the cross sections. Since the tilt of the CBs varies, averaging them together can create smeared structures that are nonrepresentative. A possible solution of aligning the CBs to angular momentum surfaces was identified, but in that case, the composite cross sections would lose a physical interpretation because each CB would be rotated at a different angle to align the angular momentum surfaces. The compositing method used in this study was considered the most robust for interpretation of CB gross characteristics, but because of the limitations discussed above, one should not interpret the cross sections as an average CB.

The spatial averaging and compositing procedure described above results in structures that are smoother than those obtained by individual case studies because the updrafts in this study are at different altitudes and stages of their life cycle (e.g., Reasor et al. 2009; Guimond et al. 2010, 2016; Didlake and Houze 2013; Rogers et al. 2015, 2016). However, it is felt that the ability to composite these structures in a CB-centric

framework, which allows for the determination of the gross characteristics of CBs and their mesoscale imprint on the vortex in a statistically robust manner, compensates for these drawbacks.

## 3. General characteristics of CBs

# a. All CBs

Figure 3 shows the perturbation vertical velocity and reflectivity fields from the composite of all CBs in the eyewall region ( $r^* = 0.75-1.25$ ) identified in this study. As expected, the perturbation vertical velocity field (Fig. 3a) shows a general area of enhanced ascent that extends  $\sim 10 \text{ km}$  radially inward and outward from the updraft center. This is a larger radial extent than the typical updraft size because some sampling points were near the edge of a major updraft. Additionally, updraft size may be misrepresented because of the relatively coarse 2-km resolution of the analyses.

On average, the ascent associated with the perturbation vertical velocity reaches a peak magnitude of  $\sim 4 \,\mathrm{m \, s^{-1}}$ at ~12-km altitude. The updraft is accompanied by perturbation downdrafts radially inward and outward from the CB core, with the stronger downdraft being radially inward and on the order of  $-1 \,\mathrm{m \, s^{-1}}$ . Both of these downdrafts are present and of a larger magnitude on the full composite-mean vertical velocity field, indicating actual areas of subsidence around a typical CB (not shown). This downdraft structure, most likely caused by divergence from the updraft core and evaporative cooling from the enhanced precipitation, is similar to that seen in a case study of Hurricane Dennis (2005) by Guimond et al. (2010). A similar relationship is seen in the perturbation reflectivity field (Fig. 3b). Accompanying the updraft are higher reflectivity values throughout the entire column. The maximum perturbation reflectivity is at  $\sim$ 6-km altitude with an  $\sim$ 8-dBZ magnitude. This maximum occurs radially inward of the axis of peak ascent, indicating that the precipitation core trails the large updraft velocities. There is also suppressed reflectivity 10-20 km radially inward of the CB core, spatially corresponding with the perturbation downdraft, and on the order of  $-2 \, dBZ$ .

In general, the CBs are accompanied by enhanced tangential winds (Fig. 4a) maximized in the low levels and radially outward of the axis of peak ascent, radial wind divergence (Fig. 4b) between ~10- and 14-km altitude, enhanced vertical vorticity at and radially inward of the updraft axis (Fig. 4c), and convergence near the base of the updraft (Fig. 4d). Multiple physical mechanisms are now explored for the root cause of these signatures.

First, for the low-level tangential jet, the low-level relative vorticity dipole may induce enhanced tangential



FIG. 3. A comparison between radial cross-sectional composites of (a) perturbation vertical velocity and (b) perturbation reflectivity from all CBs within  $r^* = 0.75-1.25$  along with the axis of peak vertical motion (black dashed line) from the perturbation vertical velocity in (a). Contour interval is  $0.5 \text{ m s}^{-1}$  for the vertical velocity and 1 dBZ for reflectivity.

winds in the CB center. Second, the low-level convergence may be associated with enhanced inward transport of absolute angular momentum surfaces just outside the axis of heating. Didlake and Houze (2011) found a similar enhancement of the tangential wind, though their observations were for rainbands, rather than the eyewall region shown here. These patterns are also consistent with case studies of deep convection that identified small-scale mesovortices embedded near the convective cores (e.g., Braun et al. 2006; Guimond et al. 2016; Hazelton et al. 2017).

