Tropical Cyclone Outflow and Warm Core Structure as Revealed by HS3 Dropsonde Data

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ABSTRACT

Dropsonde data collected during the NASA Hurricane and Severe Storm Sentinel (HS3) field campaign from 16 research missions spanning 6 tropical cyclones (TCs) are investigated, with an emphasis on TC outflow and the warm core. The Global Hawk (GH) AV-6 aircraft provided a unique opportunity to investigate the outflow characteristics due to a combination of 18–6-h flight durations and the ability to release dropsondes from high altitudes above 100 hPa. Intensifying TCs are found to be associated with stronger upper-level divergence and radial outflow relative to nonintensifying TCs in the sample, regardless of current intensity. A layer of 2–4 m s\(^{-1}\) inflow 20–50 hPa deep is also observed 50–100 hPa above the maximum outflow layer, which appears to be associated with lower-stratospheric descent above the eye. The potential temperature of the outflow is found to be more strongly correlated with the equivalent potential temperature of the boundary layer inflow than to the present storm intensity, consistent with the outflow temperature having a stronger relationship with potential intensity than actual intensity. Finally, the outflow originates from a region of low inertial stability that extends above the cyclone from 300 to 150 hPa and from 50- to 200-km radius.

The unique nature of this dataset allows the height and structure of the warm core also to be investigated. The magnitude of the warm core was found to be positively correlated with TC intensity, while the height of the warm core was weakly positively correlated with intensity. Finally, neither the height nor magnitude of the warm core exhibits any meaningful relationship with intensity change.

1. Introduction

Until recently, a single ER-2 flight over Hurricane Erin (2001) provided the only direct dropsonde observations through the full depth of the tropical cyclone (TC) outflow layer (Halverson et al. 2006). Conventional aircraft observations of TCs, such as by the U.S. Air Force C-130s and the NOAA P-3s, tend to be limited to the middle to lower levels of the cyclone with a typical flight level of 700 hPa (Aberson et al. 2006). Synoptic observations provided by the NOAA G-IV are of higher altitude, up to about 150 hPa (Rogers et al. 2002), but still frequently fail to capture the top of the outflow layer. The tropical cyclone (TC) outflow layer is challenging to observe due to the fact that it can extend from 300 hPa to as high as 100 hPa, above where most research aircraft operate. The G-IV also generally avoids overflying the TC core. Finally, special NASA DC-8 dropsondes into Super Typhoon Flo (1990) sampled the outflow up to 195 hPa (Titley and Elsberry 2000). Only just recently with the NASA Genesis and Rapid Intensification Processes (GRIP) field experiment of 2010\(^1\) (Braun et al. 2013) and the subsequent NASA Hurricane and Severe Storm Sentinel (HS3; Braun et al. 2016) investigation from 2012 to 2014 have routine observations of the TC upper levels been obtained, and only for HS3 were dropsondes deployed. By capitalizing on the ability of the unmanned NASA Global Hawk (GH) to fly at a cruising altitude above 100 hPa and also overfly the TC core (Braun et al. 2013), HS3 made it possible to not only sample the entire depth of the outflow layer, but also to sample the region in which the radial outflow originates over the deepest convection, hereafter referred to as the outflow roots. Several studies have already examined data from GRIP, focusing on both genesis and rapid intensification (e.g., Davis and Ahijevych 2013; Zawislak and Zipser 2014; Helms and Hart 2015;...
The primary goal of HS3 was to observe hurricane formation and intensity change, motivated by hypotheses related to the complex interaction between the large-scale environment and hurricane internal dynamics (Braun et al. 2016). HS3 operated research flights during 5-week campaigns in the August–September time frame during the 3-yr period of 2012–14. Two GH aircraft were available for HS3, designated “air vehicle one” (AV-1) and “air vehicle six” (AV-6). AV-6 was equipped with an NCAR–NOAA miniature dropsonde system known as the Airborne Vertical Atmospheric Profiling System (AVAPS) II, which was designed for remote operation, and could launch up to 88 dropsondes in a single flight while tracking up to eight dropsondes in the air simultaneously. There were 25 research flights conducted by HS3 from 2012 through 2014, resulting in 1506 quality-controlled dropsondes in the final dataset (Wick 2015).

In addition to the ability to operate at very high altitudes, the GH also had the unique capability to fly very long missions of ~24-h duration. This capability allowed the aircraft to perform multiple transects over the core while also sampling the environment around the TC, making it possible to render a more complete depiction of the three-dimensional structure of the cyclone and its environment from the dropsonde data alone.

While there are a wide number of potential applications of the HS3 dataset, it is clear that one of the most unique aspects of this dataset is the high spatial resolution and coverage of in situ observations above the outflow and outflow-roots region of the TC. Because of the limited nature of in situ measurements of TC outflow, previous studies (e.g., Sadler 1976, 1978; Merrill 1988a,b; Merrill and Velden 1996) used a combination of satellite-based atmospheric motion vectors (AMVs), commercial aircraft [or NASA DCS dropsondes from the Tropical Cyclone Motion (TCM-90) field experiment in the case of Merrill and Velden (1996)], and rawinsondes of limited spatial density to generate analyses of the outflow layer. Merrill (1988a) found stronger radial outflow, but weaker anticyclonic flow to be associated with the intensifying cyclones, with greater environmental vertical wind shear near or over the TC center for nonintensifying cases. Anticyclonic flow associated with intensifying TCs also tended to be more directly centered on the TC itself, whereas for nonintensifying TCs the anticyclone was displaced to the downshear right (Merrill 1988a). A number of studies (Sadler 1976, 1978; Merrill and Velden 1996) also found an association between TCs with multiple outflow channels and intensification. Additionally, Merrill and Velden (1996) found an increase in the height of the level of strongest outflow as well as an increase in the vertical depth of the outflow layer during the intensification of Super Typhoon Flo (1990). Their results show the equatorward outflow jet maximum to occur at a higher level than the poleward outflow, but with slightly stronger maximum velocities in the northern outflow. Here, we revisit some of these previous results given the superior spatial density and high vertical resolution of the HS3 dataset.

