A 13-Year Global Climatology of Tropical Cyclone Warm-Core Structures from AIRS Data

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ABSTRACT
There is uncertainty as to whether the typical warm-core structure of tropical cyclones (TCs) is featured as an upper-level warm core or not. It has been hypothesized that data from the satellite-borne Advanced Microwave Sounding Unit (AMSU) are inadequate to resolve a realistic TC warm-core structure. This study first evaluates 13 years of Atmospheric Infrared Sounder (AIRS) temperature retrieval against recent dropsonde measurements in TCs. AIRS can resolve the TC warm-core structure well, comparable to the dropsonde observations, although the AMSU-A retrievals fail to do so. Using 13-yr AIRS data in global TCs, a global climatology of the TC warm-core structure is generated in this study. The typical warm-core height is at the upper level around 300–400 hPa for all TCs and increases with TC intensity: 400 hPa (~8 km) for tropical storms, 300 hPa (~10 km) for category 1–3 hurricanes, 250–300 hPa (~10–11 km) for category 4 hurricanes, and 150 hPa (~14 km) for category 5 hurricanes. The range of warm-core height varies with TC intensity as well. A strong correlation between TC intensity and warm-core strength is found. A weaker but still significant correlation between TC intensity and warm-core height is also found.

1. Introduction
The warm-core structure is a fundamental feature of mature TCs that is frequently used to distinguish TCs from extratropical cyclones. The strength and height are the two most common variables characterizing the warm core. The strength of the warm core is usually defined by the magnitude of the maximum temperature anomaly, while the height is the level where the peak temperature anomaly is found. Previous work using the satellite-borne Advanced Microwave Sounding Unit (AMSU; Kidder et al. 2000), the Advanced Technology Microwave Sounder (ATMS; Zhu and Weng 2013), and dropsonde observations (Durden 2013) have found that the strength of the warm core generally increases with the TC intensity. However, these studies mainly focused on a limited number of cases. The extent of the correlation between the warm-core strength and TC intensity and how it varies in different geographical locations and under different environmental conditions are still open questions, especially when TC intensity is defined as the maximum sustained wind.

Until recently, it was widely believed that the typical warm-core structure is characterized by an upper-level peak (Durden 2013; Duran and Molinari 2018), and the height of warm core is positively correlated to TC intensity in terms of minimum sea level pressure (Zhang and Chen 2012; Durden 2013) and maximum sustained wind speed (Zhang et al. 2015; Gao et al. 2017). This is consistent with previous studies that demonstrated theoretically the importance of an upper-level warm core in mature (Emanuel 1986) and rapidly intensifying TCs (Zhang and Chen 2012), respectively. Through their simulations of Hurricane Wilma (2005), Chen and Zhang (2013) showed the generation of the upper-level warm core by convective bursts occurring in the eyewall. However, this conventional view was challenged by Stern and Nolan (2012, hereafter SN12). As SN12
argued, much of our understanding of the TC warm core is from one or more of three case studies using aircraft data from several decades ago (La Seur and Hawkins 1963; Hawkins and Rubsam 1968; Hawkins and Imbembo 1976). Using flight-level observations at multiple levels for Hurricanes Cleo (1958; La Seur and Hawkins 1963), Hilda (1964; Hawkins and Rubsam 1968), and Inez (1966; Hawkins and Imbembo 1976), these three studies found that the maximum warm-core anomaly was near 250–300 hPa. A second midlevel (600–650 hPa) maximum comparable in strength to the observed upper-level warm core was found for Hurricane Inez (1966) on the first flight day of 27 September. As Inez intensified rapidly on the second flight day of 28 September, the upper-level warm core deepened and became primary and the height rose from near 300 to near 240 hPa. Based on their idealized simulations, SN12 challenged the traditional view and argued that the “typical” warm-core structure is actually not well known. Although an upper-level maximum temperature anomaly was found in many of their simulations, all of them exhibited a midlevel temperature anomaly of relatively higher magnitude. This seems consistent with the midlevel warm core found by Halverson et al. (2006) for Hurricane Erin (2001) using dropsonde data. SN12 also argued that the height of warm core is not directly related to storm intensity.

Motivated by SN12’s work, Durden (2013) examined 27 TC eye soundings sampled by a combination of aircraft-deployed dropsondes and surface-based radiosondes. He found that the height of the maximum warm anomaly varies between 760 and 250 hPa and is positively related to TC intensity. But as stated by Durden (2013), the number of cases is still limited and more sounding data are needed to draw robust conclusions. Long-term satellite observations should provide some hope in this context.