Above the updraft core, there is enhanced storm center-relative inflow (outflow) radially inward (outward) of the CB, a signature consistent with the enhanced divergence at those altitudes. The radial wind field is the only one where the total field is shown. The perturbation cross section that did not yield physically intuitive results was radial wind, possibly because the radial flow can have significant asymmetries, typically with a wavenumber-1 pattern (e.g., Reasor et al. 2013; DeHart et al. 2014). Therefore, the azimuthal-mean field masks significant across-storm variabilities in the radial flow, confounding the determination of a clear perturbation signal.

Another important observation is that in the midlevels,  $\sim$ 5 km radially inward and  $\sim$ 10 km radially outward, the tangential wind speed is weakened when compared to the azimuthal mean. This negative perturbation can have a significant impact on the intensity of the TC, especially if the CB is outside the RMW. This negative perturbation in tangential velocity is consistent with the relative vorticity dipole that exists  $\sim 15 \text{ km ra}$ dially inward of the CB center. The more prominent vorticity dipole is near the CB radial location. It is characterized by higher vorticity radially inward and lower vorticity radially outward. The line of maximum ascent is displaced from the center and over the positive vorticity perturbation, indicating that both stretching of vertical vorticity and tilting of horizontal vorticity are important factors for vertical vorticity generation.

A look at individual updraft cross sections (not shown) indicates variation among the orientation of the updraft axis to the positive vorticity perturbation. In some cases, the maximum vorticity perturbation was aligned with the updraft axis, indicating that stretching is the main form of vorticity generation, which is consistent with other studies (e.g., Black et al. 1996; Hendricks et al. 2004; Guimond et al. 2010). To further investigate the mechanisms for vorticity generation, the composite structure was recalculated for CBs identified at 4-km altitude or below. The perturbation vertical velocity field (Fig. 5a) shows a peak near 4-km altitude. In this analysis, the maximum positive vorticity perturbation is closer to the axis of peak ascent (Fig. 5b). The closer alignment between the updraft axis and the peak positive relative vorticity perturbation shows that in the low levels, stretching is the dominant term for vertical vorticity generation. However, the vorticity dipole still exists, indicating that tilting of background horizontal vorticity is nonnegligible. The vorticity dipole just radially inward of the CB and at the low levels is likely tilting of horizontal vorticity generated by the strong vertical shear of the radial wind (Fig. 5d). Tilting also likely becomes more important in the midlevels, as the perturbation vorticity dipole is centered closer to the updraft axis in Fig. 4c. This is a result similar to that of the case studies of Supertyphoon Jangmi (2008) in Sanger et al. (2014) and



FIG. 4. Composite radial cross sections from all CBs of (a) perturbation tangential wind, (b) total radial wind, (c) perturbation relative vertical vorticity, and (d) perturbation divergence from all CBs within  $r^* = 0.75-1.25$ . Contour interval is  $1 \text{ m s}^{-1}$  for the tangential wind velocity,  $2 \text{ m s}^{-1}$  for the radial wind velocity, and  $2 \times 10^{-4} \text{ s}^{-1}$  for the vorticity and divergence fields.

Hurricane Rita (2005) in Didlake and Houze (2011). For CBs in the low levels, the resulting low-level tangential wind perturbation (Fig. 5c) is slightly larger than for all CBs. In all cases, the presence of CBs influences the low-level spinup of the TC. Last, low-level CBs are in an area of low-level radial wind convergence (Fig. 5d), with significantly more outflow in the midlevels than the outflow shown for all CBs (Fig. 4b).

The kinematic fields analyzed above (relative vorticity, radial wind, tangential wind, and divergence) are representative quantities that yield similar gross characteristics in all further stratifications based on radial or azimuthal location and intensity change. Therefore, they will not be shown in subsequent sections.