To understand the relationship between outflow, divergence, and the upper-level anticyclone above the TC, it is beneficial to first examine the vorticity equation in pressure coordinates (e.g., Holton 2004):

\[
\frac{D}{Dt}(\zeta + f) = -\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) - \left(\frac{\partial \omega}{\partial x} - \frac{\partial \omega}{\partial y}\right),
\]

where \(\zeta\) is the relative vorticity on a constant pressure surface; \(f\) is the Coriolis parameter; \(u\) and \(v\) are the zonal and meridional components of wind, respectively; \(p\) is pressure; and \(\omega = Dp/Dt\) is the pressure change following the motion. The two terms on the right-hand side of Eq. (1) are known as the stretching and tilting terms, respectively. TC outflow originates above the deep convective core of the cyclone as air parcels ascend parallel to lines of constant angular momentum, which are steeply slanted in the eyewall but till to become horizontal at upper levels (Emanuel 1986). In this region, convergence will result in an increase in \(\zeta\) while divergence will result in a decrease in \(\zeta\) for a given air parcel, as a result of the stretching term assuming all else being equal and at constant latitude. Divergent flow in the outflow roots region will contribute to anticyclonic \(\zeta\), which in turn intensifies and builds the upper-level anticyclone. The radius at which the outflow will expand against the ambient environment is theoretically equal to the Rossby radius of deformation (\(R_d\); Rappin et al. 2011), although in the real atmosphere it is often difficult to compute \(R_d\) accurately.\(^2\)

It has been argued that TCs primarily interact with environmental flow at the upper levels where the inertial stability \(I\) is weaker (Holland and Merrill 1984):

\[
I^2 = \left(\frac{f + 2V_t}{r}\right)(f + \zeta),
\]

where \(\zeta = (1/r)\left[\partial(rV_t)/\partial r\right]\) and \(V_t\) is the tangential component of wind (Alaka 1961). Greater \(I\) at the low to

\(^2\)The fact that \(R_d\) is inversely proportional to \(f\), which varies with latitude, also complicates matters, favoring a smaller (larger) radius anticyclone poleward (equatorward) of the TC.
midlevels resists radial inflow or outflow, especially for stronger storms (Schubert and Hack 1982; Shapiro and Willoughby 1982; Hack and Schubert 1986). In one of a very few studies to compute I using dropsonde data in the outflow region, Molinari and Vollaro (2014) found radial outflow to be maximized at the outer edge of the inertially unstable region, and hypothesized that the length of time that the inertially unstable region persists may be related to the nature or even the length of the period of TC intensification. A few recent modeling studies also examine the relationship between changes in environmental or vortex inertial stability and changes in TC structure and intensity. Rappin et al. (2011) argue that a reduction in inertial stability north of the TC associated with the insertion of a westerly zonal jet minimizes the reduction in inertial stability north of the TC.

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As is the case with outflow, the TC warm core has traditionally been infrequently documented by in situ observations because of its combination of high altitude and collocation with the core of the cyclone, making it difficult for conventional research aircraft to reach. However, documenting the magnitude and height of the warm core may be fundamental to fully understanding the balanced structure of the TC vortex. Early studies such as La Seur and Hawkins (1963), Hawkins and Rubsam (1968), and Hawkins and Imbembo (1979) all find maximum warm anomalies of 9°C–16°C close to 250 hPa. Hawkins and Imbembo (1976) also found a lower secondary maximum in Hurricane Inez (1966) from 650 to 600 hPa. However, these flights all inferred the height and strength of the warm core based upon interpolation of flight-level data from three to five different vertical levels, so uncertainty was large. Halverson et al. (2006) found a warm core maximum of 11°C at 500 hPa in Hurricane Erin (2001) using direct dropsonde measurements. They attributed finding such a low-altitude warm core to the fact that the storm was steadily weakening during the observational period. Examining a combination of dropsondes, aircraft flight-level data and WP-3D-based radar data, Dolling and Barnes (2012) argue that there is an early warm-core-like feature with a maximum temperature anomaly of 7°C as low as 800 hPa during the development of Tropical Storm (TS) Humberto (2001). However, the occurrence of a warm core at such low levels does not appear to be typical for mature TCs. Durden (2013) examined dropsondes and surface-based soundings from various NASA field campaigns with data up to ~250 hPa, and found that the level of the strongest warm core varies between 760 and 250 hPa, with the higher warm cores more often associated with stronger TCs but with a large degree of variability. Durden (2013) also noted that Danielle (1998) and Georges (1998) had prominent double warm cores, while Earl (2010) also featured a double warm core early on. However, the lower warm core for Earl is only evident when comparing against local environmental dropsondes as a reference profile, and is no longer apparent when using the Dunion (2011) moist tropical sounding as a reference profile.

While observations of the warm core remain limited, there have been numerous modeling studies focused on the warm core. Stern and Nolan (2012) found the primary warm core to be in the 4–8-km-altitude range using idealized WRF Model simulations, in contrast to the more conventionally accepted 10 km or greater found in earlier observational studies. They also note a secondary, weaker warm core at 13–14 km in some simulations. Using idealized Nonhydrostatic Icosahedral Atmospheric Model (NICAM) simulations, Ohno and Sato (2015) found initial warm core development from 8 to 9 km. However, a secondary warm core developed around 16 km at the top of the troposphere, penetrating into the lower stratosphere, which eventually became the primary warm core as the system intensified to <900 hPa. It has been argued for some time now that dry descent in the eye of the TC contributes to intensification (e.g., Malkus and Riehl 1960). This idea has since been supported by ER-2 observations of Hurricane Bonnie in 1998 (Heymsfield et al. 2001) and a number of recent modeling studies (Chen and Zhang 2013; Miller et al. 2015; Ohno and Sato 2015). In fact, these recent modeling studies suggest that eye descent is likely to be associated with horizontal convergence above the outflow layer and may originate in the stratosphere. While the lower stratosphere is at the limits of what the GH AVAPS system is capable of sampling, the HS3 data will be examined with these hypotheses in mind.

2. Methodology

Twenty-five research flights were conducted from 2012 to 2014 during HS3, 16 of which are included in this study. This subset includes all but three TC missions of at least tropical depression (TD) status and one mission into Gabrielle (2013) prior to genesis. The three TC
missions that are not included are the 7 September 2012 flight into Hurricane Leslie during an aircraft ferry flight in which only part of the environment around the TC was sampled; the mission on 20 August 2013 that observed the remnants of Erin, in which case there was an early failure of the dropsonde system; and an attempted mission over Hurricane Ingrid on 9 September 2013 that ended prematurely because of cold fuel temperatures (Braun et al. 2016). The 16 missions analyzed in the present study span the evolution of six TCs, as well as cyclone intensification sampled six times, while five TCs of steady intensity, and five cases of weakening are captured, per the National Hurricane Center’s (NHC) second-generation NHC “best track” hurricane database (HURDAT2; Landsea and Franklin 2013). Throughout the text, the storm intensity will refer to the HURDAT2 intensity at the temporal median of the dropsonde release pattern, while intensity trend refers to the total change in intensity from the time of release of the first dropsonde to the time of release of the final dropsonde.