## b. Azimuthal variation

The general CB structure is now examined as a function of shear-relative quadrant. Boxplots showing the median, 25th and 75th percentiles, and outlier values of the distributions of peak updraft magnitude (Fig. 6a), height of the peak updraft (Fig. 6b), and height of the

15-dBZ echo top (Fig. 6c) are presented. Mean values are also shown here (as blue stars) to further understand the distribution of the measured values. In general, the peak updraft magnitude, height of the peak updraft, and echo-top heights are weaker and lower in this dataset than those shown for tropical cyclone convection in Heymsfield et al. (2010). This difference likely reflects differences in the resolution and sensitivity of the two radar systems. Nevertheless, a systematic variation in the structure of the convection as a function of shearrelative quadrant is seen here. The median updraft velocities have the weakest magnitude and are at the lowest altitude in the DSR quadrant. This quadrant typically contains the strongest low-level convergence (Reasor et al. 2013) and can be thought of as the quadrant where convection is most frequently initiated. In the quadrants downwind (i.e., DSL and USL), the updrafts are identified at steadily higher altitude and stronger magnitudes, suggesting that they are maturing. The USL quadrant has the largest median updraft magnitude and the highest altitude of peak updraft and



FIG. 5. Composite radial cross sections along with the axis of peak ascent (black dashed line) from all CBs identified at or below an altitude of 4 km and within  $r^* = 0.75-1.25$  of (a) perturbation vertical velocity, (b) perturbation relative vertical vorticity, (c) perturbation tangential wind, and (d) total radial wind. Contour interval is  $0.5 \text{ m s}^{-1}$  for the vertical velocity,  $2 \times 10^{-4} \text{ s}^{-1}$  for the vorticity,  $1 \text{ m s}^{-1}$  for the tangential wind velocity, and  $2 \text{ m s}^{-1}$  for the radial wind velocity.

height of 15-dBZ echo top. In this quadrant, there is a large spread in the distribution of updraft velocities but a small spread in peak updraft heights, which are mostly occurring between 10- and 13-km altitude. This indicates that the strongest updrafts at the highest altitudes are most prevalent in this quadrant. For the USR quadrant, where the kinematic and thermodynamic conditions are unfavorable for convection (e.g., Reasor et al. 2013; DeHart et al. 2014; Rogers et al. 2016; Zawislak et al. 2016), and where the updraft could have made the local environment unfavorable itself due to downdrafts (e.g., Molinari et al. 2013; Zhang et al. 2017b), the median updraft velocity and height has decreased from the USL. The height of the 15-dBZ echo top is at its lowest altitude, indicating a consistent lack of upper-level clouds and moisture in this quadrant.

The quadrant-averaged structural differences in the updrafts are shown using the radial cross-sectional composites of perturbation vertical velocity (Fig. 7). While

there are many similarities among the perturbation vertical velocities in the four quadrants, there are some differences to be noted. As with the statistical analysis shown in Fig. 6, the USL quadrant has the strongest perturbation updrafts at the highest altitudes, with downdraft perturbations both radially inward and outward of the CB. The extensive coverage of downdrafts USL can be attributed to the general shear-induced subsidence seen in this quadrant (Reasor et al. 2013; DeHart et al. 2014). The DSR shows updrafts peaking at  $\sim$ 12 km, significantly higher than that suggested by Fig. 6b. The higher peak altitude was determined to be from a small cluster of strong updrafts and high altitudes (where there is less data coverage) that was able to dominate the mean structure. Combining the statistics and composite cross sections suggests that CBs with the maximum vertical velocity in the midlevels also had large vertical velocities aloft. However, CBs that had a maximum vertical velocity in the upper levels did not extend into the midlevels. The DSL and USR



FIG. 6. (a) Boxplot comparison of the maximum updraft magnitude between the shear-relative quadrants; (b) as in (a), but for the height of the maximum updraft; (c) as in (a), but for the maximum height of the 15-dBZ echo top. In the boxplots, the red line in the box denotes the median value, while the upper and lower edges

quadrants show a combination of characteristics between the other two quadrants. Many of these variations will be further discussed in the following sections, as the quadrant-averaged signals are different between TC intensity change categories.

# 4. CB structure variation as a function of TC intensity change

## a. Azimuthal variation

The relative azimuthal distributions of CBs as a function of intensity change are shown in Fig. 8. Similar to Rogers et al. (2013), IN storms have more than twice as many CBs in the eyewall region than SS storms (1529 vs 690). In terms of the azimuthal distribution, SS storms have the majority of CBs on the downshear (DSR + DSL)side, while IN storms have the majority of CBs left of shear (DSL + USL). Two quadrants in particular are noteworthy. In the USL quadrant, the proportion of CBs is 3 times higher for IN storms, compared to SS storms. This relationship is consistent with satellite composite studies (e.g., Zagrodnik and Jiang 2014) and an airborne case study of Hurricane Edouard (Rogers et al. 2016). Rogers et al. (2013) found that in the USL quadrant, there is a general area of upper-level subsidence in the eyewall region for SS storms, compared to a region of upward motion for IN storms. It is not clear whether the differences in composite vertical motion simply reflect different amounts of CBs in the USL quadrant for IN versus SS cases, or if the mesoscale vertical motion differences are a potential cause of the USL variability in CB numbers. A marked difference in CB proportion is also seen DSR, where SS storms have 6 times the proportion of CBs than IN storms. These results indicate that intensifying TCs have a greater proportion of CBs on the upshear side than SS TCs, whose CBs remain downshear.

Table 1 summarizes the differences between IN and SS storms in terms of average peak updraft value, height of that peak updraft, and height of the 15-dBZ echo within the eyewall region. Statistically significant differences in these categories between the CBs from IN and SS storms at greater than 95% certainty are bolded

 $<sup>\</sup>leftarrow$ 

of the box represent 75th and 25th quartiles, respectively. The difference between the 75th and 25th percentiles represents the interquartile range (IQR). Whiskers extending above and below the box represent either  $1.5 \times IQR$  above the 75th and below the 25th percentiles respectively, or to the maximum/minimum values. Beyond the whiskers, values are statistical outliers and are represented as red plus signs.



FIG. 7. Composite perturbation vertical velocity for CBs in the  $r^* = 0.75$ –1.25 region and stratified by shearrelative quadrant. With the shear vector pointing right, the quadrants are (a) USL, (b) DSL, (c) USR, and (d) DSR. Contour interval is  $0.5 \text{ m s}^{-1}$  for all plots.

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and italicized.1 Significant differences in the characteristics of the updrafts are also noted in the boxplots comparing the distributions of updraft value (Fig. 9a), height of peak updraft (Fig. 9b), and height of 15-dBZ echo top (Fig. 9c) between IN and SS storms.

Radial Distance From Burst Center (km)

-20

In the DSR quadrant, the CBs from SS storms had stronger peak updrafts by  $0.31 \,\mathrm{m \, s^{-1}}$ , with a significance level greater than 95%. Even though it is not statistically significant, there is a 0.89-km-higher 15-dBZ echo top in the SS storms, indicating more-developed convective cores in this quadrant. No significant difference in the height of the peak updraft was seen, which is consistent with having a mix of developed and initiating convection in this quadrant.

In the DSL quadrant, where the most CBs were identified in both IN and SS storms, the CBs from IN

storms were marginally stronger than in SS storms. Statistically significant differences were seen in the height of the 15-dBZ echo top, with IN storms being at a higher altitude by 0.85 km. While the mean values of the other two quantities were not statistically significant, the median values in both the updraft velocity and height were higher for IN storms by  $0.41 \,\mathrm{m \, s^{-1}}$  and  $0.5 \,\mathrm{km}$ , respectively.

20

0

Radial Distance From Burst Center (km)

The USL quadrant contains the largest and most statistically significant differences in the updraft strength and structure. The differences in mean updraft strength, height, and 15-dBZ echo-top height are  $1.67 \text{ m s}^{-1}$ , 3.55 km, and 2.34 km, respectively, with CBs from IN storms being stronger and at a higher altitude (Table 1). The distribution of peak updraft height for IN storms is small, compared to all the other quadrants with the 25th and 75th percentiles of the distribution, being separated by 2.5 km in this quadrant. The vast majority of the updrafts peak near 12-km altitude. The narrow distribution in altitudes (Fig. 9b) and high occurrence rates (Fig. 8) indicate a preferential azimuthal location and altitude of CBs for intensification. The increase of updraft strength, height, and echo top from DSL to USL for IN storms suggests that CBs continue to develop

<sup>&</sup>lt;sup>1</sup> It should be noted that the statistical part of this study should not be directly compared to the composite radial cross sections, since the composite cross sections are based off of 8-km averaging in azimuth, while the statistics are calculated from individual grid points.

Azimuthal Distribution of CBs in Eyewall Region



FIG. 8. The relative distribution of CBs for IN and SS storms in the  $r^* = 0.75-1.25$  region based on shear-relative quadrant. The total amount of CBs is 1529 in IN storms and 690 in SS storms.

as they travel around to the upshear side of intensifying TCs. A similar structure was seen in an azimuth-height cross section in Hurricane Edouard [cf. Fig. 7 of Rogers et al. (2016)]. By contrast, CBs have nearly the same strength (or slightly weaker strength) in the USL as in the DSL quadrants for SS TCs.