<table>
<thead>
<tr>
<th>Date (first dropsonde)</th>
<th>System name</th>
<th>Classification</th>
<th>Intensity (kt)</th>
<th>Trend (kt)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 Sep 2012</td>
<td>Nadine</td>
<td>TD</td>
<td>30</td>
<td>0</td>
</tr>
<tr>
<td>14 Sep 2012</td>
<td>Nadine</td>
<td>H</td>
<td>70</td>
<td>+10</td>
</tr>
<tr>
<td>19 Sep 2012</td>
<td>Nadine</td>
<td>TS</td>
<td>50</td>
<td>0</td>
</tr>
<tr>
<td>22 Sep 2012</td>
<td>Nadine</td>
<td>TS</td>
<td>45</td>
<td>+5</td>
</tr>
<tr>
<td>26 Sep 2012</td>
<td>Nadine</td>
<td>TS</td>
<td>50</td>
<td>+5</td>
</tr>
<tr>
<td>29 Aug 2013</td>
<td>Gabrielle</td>
<td>Pregenesis</td>
<td>&lt;25</td>
<td>0</td>
</tr>
<tr>
<td>4 Sep 2013</td>
<td>Gabrielle</td>
<td>TD</td>
<td>30</td>
<td>0</td>
</tr>
<tr>
<td>7 Sep 2013</td>
<td>Gabrielle</td>
<td>Disturbance</td>
<td>25</td>
<td>0</td>
</tr>
<tr>
<td>16 Sep 2013</td>
<td>Humberto</td>
<td>TS</td>
<td>35</td>
<td>-5</td>
</tr>
<tr>
<td>26 Aug 2014</td>
<td>Cristobal</td>
<td>H</td>
<td>70</td>
<td>+5</td>
</tr>
<tr>
<td>29 Aug 2014</td>
<td>Cristobal</td>
<td>H</td>
<td>70</td>
<td>-10</td>
</tr>
<tr>
<td>2 Sep 2014</td>
<td>Dolly</td>
<td>TS</td>
<td>40</td>
<td>-5</td>
</tr>
<tr>
<td>12 Sep 2014</td>
<td>Edouard</td>
<td>TS</td>
<td>35</td>
<td>+5</td>
</tr>
<tr>
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<td>Edouard</td>
<td>H</td>
<td>80</td>
<td>+15</td>
</tr>
<tr>
<td>16 Sep 2014</td>
<td>Edouard</td>
<td>H</td>
<td>85</td>
<td>-20</td>
</tr>
<tr>
<td>18 Sep 2014</td>
<td>Edouard</td>
<td>H</td>
<td>65</td>
<td>-10</td>
</tr>
</tbody>
</table>

Most systems flown during HS3 were well sampled in all four quadrants using either butterfly or lawnmower patterns [for descriptions and figures of these patterns, see Helms and Hart (2013)]. Therefore, it is possible to generate a reasonable three-dimensional representation of the TC, the outflow, and the near-storm environment from dropsondes alone. These dropsondes were released over a 12–18-h period, and occasionally even longer, for numerous missions throughout HS3. Storm motion can be substantial over these time periods, especially for fast-moving systems. As such, results will be presented in a storm-relative coordinate system computed relative to 2-min interval center fixes provided by the NOAA/Hurricane Research Division (HRD; available online at http://www.aoml.noaa.gov/hrd/data_sub/hurr.html). When the HRD centers are not available, the TC center is calculated using a linear interpolation of NHC’s 3- (when available) and 6-hourly center fixes to the nearest minute of each dropsonde release. The dropsondes’ onboard GPS is utilized to account for drift and data are averaged vertically over 5-hPa increments to reduce noise. Finally, data are interpolated onto a Cartesian grid with 25-km horizontal grid spacing using a natural-neighbor interpolation method (Sibson 1981), which can be thought of as a compromise between linear and cubic interpolation. Analyses will be presented on constant pressure surfaces, where the $x$ and $y$ coordinates refer to the distance in kilometers from the center of the TC. Radius–pressure $(r-p)$ azimuthal–mean sections, as well as three-dimensional $(x-y-p)$ plots will also be shown. For the radius–pressure plots, the horizontally interpolated data are averaged in 25-km-radius bins and plotted at the midpoint of each bin (e.g., data in the 0–25-km bin are plotted at 12.5 km).

The top of the outflow layer is typically capped by the high static stability at the tropopause level. Here, we are interested in assessing the vertical structure of the outflow relative to the tropopause. As such, static stability is quantified in terms of the lapse rate $\gamma$:

$$\gamma = \frac{dT}{dz},$$

where $T$ is the temperature in degrees Celsius and $z$ is the altitude in kilometers. It has also been established that inertial stability is relevant to various aspects of TC structure, including the behavior of the outflow. In calculating $I$ [Eq. (2)], note that it is trivial to decompose...
the wind into radial ($V_r$) and tangential ($V_t$) components once the dropsondes are adjusted to storm-relative coordinates. However, caution must be used in interpreting the results, as winds, particularly $V_t$, can be quite sensitive to small errors in storm center location [implied by Willoughby (1992) and Marks et al. (1992), quantified by Ryglicki and Hodys (2016)]. Also note that $V_r$ and $V_t$ in this study are computed with storm motion removed. Inertial instability or $I^2 < 0$ corresponds to an imbalance of pressure gradient and total centrifugal forces, and is possible in regions of negative absolute vorticity in the Northern Hemisphere or when the absolute angular momentum decreases outward, although meridional and radial mixing generally prevent this from occurring (Knox 1997). Of significance to the present study, Molinari and Vollaro (2014) found inertial instability in the outflow region of from roughly 10.5 to 12 km and from 350- to 450-km radius in an azimuthal mean of G-IV dropsondes for Hurricane Ivan (2004).

Motivated by the Davis and Ahijevych (2012) technique for estimating vertical wind shear from a set of dropsondes alone, an approximate estimate of the environmental flow is obtained by averaging dropsonde winds in each of the four quadrants around the cyclone, then averaging those four wind vectors. Averaging of the four quadrants will largely eliminate the wavenumber-0 vortex component to the wind. The advantage of this technique is that, as long as all four quadrants are sampled at fairly comparable radii, the results are relatively robust despite an unequal number of dropsondes in each quadrant.

### 3. Results

#### a. Upper-level anticyclone and outflow

First, the upper-level structure of Hurricane Nadine as sampled on 14 September 2012 is examined. Nadine was a 60-kt (1 kt = 0.5144 m s$^{-1}$) TS at the time of release of the first dropsonde, and strengthened into a 70-kt hurricane by the temporal midpoint of the dropsonde deployment (Table 1). At the time, Nadine was under a region of relatively strong (15–20 m s$^{-1}$) southwesterly upper-level environmental flow. This strong environmental flow overpowered the weak upper-level cyclonic circulation, resulting in pronounced asymmetries, including net anticyclonic tangential flow northwest of the center of the TC and inflow southwest of the storm center (Figs. 1a–d). In the composite horizontal cross sections, by convention the flow associated with the TC itself will be defined as the contiguous region of positive $V_r$ closest to the origin (0, 0), the outflow will be any contiguous region of positive $V_r$ originating within 500 km of the TC, the upper-level anticyclone will be any contiguous region of negative $V_r$ within 1000 km of the TC that exists over at least two quadrants, and the environmental flow will be computed as the sum of the mean wind computed in each of four quadrants surrounding the TC separately. For Nadine, cyclonic tangential flow becomes weaker and more spatially confined with decreasing pressure above 200 hPa. Consistent with southwesterly environmental flow, a positive radial wind maximum of 25–30 m s$^{-1}$ was observed at 200 hPa in the northeast quadrant of the storm, with net inflow of up to 20 m s$^{-1}$ in the southwest quadrant (Fig. 1a). This structure to the radial wind field is qualitatively consistent with the aforementioned structure at 150 and 130 hPa as well (Figs. 1b,c), but with winds weakening with height, while the secondary circulation appears to collapse completely by 100 hPa (Fig. 1d). Overall, the upper-level anticyclone contracts with height, with the minimum $V_r$ of less than −30 m s$^{-1}$ at 200 and 150 hPa occurring 500 km or more from the center both beyond and close to, respectively, the edge of the dropsondes, while the minimum $V_r$ appears to fall well within the dropsonde pattern at 130 hPa, closer to 400-km radius. Unfortunately most of the northern dropsondes did not begin reporting until below 100 hPa, but data from 120–110 hPa (not shown) suggest that the anticyclone continues to contract above 130 hPa. Whether or not the contraction of the radius of minimum $V_r$ is related to the decrease in magnitude and area of positive $V_r$ with height is yet to be determined. Additional cases will be examined in order to determine the variability in the rate of contraction of the upper-level anticyclone with height among the HS3 dataset.