The significant structural differences for the convection in the USL quadrant between IN and SS storms is reflected in radial cross-sectional composites of perturbation vertical velocity and reflectivity fields for this quadrant (Fig. 10). The maximum in vertical velocity for CBs in IN storms is stronger, at a higher altitude, and spans a larger radial distance than CBs in SS storms. Additionally, the reflectivity perturbations are maximized at a higher altitude and extend over a deeper layer for CBs from IN storms, compared with SS storms. These structural differences indicate that the convection in SS storms may be weakening, possibly due to an unfavorable local environment in the USL quadrant, as suggested in Zawislak et al. (2016) and Rogers et al. (2016).

The USR quadrant also showed significant differences in the structure of the CBs (Fig. 9). The IN storms had CBs with stronger mean updraft strength, height of peak updraft, and height of 15-dBZ echo top, with differences of  $1.1 \text{ m s}^{-1}$ , 4.0 km, and 1.03 km, respectively. While these results are statistically significant, the sample size is much smaller, and there was more variability in this quadrant than USL, making the results less statistically robust. Nevertheless, they are consistent with the notion that the local environment on the upshear side is more favorable for the development and maintenance of CBs in IN TCs, compared to SS TCs.

# b. Radial variation

The relative radial distributions of CBs as a function of intensity change are shown in Fig. 11. For both the IN and SS storms, the peak distribution of CBs was in the

TABLE 1. The differences between the IN and SS storms (IN - SS) in average CB properties for the shear-relative quadrants. Boldface values are significant at the 95% confidence level.

Quadrant	Peak updraft value $(m s^{-1})$	Height of peak updraft (km)	Height of 15-dBZ echo top (km)
DSL	0.11	-0.03	0.85
USL	1.67	3.55	2.34
USR	1.10	3.99	1.03
DSR	-0.31	-0.11	-0.89

 $r^* = 1.0-1.25$  radial band, with the second-highest frequency in the  $r^* = 0.75 - 1.0$  radial band and third-highest frequency in the  $r^* = 1.25 - 1.5$  radial band. Note that this result is different from that found in Rogers et al. (2013), which found that the peak in the distribution of CBs for IN (SS) storms was inside (outside) the 2-km altitude RMW. This difference is because Rogers et al. (2013) defined CBs based on the vertical velocity at 8-km altitude, whereas in the study, CBs were between 0.5 and 16 km. Since updrafts slope outward with height (Marks and Houze 1987; Stern et al. 2014; Hazelton et al. 2015), the distribution of CBs in this paper is slightly different. Despite the differences in the radial distributions of CBs, there is still a larger proportion of CBs in the 0.75-1.25 radial band for IN storms than for SS storms and a higher proportion of CBs in the 1.25-1.5 band for SS storms than for IN storms, as shown in Rogers et al. (2013).

In addition to the differences in the radial distribution of CBs between IN and SS storms, the storms' structures can also vary significantly. Table 2 shows the differences of the mean values in the three parameters tested between IN and SS storms. In the  $r^* = 0.75-1.0$  region, the IN storms had CBs with a greater mean peak updraft speed of  $0.92 \text{ m s}^{-1}$ , height of peak updraft of 1.93 km, and height of 15-dBZ echo top of 1.85 km than the SS storms. These differences are beyond the statistical 95% confidence limit. Boxplots showing the distributions of these fields are presented in Fig. 12. In this radial band, the median and 75th percentiles of the distribution of peak updraft value for CBs in IN storms were notably larger than for SS storms. A similar relationship between CBs in IN and SS storms was seen for the height of peak updraft and 15-dBZ echo top.