For other hurricanes sampled during HS3, including 80-kt Hurricane Edouard on 14 September 2014 (Figs. 2a–d) and 70-kt Hurricane Cristobal on 26 August 2014 (Figs. 3a,b), the region of $V_t = 0$, or the interface between cyclonic and anticyclonic flow, clearly contracts with height. However, there are not enough data at large radii to convincingly determine whether or not the minimum in $V_t$ also contracts with height for either Cristobal or Edouard. In the case of Edouard, outflow is maximized along a single outflow channel to the northwest that is strongest at 200 hPa and decreasing but of the same direction there above. Similar to Nadine, the strongest outflow occurs downwind of the environmental flow at that level. In the case of Cristobal, there are

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3 Note that an open circle without a wind barb indicates the location of a dropsonde but with no or missing data at that level.
two partially connected outflow channels: one to the northeast and one to the south. Within the sampled region, the northeast channel is maximized at 180 hPa (Fig. 3a) while the one to the south is strongest at 130 hPa (Fig. 3c). The strongest outflow from 200 to 150 hPa exists almost due downwind of the southwest-erly environmental flow, while the outflow maximum at 130 hPa is actually upwind. Consistent with the results of Merrill and Velden (1996), the jet maximum of the southern outflow channel occurs at higher altitude than that of the northern channel, with stronger radial velocities in the northern outflow channel. It was initially hypothesized that the higher outflow to the south would be associated with a higher tropopause. The data reveal that the cold-point tropopause is indeed slightly colder on average to the south (Fig. 3c). However, and somewhat unexpectedly, tropopause pressures are roughly equal north and south of the TC (Fig. 3d). Examination of individual dropsondes reveals the cold point to be slightly “sharper” south of the TC but of roughly the same level as to the north, explaining the discrepancy (not shown). However, why this difference in vertical structure just below the tropopause is observed in the dropsondes remains elusive at this time. Note that while Cristobal (2014) was an intensifying TC at the time of development of dual outflow channels, there were insufficient samples to relate the development of multiple outflow channels to intensification.

Although the tropopause does not appear to be meaningfully higher to the south compared to the north of Cristobal, there does appear to be a slight upward bulge of the tropopause, as indicated by the overall lower tropopause pressures directly above the TC (Fig. 3d). To systematically investigate whether or not there is an upward bulge in the tropopause across the broader HS3 sample, the difference between the mean 0–100-km radius and the 400–600-km-radius tropopause pressure results is calculated and plotted versus storm intensity (Fig. 4a). The results appear to be mixed, as there are eight cases where the tropopause is elevated to lower pressures above the TC, while in the other eight cases the tropopause actually

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**Fig. 1.** Tangential wind (shaded, 5 m s$^{-1}$ increments) and radial wind (positive, contoured; negative, dashed; 5 m s$^{-1}$ increments) for 70-kt Hurricane Nadine on 14 Sep 2012. Analyses are at (a) 200, (b) 150, (c) 130, and (d) 100 hPa in storm-relative coordinates (km from center in the x and y directions). Gray dots indicate locations of dropsondes.
dips downward to higher pressures above the cyclone. However, there appears to be a greater occurrence of having an elevated tropopause above the TC for stronger storms, for which the sample is unfortunately quite small. Five out of six TCs of 65 kt or greater intensity are associated with an elevated tropopause above the TC, with the greatest pressure difference of $-12 \text{hPa}$ for 85-kt Hurricane Edouard, the strongest storm in the sample. Overall, the tropopause is also found to be somewhat warmer (colder) above the TC core for most stronger (weaker) storms in the sample, but with a large degree of variability (Fig. 4b).

It is also critical to examine the divergence field in order to gain a more complete understanding of the upper-level anticyclone, since divergence can generate negative $\zeta$ via the stretching term [Eq. (1), which may ultimately act to build or strengthen a preexisting upper-level anticyclone. Upper-level divergence at 180 hPa is examined for four individual cases: 70-kt Hurricane Nadine on 14 September 2012 (Fig. 5a), 80-kt Hurricane Edouard on 14 September 2014 (Fig. 5b), 70-kt Hurricane Cristobal on 29 August 2014 (Fig. 5c), and 85-kt Hurricane Edouard on 16 September 2014 (Fig. 5d). These cases are chosen as they allow us to compare two intensifying TCs to two weakening TCs, all of similar intensities. Note that the uncertainty in the divergence calculation is somewhat greater for Cristobal on 29 August because of a more rapid forward motion versus the other TCs in the sample. For the Nadine, Cristobal, and the earlier Edouard analyses, it is evident that the region of strongest divergence is downstream of the environmental flow and, perhaps, slightly left of downstream for Edouard and Cristobal. Since the upper-level flow vector is roughly consistent with the shear vector for each of these three cases (not shown), this observation is consistent with having the strongest convection developing downshear left. Also apparent is the fact that the two intensifying systems, Nadine on 14 September 2012 and Edouard on 14 September 2014, are both associated with much stronger upper-level divergence than either of the two weakening TCs, despite the fact that these systems are all of similar intensities.

To further investigate the relationship between the strength of upper-level divergence and both present intensity and intensity change, data are binned for each of four categories: hurricanes (Fig. 6a), tropical storms and tropical depressions (Fig. 6b), intensifying TCs
(Fig. 6c), and nonintensifying and weakening TCs (Fig. 6d). For each category, 300–100-hPa-layer mean upper-level divergence is computed for each case individually; the results are then averaged over all cases in each bin. Note that there are only 6 cases in both the hurricanes and intensifying samples (but not the same set of six cases) while there are 10 in both the weak TC and nonintensifying samples. Clearly, hurricanes are associated, on average, with stronger upper-level divergence over a larger area close to the circulation center (at the origin) than TDs and TSs. The association between upper-level divergence and intensity change also appears to be robust, as intensifying systems in the HS3 dataset are overall associated with stronger upper-level divergence than weakening or steady systems, regardless of current intensity. Additionally, stronger divergence tends to be vertically collocated with the center of circulation for intensifying systems, while it is displaced well to the northeast in the nonintensifying composite. This displacement of the divergence implies that the convection is also significantly displaced from the center of circulation, and is consistent with strong westerly or southwesterly shear in a large number of nonintensifying cases sampled during HS3.

b. Azimuthally averaged cross sections

Azimuthally averaged cross sections in radius–pressure coordinates are examined to better elucidate the mean structure of the sampled TCs. We begin examining the primary circulation by computing composites of $V_t$ for hurricanes (Fig. 7a), TDs and TSs (Fig. 7b), intensifying TCs (Fig. 7c), and nonintensifying and weakening TCs (Fig. 7d). Recall that data are first interpolated onto a storm-relative Cartesian grid on pressure levels, then averaged in 25-km-radius bins and plotted at the midpoint of each bin. The number of dropsondes in each bin is indicated in the figures. Data are only plotted out to 500-km radius, as the HS3 flight patterns became quite asymmetric about the TC at
larger radii. The mean $V_r$ at the radius of maximum winds (RMW) is significantly stronger for hurricanes than for TDs and TSs, as expected. More notably, the RMW is much larger for TDs and TSs than for hurricanes, likely because of the fact that both early developing TCs and TCs late in their life cycle undergoing extratropical transition tend to be associated with broader wind fields than mature TCs. While there are comparatively fewer dropsondes in the two lowest-radii bins than there are at larger radii in the TD and TS composite, a complete lack of inner-core data cannot be to blame for the unusually large RMW. Negative $V_r$ at upper levels is also notably stronger in the hurricane sample, associated with a stronger upper-level anticyclone. This observation could be due to either the anticyclone being stronger or being more vertically collocated with the TC in the hurricane composite, or both. When comparing the intensifying to the non-intensifying composite $V_r$, it is apparent that intensifying systems tend to be associated with a smaller RMW, stronger cyclonic $V_r$ at the RMW, and a deeper vortex. However, in terms of negative $V_r$, the strength of the upper-level anticyclone appears to be a function of the strength of the vortex itself and not a function of intensity change. Intensifying cyclones include weaker systems as well as some stronger TCs, so both cyclones and anticyclones are weaker than in the hurricane composite. Weaker but intensifying TCs are not necessarily in an environment that precludes the development of a stronger anticyclone, but the storm needs time to intensify long enough to realize that anticyclone.

Next, the structure of the secondary circulation is examined by computing the azimuthally averaged radial wind $V_r$. Four representative cases are chosen to demonstrate both the degree of variability within the HS3 sample as well as some commonalities across differing scenarios. These cases include steady-state 50-kt TS Nadine on 19 September 2012, intensifying 70-kt Hurricane Cristobal on 26 August 2014, intensifying 80-kt Hurricane Edouard on 14 September 2014, and weakening 85-kt Hurricane Edouard on 16 September 2014. The secondary circulation is evident in all four representative cases, with the strongest inflow at the low levels maximized from 1000 to 950 hPa and the strongest outflow between 300 and 100 hPa. However, the magnitude of the upper-level outflow is notably weaker, particularly at small radii for Cristobal and at all radii for Nadine when compared against the two Edouard cases. For all four cases, vertical gradients in $V_r$ are quite sharp, sharper than gradients in $V_r$ in the composites (Fig. 7) or even when comparing individual cases (not shown). A small inflow layer of $-2$ to $-4 \text{ m s}^{-1}$ and 20–50 hPa deep is also observed 50–100 hPa above the region of maximum outflow in all four cases. It is hypothesized that this inflow is associated with dry-adiabatic descent above the eye in the lower stratosphere, resulting in a small reverse secondary circulation in which the primary outflow layer also serves as the “out” component of an “in–down–out” circulation pattern. If this were the case, one might expect to observe a lowering of the tropopause above the eye. However, this result is not inconsistent with the earlier finding of the tropopause being elevated above the core of stronger TCs (Fig. 4a), as it is possible for there to be a net increase in height of the tropopause above the cyclone core along with a simultaneous highly localized decrease in the height of the tropopause pressure difference (hPa) between the mean from $r = 0$ to 100 km and the mean from $r = 400$ to 600 km, with negative values indicating a higher tropopause over the TC center, and (b) the cold-point tropopause temperature difference ($^\circ\text{C}$) between the mean from $r = 0$ to 100 km and the mean from $r = 400$ to 600 km, with negative values indicating a colder tropopause over the TC center.

FIG. 4. Scatterplot of azimuthal-mean quantities for all cases with a current intensity $\geq 30$ kt (see Table 1). Included are (a) the cold-point tropopause pressure difference (hPa) between the mean from $r = 0$ to 100 km and the mean from $r = 400$ to 600 km, with negative values indicating a higher tropopause over the TC center, and (b) the cold-point tropopause temperature difference ($^\circ\text{C}$) between the mean from $r = 0$ to 100 km and the mean from $r = 400$ to 600 km, with negative values indicating a colder tropopause over the TC center.
tropopause directly above the eye. Warming from adiabatic descent above the eye is also consistent with the earlier finding of having a slightly warmer tropopause directly above the TC for stronger storms (Fig. 4b).

Multiple inflow layers are also observed below the outflow channel, generally \(-2 \text{ m s}^{-1}\) or weaker. We originally speculated that the rightward deflection of radial outflow by Coriolis would produce a minimum in \(V_{t}\) at the same level as the maximum \(V_{r}\). Somewhat unexpectedly, for most cases the level of maximum \(V_{r}\) is found to occur 40–80 hPa below the level of minimum \(V_{t}\). The level of maximum \(V_{r}\) is still below the minimum \(V_{t}\) even when plotted on theta surfaces (not shown), so this result is not simply a function of theta surfaces sloping downward with increasing radius. It is unclear whether this discrepancy is due to having a preexisting upper-level anticyclone prior to TC genesis and/or intensification of a synoptic-scale anticyclone that exists above the level of maximum outflow, a decrease in height of the level of maximum radial outflow with time resulting in a “remnant” diabatically generated remnant upper-level anticyclone as the outflow descends, or some other process.

Composite of \(V_{r}\), inertial stability, and \(\Gamma\) in radius–pressure coordinates for all intensifying (Figs. 8a,b) and nonintensifying and weakening (Figs. 8c,d) TCs are also examined. The level of maximum \(V_{r}\) is fairly consistent between intensifying (Fig. 8a) and nonintensifying (Fig. 8c) TCs, with a peak between 200 and 150 hPa in the means. Among the full set of individual cases, there does not appear to be a meaningful relationship between the level of strongest radially averaged positive \(V_{r}\) (computed as the level at which the strongest radially averaged positive \(V_{r}\) from 100 to 500 km occurs) and the current intensity, regardless of whether this value is computed as a function of pressure (Fig. 9a) or \(\theta\) (Fig. 9c). However, there is an overall slight tendency for intensifying TCs to have a higher level of peak azimuthally averaged outflow compared to weakening TCs, but with very large variability (Fig. 9b). Upper-level outflow is overall much stronger in the intensifying than the nonintensifying composite at virtually all radii: 5–7 m s\(^{-1}\) at the maximum for intensifying cases versus 2–4 m s\(^{-1}\) for nonintensifying and weakening cases, which would suggest via the continuity equation stronger...
vertical motion and thereby a stronger secondary circulation (Figs. 8a,c). This result is consistent with the earlier finding that most intensifying systems are associated with stronger upper-level divergence than non-intensifying systems. A layer of inflow \(-1 \text{ to } -2 \text{ m s}^{-1}\) is also evident above the outflow in both composites from 125 to 100 hPa, although subtle differences in the exact height of this feature between various cases cause it to appear to be weaker in the mean than what was observed in the individual cases (Fig. 10). Interestingly, this inflow is slightly stronger in the nonintensifying composite at radii \(\geq 200 \text{ km}\), but is approximately equal between the two samples within 200 km of the center.