In terms of the composite radial structure, there is a clear maximum in perturbation vertical velocity at  $\sim$ 12-km altitude for IN storms (Fig. 13a), while for SS storms, the vertical velocity field is weaker, with a suggestion of a double maximum (Fig. 13c). This indicates that the region has better-developed and more mature updrafts inside the RMW while the storm is intensifying. Stronger and deeper convection just inside the RMW is favorable for intensification because this region is characterized by



FIG. 9. (a) Boxplot comparison of the maximum updraft magnitude between the shear-relative quadrants and TC intensity change; (b) as in (a), but for the height of the maximum updraft; (c) as in (a), but for the maximum height of the 15-dBZ echo top.

higher vorticity and inertial stability [see, e.g., Fig. 15 in Rogers et al. (2013) using the same dataset], making it a more efficient region for converting diabatic heating from CBs into increased tangential winds (Schubert and Hack 1982; Nolan et al. 2007; Vigh and Schubert 2009; Pendergrass and Willoughby 2009). Also, this region provides a favorable location for the convergence of angular momentum surfaces in the boundary layer inside the RMW (Smith and Montgomery 2016), accompanied by vortex spinup.

In the  $r^* = 1.0-1.25$  region, we see similar characteristics for the distribution of CB peak velocities and altitudes. The respective means for the three quantities analyzed in IN and SS storms are larger for the  $r^* = 1.0-1.25$ region than in the  $r^* = 0.75 - 1.0$  region (Fig. 12). There was a greater increase in the CBs associated with SS storms, but there is still a statistically significant difference in mean velocities of  $0.36 \,\mathrm{m\,s^{-1}}$  and difference in peak updraft height of 1.41 km (cf. Table 2). The 15-dBZ echotop difference is not statistically significant. In this radial band, the CBs from SS storms also spread through a larger range of strengths and heights than for the  $r^* = 0.75-1.0$ region, which was mostly constrained to the weaker and lower-altitude updrafts (Figs. 12a,b). Because this region has the most minimal differences between CBs in IN and SS storms, the composite cross sections are not shown.

The opposite relationships between CBs from IN and SS storms are seen in the  $r^* = 1.25 - 1.5$  region. SS storms have larger mean peak updrafts by  $0.49 \,\mathrm{m \, s^{-1}}$  (Table 2), even though their median values are very similar (Fig. 12a). The echo tops in both IN and SS storms are at their highest altitudes, but the differences between them in this radial band are not statistically significant (Fig. 12c). Differences in the peak updraft strength for CBs in SS versus IN storms are mostly due to a decrease in the average peak updraft for IN storms. For example, the average peak CB vertical velocity for the  $r^* = 1.25-1.5$ region in SS storms is  $7.56 \,\mathrm{m \, s^{-1}}$ , which is weaker than the  $r^* = 1.0-1.25$  region with a mean CB velocity of 7.71 m s<sup>-1</sup>. For the IN cases, however, the mean CB velocity in the  $r^* = 1.25 - 1.5$  region is  $7.07 \,\mathrm{m \, s^{-1}}$ , while for the inner eyewall region, it is  $8.06 \,\mathrm{m \, s^{-1}}$ . This indicates that the strength of the convection within IN storms decreases rapidly outside the eyewall region but stays relatively constant in SS storms.

In terms of structural characteristics, the CBs from SS storms in the  $r^* = 1.25$ –1.5 radial band (Fig. 12d) have a stronger peak updraft, confined to a higher altitude, than CBs in the  $r^* = 0.75$ –1.0 radial band (Fig. 12b). When comparing with the CBs from IN storms in the  $r^* = 1.25$ –1.5 radial band (Fig. 12c), the average perturbation vertical velocity in CBs from SS storms has a larger value. The



FIG. 10. Radial cross-sectional composites for CBs in the USL quadrant for (a) IN perturbation vertical velocity, (b) SS perturbation vertical velocity, (c) IN perturbation reflectivity along with the axis of peak ascent (black dashed line) in (a), and (d) SS perturbation reflectivity along with the axis of peak ascent (black dashed line) in (b). Contour interval is  $0.5 \text{ m s}^{-1}$  for the vertical velocity plots and 1 dBZ for the reflectivity plots.

greater number of CBs and stronger convection outside the RMW for SS storms may also reflect the development of active outer rainbands or secondary eyewalls (Didlake and Houze 2013; Rozoff et al. 2012).