A meaningful relationship between the \(\theta_e\) of the maximum 100–500-km radially averaged \(V_r\) and the present intensity was not observed (Fig. 9c). However, a notable positive relationship between the \(\theta_e\) of the outflow layer and the \(\theta_e\) of the boundary layer inflow, which is a proxy for low-level moisture and SSTs, does occur (Fig. 9d). These values are obtained by first computing azimuthal-mean \(V_r\) in radius–\(\theta_e\) (radius \(\theta_e\)) coordinates, averaging \(V_r\) from 100- to 500-km radius, and finding the \(\theta(\theta_e)\) level corresponding to the maximum (minimum) value of this radially averaged \(V_r\). While one might expect the \(\theta_e\) of the inflow to be directly proportional to storm intensity, the notable differences between these two figures suggest that this is not always the case. This may have further implications for the TC potential intensity (Emanuel 1986).

The intensifying composite features a sharp layer of greater static stability at lower levels in the inner core in a region of \(\sim 2^\circ\text{–}3^\circ \text{C km}^{-1}\) lapse rates from 850 to 900 hPa (Fig. 8b). In contrast, the nonintensifying composite is associated with a deeper but less well-defined layer of greater static stability of \(\sim 3^\circ\text{–}4^\circ \text{C km}^{-1}\) from 950 to 800 hPa (Fig. 8d). As will be shown in section 3c, subtle differences between the two stability profiles at low radii are likely a consequence of differences in the vertical extent of the warm core. Note that the top of the outflow

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4 While \(\theta_e\) of the inflow continues to increase with decreasing radius and should better match the actual values of outflow \(\theta_e\), inflow becomes increasingly noisy inside of 100-km radius in the individual cases, with distinct local maxima that occur on \(\theta_e\) levels that differ by as much as 5 K for a single case. Because of large uncertainty in the 0–100-km \(V_r\), 100–500-km \(V_r\) is depicted instead.

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**Fig. 6.** Composites of 300–100-hPa layer-average divergence in storm-relative coordinates for all cases of (a) hurricane intensity, (b) TS or TD intensity, or for cases that are (c) intensifying or (d) nonintensifying or weakening. The number of cases in each sample \(n\) is indicated.
layer (Figs. 8a,c) is constrained in the vertical by the very high static stability at the tropopause, indicated by a lapse rate that approaches zero. Greater static stability above the outflow is consistent with a sharper vertical gradient in $V_r$ above rather than below the peak outflow. For the intensifying (nonintensifying) composite, the radial outflow is found to originate radially just inside (outside) of the region of higher inertial stability values of $1 \times 10^{-4}$ s$^{-1}$, but the bulk of the outflow is primarily outside of this region. This region, previously referred to as the outflow roots, is found from approximately 50- to 200-km radius between 250 and 150 hPa. For larger (smaller) systems, the outflow roots will be of slightly larger (smaller) radius. In the composites, the radius at which $V_r$ exceeds 1 m s$^{-1}$ is greater in the nonintensifying than the intensifying composite (Figs. 8a,c), consistent with a larger mean RMW (Figs. 7a,c). With higher values of $I$ at lower levels or closer to the core, the recirculating wind is too strong for any appreciable $V_r$ to develop, as described by Schubert and Hack (1982), Shapiro and Willoughby (1982), and Hack and Schubert (1986). We do not find an inertially unstable ($I < 0$) region in either composite. However, consistent with Molinari and Vollaro (2014), $V_r$ continues to increase with radius just below the tropopause as $I$ continues to decrease with radius. Unfortunately, the data coverage at radii $\gtrsim$500 km here does not allow us to compare the radius of minimum $I$ to the radius of maximum $V_r$.

To better conceptualize the mean vertical structure and variability of the secondary circulation, the 100–500-km mean $V_r$ is plotted as a function of height with the $\pm 1 \sigma$ region shaded, comparing hurricanes to TDs and TSs (Fig. 11a) and intensifying to nonintensifying and weakening TCs (Fig. 11b). The statistical significance of the difference between the two samples is computed at the 5% level using a two-sample Welch’s $t$ test in 5-hPa increments. While the mean outflow from 250 to 150 hPa is 1.5–2 m s$^{-1}$ stronger for hurricanes than for TDs and TSs (Fig. 11a) and intensifying to nonintensifying and weakening TCs (Fig. 11b). The statistical significance of the difference between the two samples is computed at the 5% level using a two-sample Welch’s $t$ test in 5-hPa increments. While the mean outflow from 250 to 150 hPa is 1.5–2 m s$^{-1}$ stronger for hurricanes than for TDs and TSs (Fig. 11a) and intensifying to nonintensifying and weakening TCs (Fig. 11b). The statistical significance of the difference between the two samples is computed at the 5% level using a two-sample Welch’s $t$ test in 5-hPa increments. While the mean outflow from 250 to 150 hPa is 1.5–2 m s$^{-1}$ stronger for hurricanes than for TDs and TSs (Fig. 11a).
the intensity change, but once again with large variability (Fig. 11d). Nonetheless, intensifying TCs are associated with 2–3 m s$^{-1}$ stronger $V_r$ than nonintensifying TCs in the 250–150-hPa layer, which is statistically significant in the 160–210-hPa layer (Fig. 11b). One notable outlier in the scatter of individual cases is 85-kt Hurricane Edouard on 16 September 2014, which exhibited relatively strong radial outflow despite having weakened by 20 kt throughout the dropsonde release period. Therefore, it is clear that present intensity also remains a factor, but these data suggest that perhaps the intensity change is the more significant of the two.

c. Height and structure of the warm core

During HS3, the Global Hawk often made multiple transects over the TC core while releasing dropsondes from the lower stratosphere. This sampling allowed for unprecedented sampling of the warm core through the entire depth of the cyclone. To first gain an understanding of both the commonalities and the variability in the magnitude and structure of the warm core within the dataset, we will first examine the warm core in binned radius–pressure azimuthal-mean cross sections for a subset of six HS3 flights (Fig. 12). These cases include 50-kt TS Nadine on 19 September 2012, 45-kt TS Nadine on 22 September 2012, 35-kt TS Humberto on 16 September 2013, 70-kt Hurricane Cristobal on 26 August 2014, 80-kt Hurricane Edouard on 14 September 2014, and 85-kt Hurricane Edouard on 16 September 2014. The potential temperature anomaly is computed with respect to the mean $\theta$ profile of the interpolated data of radius $500 \leq r \leq 1500$ km in each mission. For both Nadine flights (Figs. 12a,b), both Edouard missions (Figs. 12e,f), and Cristobal (Fig. 12d), the maximum $\theta$ anomaly occurs very near 300 hPa. For Humberto, there are unfortunately no dropsondes within a radius of 25 km from the center of the TC.

5 Occasionally during HS3, a few dropsondes were deployed well outside of the target TC for a variety of reasons, such as to test the dropsonde system early in the flight.
However, the natural-neighbor interpolation allows for estimation of the local temperature maximum based upon the horizontal temperature gradient (Fig. 12c), and the storm had a broad circulation pattern with a large RMW, so a lack of data near $r = 0$ km should not be a problem. This analysis suggests a distinct double-warm-core structure, with one maximum around 200 hPa and a second maximum around 800 hPa, as well as negative $u$ anomalies between these two maxima. However, further analysis suggests that this pseudo-upper-level warm core is actually a stratospheric intrusion, and the actual upper-level vortex was associated with the observed cold core. As a result of this hybrid structure, Humberto is not representative of a typical, fully tropical TC. Nadine on 22 September 2012 also features two different maxima in $\theta$, with the upper-level warm core the stronger of the two with a 7-K anomaly versus a 5-K anomaly associated with the lower maximum. Nadine’s upper-level warm core is also deeper in the vertical than the lower warm anomaly. However, there were potential issues in the sampling of the environment around Nadine. On 22 September, Nadine existed in a region of significant northwest (NW)-to-southeast (SE) baroclinicity, with colder air to the NW where many of the “reference” dropsondes were deployed. While there is still evidence of a weak low-level warm core, the low-level warm anomaly associated with Nadine likely results from baroclinicity and the asymmetric sampling of the environment. Finally, the warm core of Edouard has become stronger at virtually all levels between 14 and 16 September 2014, with the strength of the warm core increasing from 10 to 15 K, primarily as a result of intensification. The level of the peak warm anomaly remains unchanged. Warming on 16 September at larger radii (300–500 km) is likely an effect of the TC also moving northward into an environment with a cooler reference profile, which also contributes to some additional “apparent” warming of the core. However, this is believed to be a smaller factor than the warming due to intensification.

**Fig. 9.** Pressure (hPa) of the maximum azimuthal-mean radial wind averaged from 100- to 500-km radius for each case, plotted as a function of (a) intensity (kt) and (b) intensity change. Also included are the mean potential temperature ($\theta$, K) of the level of strongest azimuthally averaged outflow plotted as a function of (c) intensity (kt), and (d) the mean equivalent potential temperature ($\theta_e$, K) of the level of strongest 100–500-km radius-mean azimuthally averaged radial inflow.
Upon examination of the entire HS3 sample of TCs with a present intensity of $\geq 30$ kt, there is a very evident positive relationship between the maximum warm core temperature anomaly (regardless of height) and the intensity (Fig. 13a), consistent with the literature (e.g., Durden 2013). The pressure at which the maximum temperature anomaly occurs does not appear to be strongly correlated with intensity (Fig. 13c). However, while having a warm core above 500 hPa is equally likely as having one below 500 hPa for TCs $\leq 45$ kt, all but one of the TCs in the sample $>45$ kt have a warm core between 350 and 250 hPa. Neither the magnitude (Fig. 13b) nor the pressure (Fig. 13d) of the warm core shows any meaningful correlation with the intensity change. No evidence was found for a relationship between the presence of a single versus a double warm core and storm intensity or intensity change.

Finally, the 0–25-km radial-mean temperature anomaly and standard deviation as a function of pressure is examined. These data originate from a combination of 18 dropsondes across HS3 within 25 km of the center of circulation and the interpolated values. Differences between the hurricane category and the TD and TS category are statistically significant through a deep layer from just above 600 hPa through slightly above 200 hPa (Fig. 14a). The hurricane warm core was, on average, approximately 4–5 K warmer than the mean for TDs and TSs in the HS3 sample. The greatest variability in the strength of the warm core for weaker systems occurs at and just below the level of the greatest warm anomaly, with a secondary peak in variability at lower levels from 950 to 800 hPa and less variability at middle levels. The larger standard deviation observed at low levels appears to be associated with some TDs and TSs having a secondary low-level warm core, while others do not have this feature at all. For hurricanes, the standard deviation is fairly uniform throughout the column but greatest from approximately 850 to 550 hPa. In this region, the larger variance appears to be a function of how far
downward the primary warm core extends below the maximum. While there is much greater variability in the height and strength of the warm core for nonintensifying and weakening TCs than there is for strengthening TCs, the differences in the means are not statistically significant at any level (Fig. 14b). Note, however, that there is a greater downward extension of the primary warm core in the nonintensifying than in the intensifying composite. This structure would result in the greater static stability observed in the nonintensifying composite near 800 hPa in section 3b.

4. Discussion and conclusions

In this study, storm-relative constant pressure, azimuthal-mean radius height, and three-dimensional analyses of 16 HS3 missions investigating six different TCs are examined. This unique dataset has made it possible to develop unprecedented analyses of both TC outflow and the warm core structure entirely from dropsonde data. The primary focus of this study is on the strength, height, and structure of the TC outflow, and how both the commonalities and the variability within the sample relate to either the current TC intensity or the intensity change. The height and structure of the warm core and its relationship to intensity and intensity change are also examined.

Overall, hurricanes in the HS3 sample tend to be associated with stronger upper-level divergence than TDs and TSs. Perhaps more intriguingly, stronger upper-level divergence is found for intensifying rather than nonintensifying and weakening systems regardless of current intensity. The divergence also tends to be more concentrated close to the center of circulation for intensifying rather than weakening TCs. Similarly, azimuthal-mean radius–pressure cross sections reveal stronger radial wind in the outflow region for intensifying than nonintensifying TCs, with statistical significance, throughout the 250–150-hPa layer. This finding is consistent with the results of Merrill (1988a). A weaker but nonetheless positive relationship between the strength of the outflow and present storm intensity was also observed. The strongest
outflow is generally found to occur from 50 to 100 hPa below the tropopause, with virtually all the outflow occurring within 150 hPa of the tropopause. The pressure at which the outflow maximum occurs is also found to be loosely correlated with intensity change, where intensifying systems often have their strongest outflow at higher levels. Additionally, while the $\theta$ of the outflow does not appear to be strongly tied to the intensity, the $\theta$ of the outflow exhibits a positive relationship with the mean $\theta_e$ of the level of strongest low-level inflow. This result is somewhat unexpected, as the low-level $\theta_e$ is expected to be roughly proportional to TC intensity.
Both $V_r$ and $V_t$ are associated with sharp vertical gradients in the outflow layer, although the gradient is found to be sharper for $V_r$ than $V_t$, especially above the level of maximum $V_r$. This result is believed to be a consequence of the greater static stability at the tropopause above the outflow region. The level of minimum $V_t$ associated with the upper-level anticyclone is also found to be notably (~40–80 hPa) above the level of maximum $V_r$, suggesting some disconnect between the height of the upper-level anticyclone at large radius and divergence near the outflow roots at small radius. This finding could possibly be due to the presence of a pre-existing upper-level anticyclone above the primary outflow layer collocated with the $T$; a lowering of the level of maximum radial outflow with time, which leaves a remnant diabatically induced upper-level anticyclone above; or some combination of the two. Since $\theta$ surfaces slope upward with increasing radius at the upper levels away from the TC for most cases (Fig. 12), part of this phenomenon could also be the result of isentropic outflow rising to lower pressure levels with increasing radius, and deflection to the right due to Coriolis, although this effect does not appear to be sufficient to fully explain the discrepancy. Finally, the strength of the upper-level anticyclone does not appear to be a discriminating factor between intensifying and nonintensifying TCs.