#### 5. Summary and conclusions

The structure of CBs in the eyewall region of TCs has been examined using a composite airborne Doppler radar database for IN and SS storms. CBs were identified as horizontal columns whose vertical velocity exceeded the 99th percentile of  $6.04 \text{ m s}^{-1}$  and do not represent individual convective cores. In general, and consistent with numerous observational and modeling case studies, CBs in TCs are associated with precipitating updrafts in the mid- to upper troposphere with subsidence radially inward and outward. They are associated with positive perturbation reflectivity in the updraft core and negative perturbation reflectivity at approximately 10km radially inward and outward from the updraft center. Relative vertical vorticity is enhanced along the updraft axis. It is also enhanced radially inward and reduced radially outward of the CB. The analysis indicated that stretching of vertical vorticity was prominent throughout the whole column, while tilting of horizontal vortex lines contributed to the generation of vorticity in the midlevels. There is also enhanced tangential wind



FIG. 11. The relative distribution of CBs for IN and SS storms, based on radial location relative to the RMW ( $r^* = 0.25$  bin size). The total amount of CBs in these radial bands is 2216 in IN storms and 1050 in SS storms.

TABLE 2. The differences between the IN and SS storms (IN - SS) in average CB properties for three radial bands. Boldface values are significant at the 95% confidence level.

Radial band	Peak updraft value (m s $^{-1}$ )	Height of peak updraft (km)	Height of 15-dBZ echo top (km)
0.75-1.0	0.92	1.93	1.85
1.0-1.25	0.36	1.41	-0.16
1.25–1.5	-0.49	-0.11	-0.36

associated with the updraft-induced positive perturbation of relative vorticity and divergence coming out of the updraft in the upper troposphere. These signals could be associated with smaller-scale mesovortices that have been shown to be prevalent in the CB life cycle (e.g., Braun et al. 2006; Guimond et al. 2016; Hazelton et al. 2017).

Differences in CB structure are observed among shear-relative quadrants of the eyewall. In the DSR quadrant, eyewall updrafts are seen at a broad range of altitudes, indicating convection sampled at various stages of its life cycle. Conversely, in the USL quadrant, updrafts are confined to the highest altitudes, and updraft strength and height are maximized in this quadrant. The DSL and USR quadrants show values of updraft strength and height intermediate to those DSR and USL. This azimuthal distribution suggests a life cycle of initiation of CBs (DSR), maturation (DSL), peak strength and height (USL), and weakening (USR). A similar type of evolution was shown in DeHart et al. (2014). It is important to note that the composite results are generalized and contain variability among individual CBs. Caution should be taken in interpreting the structure when comparing to individual case studies.

Further stratifying by TC intensity change highlights noteworthy differences in both the structure and distribution of CBs. The greatest differences in terms of shearrelative azimuth are seen in the DSR and USL quadrants. In general, for IN storms, there are more CBs with stronger updrafts at a higher altitude and higher echo tops in the USL quadrant, compared with SS storms. Conversely, for SS storms, there are more CBs with stronger updrafts and higher echo tops in the DSR quadrant, compared with IN storms. In terms of the radial distribution, CBs in IN storms are preferentially located inside the RMW, with stronger updrafts and higher echo tops inside the RMW, compared with SS storms. All of the above relationships are significant at the 95% confidence level or greater.

Figure 14 provides a descriptive summary schematic of the differences in the structure and distribution of CBs for IN versus SS TCs as a function of altitude, shearrelative azimuth, and radial location normalized by the 2-km RMW. It is worth noting that this schematic is only



FIG. 12. (a) Boxplot comparison of the maximum updraft magnitude between radial band and TC intensity change; (b) as in (a), but for the height of the maximum updraft; (c) as in (a), but for the maximum height of the 15-dBZ echo top.



FIG. 13. Radial cross-sectional composites of perturbation vertical velocity of (a) IN storms  $r^* = 0.75-1.0$ ; (b) as in (a), but for SS storms; (c) IN storms  $r^* = 1.25-1.5$ ; (d) as in (c), but for SS storms. Contour interval is  $0.5 \text{ m s}^{-1}$  for all plots.