In this study, an attempt has been made to quantify a range of pressure levels and a range of radii from which the outflow originates. This outflow roots region was also quantified in terms of static and inertial stability. Of course, the interaction between the radial and tangential flow is a complex issue. Positive $V_r$ tends to be associated with divergence above the TC, which in turn generates negative vorticity. On the other hand, stronger swirling winds associated with greater inertial stability should delay the outward acceleration of air parcels until they reach sufficient distance from the primary circulation. This effect is why regions of reduced inertial stability in the environment of the TC tend to promote enhanced

Fig. 13. Scatterplots of (a),(b) the warm core temperature anomaly (°C) and (c),(d) the pressure at which the maximum temperature anomaly occurs (hPa) vs (a),(c) the current storm intensity (kt) and (b),(d) the 12-h intensity change (kt). The warm core is computed as the difference between the radial average of the horizontally interpolated data between 0 and 25 km and radius 500 ≤ r ≤ 1500 km.
outflow in the direction of the inertial stability minimum (e.g., Rappin et al. 2011). Therefore, it is necessary to quantify the inertial stability to adequately describe the secondary circulation as a whole. Consistent with Molinari and Vollaro (2014), greater Vr tends to occur at higher levels (but below the tropopause) and larger radii where values of I are lower. From these data, the outflow roots region in which the azimuthal-mean Vr exceeds 1 m s$^{-1}$ occurs for I values close to or slightly above 1 $\times$ 10$^{-4}$ s$^{-1}$, which exists in a region from 300 to 150 hPa and from 50- to 200-km radius in the HS3 mean.

One or more inflow layers above and below the outflow channel are also observed in several cases, with peak negative Vr of $-2$ to $-4$ m s$^{-1}$ over a shallow layer 20–50 hPa deep. The highest observed inflow occurs 50–100 hPa above the level of strongest outflow. This inflow is believed to be associated with a “reverse” secondary circulation pattern, associated with dry-adiabatic descent above the eye in the lower stratosphere. This upper-level inflow layer is also evident in both intensifying and non-intensifying composites from 125 to 100 hPa, although subtle differences in the exact height of this feature between various cases cause it to appear to be weaker in the mean than what was observed in the individual cases. Interestingly, this inflow is slightly stronger in the non-intensifying composite at radii $\geq$200 km, but is approximately equal between the two samples within 100 km of the center. This finding is generally inconsistent with the idea that upper-tropospheric (Heymsfield et al. 2001) or lower-stratospheric (Ohno and Satoh 2015; Miller et al. 2015) descent radially inward of deep convective bursts along the eyewall should be more characteristic of intensifying, if not rapidly intensifying, TCs, and not steady-state or weakening TCs.

It is unclear from this dataset whether multiple outflow channels are correlated with TC intensification. The only case of dual outflow channels developing for an intensifying TC during HS3 occurred with Hurricane Cristobal (2014); however, the first flight into Cristobal occurred near the end of the intensification phase. Additionally, six other cases of intensification in the dataset occurred in the presence of only a single outflow channel. The northern outflow channel for Cristobal is at its strongest at 180 hPa, while the southern outflow channel peaks near 130 hPa. This discrepancy was originally hypothesized to be associated with a downward-sloping tropopause with latitude, but while the tropopause cools from north to south in the environment surrounding Cristobal, the tropopause is not correspondingly higher south of the TC. However, the tropopause was found to be systematically higher above the TC for five out of six TCs of 65-kt intensity or stronger, with a tropopause pressure 12 hPa lower than the surrounding environment above 85-kt Hurricane Edouard on 16 September 2014, the strongest TC in the sample.

Finally, stronger TCs are found to be associated with stronger warm cores, with a maximum $\theta$ anomaly of $>14$ K at 260 hPa for 85-kt Hurricane Edouard and a warm anomaly of at least 6.5 K for all TCs of $\geq$65 kt. The height of the primary warm core was found to occur between 250 and 350 hPa for more than half of the cases, but was found as low as 850–700 hPa for a few weaker systems. Stronger TCs were more consistently associated with higher warm cores than weaker TCs, in agreement with Durden (2013) and

![Fig. 14. Potential temperature (K) anomalies (solid lines) and $\pm 1$ standard deviation region (shaded) of the inner-most 0–25-km radius bin as a function of pressure (hPa), comparing (a) hurricanes (blue) to TDs and TSs (red), and (b) intensifying (green) to nonintensifying and weakening (red) TCs. Small circles denote values in which the difference between the two subsets of data is significant at the 5% level using a two-sample Welch's $t$ test.](image-url)
Ohno and Satoh (2015), although several TDs and TSs also had warm cores at or above 300 hPa. The fact that the warm core occurred at 350 hPa or above for all systems of 70-kt intensity or above is somewhat in disagreement with the midlevel warm core suggested by Stern and Nolan (2012). While Halverson et al. (2006) suggest that the TC weakening phase may be associated with a lowering of the warm core, we do not find a relationship between the height (or strength) of the warm core and the intensity change. Several TCs exhibited double-warm-core structures with a large degree of variability. TS Nadine on 22 September 2012 featured a weaker secondary warm core around 850–900 hPa. TS Humberto was associated with a cold-core-over-warm-core hybrid structure, with the primary warm core around 800 hPa. An upper-level warm anomaly at 200 hPa of comparable magnitude to the low-level warm core was also observed for Humberto. However, this feature appears to be associated with a lowering of the stratosphere and not a tropospheric warm core. Finally, the warm core associated with Edouard strengthened at virtually all levels between 14 September 2014 and 16 September 2016 as the TC first strengthened from 80 to 105 kt, then weakened to 85 kt. The level of the warm core remained steady close to 300 hPa and the overall structure did not change during this period.

Overall, these results suggest that there is significant variability as a function of height as well as significant case-to-case variability in terms of the structure of the tropical cyclone outflow layer and the height and magnitude of the warm core. High-altitude dropsonde observations from HS3 made it possible to observe and document these features in greater detail and over a greater number of cases than ever before. As always, the results presented in this paper should be reevaluated as additional in situ data in the upper levels of the TC become available.

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