showing the region of active convection in the CBs and does not include the anvils, even though they can contain substantial amounts of mass/moisture. Given limitations in the radar dataset and the nature of compositing, there is no way to identify the physical mechanisms underlying these differences. However, some inferences can be made regarding possible mechanisms responsible for the observed differences between IN and SS TCs based on the results of previous studies. For example, the differences in the radial distribution of CBs may be attributed to the inertial stability distribution outside the RMW. For this dataset, the outer-core wind field was stronger relative to the wind speed at the RMW for SS storms (Rogers et al. 2013), indicating a region of higher inertial stability in that region that may reduce radial inflow and preferentially lead to low-level convergence and CB initiation. Conversely, TCs with a weaker outer-core wind field and inertial stability may allow a greater penetration of lowlevel inflow inside the RMW, where strong low-level convergence, combined with a longer time for destabilization via surface enthalpy fluxes, allows for deeper

convection inside the RMW and TC intensification. These wind field differences were also seen by Zhang et al. (2017a) in simulations using the Hurricane Weather Research and Forecasting (HWRF) Model. This stronger inflow for IN cases can also import angular momentum at a higher rate than cases with weaker inflow, allowing for TC spinup (Smith and Montgomery 2016).

Possible explanations for the differences in strength and distribution in CBs as a function of shear-relative azimuth are related to the TC's response to varying environmental conditions. For TCs experiencing large amounts of shear, one would expect the downshear quadrants to have the highest concentration of CBs due to enhanced shear-induced low-level convergence. The difference in the azimuthal distribution for IN and SS storms may be related to the ability of CBs to persist into the upshear quadrants once initiated downshear.

In the USL quadrant specifically, the structural differences seen in CBs from IN and SS storms suggest that the local environment of CBs for SS storms was unfavorable for the persistence of deep convection upshear,



FIG. 14. Simple schematic showing average updraft profile and shear-relative and RMW-normalized distribution of CBs for (a) intensifying TCs and (b) steady-state TCs. Shear direction, shear-relative quadrants, and RMW indicated. Scalloped areas denote locations of CBs; dark (light) shading denotes CBs located downshear (upshear). Thickness and height of arrows denote relative strength and altitude range of peak updrafts.

compared with the local environment of CBs for IN storms. Some theories for the suppression of convection in the USL quadrant are the entrainment of environmental dry air (Nguyen et al. 2017), lack of midlevel background humidity and surface enthalpy fluxes (Onderlinde and Nolan 2016; Zawislak et al. 2016; Rogers et al. 2016), and strength and direction of the shear (Leighton et al. 2018).

Regardless of the mechanism, it appears from these results that the upshear side of the storm is a crucial azimuthal location for analyzing the local environment of the vortex and the structure of convection when considering the potential for intensification of the TC. Further research into these mechanisms is ongoing.

In an operational framework, without the presence of aircraft observations, it would be ideal to use more readily available satellite measurements to estimate where deep convection is located. Some possibilities for this are the National Aeronautics and Space Administration (NASA) Global Precipitation Measurement (GPM) space-based radar measurements; the upcoming Time-Resolved Observations of Precipitation structure and storm Intensity with a Constellation of Smallsats (TROPICS) cubesat constellation, which is set to measure ice scattering with high time resolution; or a lightning detection network, such as that on *Geostationary Operational Environmental Satellite-16 (GOES-16)*.

Expanding on this study, it would be ideal to incorporate more cases from recent years into the database. This should lead to more data coverage and a greater analysis of CB structure, especially in the USR quadrant, to give further insight into the TCs' response to deep-layer shear. Additionally, the larger database could allow for an analysis of independent convective cores with significance tests that use truly independent samples. Analysis of the broader spectrum of updrafts, rather than simply the strongest updrafts, would also shed light on the role of weak and moderate updrafts on TC structure and evolution. This is important, since these weaker updrafts may accomplish the bulk of the vertical mass flux in the hurricane eyewall (Braun and Wu 2007; Rogers 2010; Rogers et al. 2013). An examination of downdraft statistics and structure would also yield important insight on the role of these structures in warming and stabilization within the inner core. Adding thermodynamic data to analyses would be ideal to do a shear-relative quadrant analysis of the thermodynamic environment and its relationship to CB structural variations. This type of analysis is currently not able to be done on a composite basis due to the generally low sampling frequency of CBs from dropsondes and lack of mid- to upper-tropospheric sampling; the P-3 generally flies at an altitude below where most of the CBs exist. However, high-resolution modeling studies can be done to complete a temporal analysis of the structure of CBs and their local environments throughout their and the TCs' life cycles to help quantify more environmental factors inhibiting the convection (e.g., entrainment of dry air).

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