Knaff et al. (2004, hereafter KSDD) pioneered a statistical study of the TC warm core using AMSU-derived temperature profiles. An upper-level warm core was found to be located around 12 km. When referring to the warm-core height, some studies used the pressure level while others used the geopotential height. To easily compare among different studies, a general conversion between pressure levels and geopotential heights is needed. Table 1 provides such conversion as determined by the hydrostatic balance. The detailed values of Table 1 are from the mean tropical sounding of Jordan (1958) and the mean moist tropical sounding of Dunion (2011), both of which gave approximately the same geopotential height values for corresponding pressure values. According to Table 1, the upper-level warm-core height of 12 km found by KSDD equates to approximately the 200-hPa pressure level. They also showed that the height of the warm core decreased as the environmental vertical wind shear magnitude increased, which is different from recent theoretical studies by Stern and Zhang (2013a,b) that showed the warm-core height is not altered by the environmental shear. But the AMSU-based statistics produced an unrealistic cool anomaly below 7–8 km (~400 hPa) altitude over all. SN12 questioned KSDD’s results and claimed that the coarse horizontal resolution of AMSU data (about 50 km) could not resolve the warm-core structure below 10-km altitude (~300 hPa). In fact, the vertical resolution of AMSU-based temperature retrievals is also very coarse. AMSU has only 12 microwave channels in the 50–60-GHz portion of the oxygen band to provide temperature soundings from the surface to about 1 hPa and only 6 channels from the surface to 100-hPa height. Theoretically, only 6 levels can be independently retrieved between the surface and 100 hPa. The coarse vertical resolution might contribute more than the coarse horizontal resolution on the unrealistic cool anomaly problem below 10 km by AMSU. The coarse resolution of AMSU made it harder to correct the heavy precipitation contamination in lower levels, which is the main reason for the cool anomaly problem (Zhu and Weng 2013; Kidder et al. 2000). Currently, a higher horizontal and vertical resolution satellite instrument set is available for observing temperature profiles globally, which is the Atmospheric Infrared Sounder (AIRS and companion AMSU; Aumann et al. 2003; Chahine et al. 2006) onboard the NASA Aqua satellite launched in 2002. AIRS is an infrared spectrometer with 2378 spectral channels and a 12.5-km footprint at nadir. The high spectral resolution of AIRS allows it to provide temperature retrievals with a vertical resolution of ~1 km. The companion AMSU is a 15-channel microwave radiometer with 45-km footprint size at nadir. The AIRS+AMSU (in the following text, the AIRS+AMSU product will be referred to as AIRS product for short) level 2 product includes an advanced cloud-clearing technique (Chahine et al. 2001; Susskind et al. 2003) that employs microwave observations from AMSU along with the AIRS observations to remove cloud contaminations and retrieve temperature and humidity profiles at 100 vertical levels.

<table>
<thead>
<tr>
<th>Pressure (hPa)</th>
<th>1000</th>
<th>925</th>
<th>850</th>
<th>700</th>
<th>600</th>
<th>500</th>
<th>400</th>
<th>300</th>
<th>250</th>
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</thead>
<tbody>
<tr>
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<td>0.8</td>
<td>1.5</td>
<td>3.2</td>
<td>4.4</td>
<td>5.9</td>
<td>7.6</td>
<td>9.7</td>
<td>10.9</td>
<td>12.4</td>
<td>14.2</td>
<td>16.6</td>
</tr>
</tbody>
</table>

**Table 1.** Corresponding geopotential heights of all pressure levels used in this study according to the mean tropical sounding of Jordan (1958) and the mean moist tropical sounding of Dunion (2011). Adapted from Dunion’s (2011) Table 1 and Table 2.
AIRS is expected to produce retrievals of temperature and water vapor with high accuracy under clear and partly cloudy (cloud fraction up to 80%; Tobin et al. 2006) conditions. Global validations found that the accuracy of AIRS level 2 temperature retrievals is about 1–2 K km$^{-1}$ in cloudy conditions (Wong et al. 2015). However, the specific accuracy of AIRS retrievals in TCs is still unknown. Recently, a preliminary statistical study by Gao et al. (2017) has showed some realistic warm-core structures of TCs in the northwest Pacific basin using AIRS level 2 data from 2002 to 2013. They also found a strong relationship between warm-core strength and TC intensity.

In this study, we will first compare AIRS temperature profiles in TCs with data from aircraft-deployed dropsondes, then construct the global climatology of TC warm-core structures using AIRS observations from 2002 to 2014. Specifically, we will address the following questions: 1) What is the typical TC warm-core structure as seen by the AIRS data? How does the structure vary in different TC-prone basins? 2) How good the correlation is between the warm-core strength and TC intensity? How does the relationship vary in different basins? 3) Is the height of warm core related to TC intensity? If so, to what degree? Section 2 provides a description of the data and methods applied in this study. The validation of AIRS temperature profiles in TCs against aircraft dropsonde data is presented in section 3. Section 4 presents the global climatology of TC warm-core structures and their relationships to TC intensity and geographical location. The discussion and conclusions are presented in sections 5 and 6, respectively.

2. Data and methodology

a. The AIRS/AMSU instrument and temperature retrieval product

AIRS and its companion microwave instrument AMSU flying on board the Aqua satellite have provided integrated atmospheric sounding information since its launch in May 2002. As mentioned in the introduction section, the AIRS instrument views the atmosphere in 4 near-infrared and visible channels and 2378 infrared channels. AMSU operates in 15 channels, with center frequencies ranging from 23 to 90 GHz (Chahine et al. 2006). The AIRS version 6 level 2 products provide AIRS+AMSU-retrieved profiles of temperature and humidity since August 2002, covering a roughly 1650-km swath, and are available twice daily at the local time (LT) of about 0130 and 1330 LT on a sun-synchronous orbit from a 705-km altitude. Both support and standard products are available to the research community. The support product contains temperature profiles at 100 vertical levels with a horizontal resolution of about 45 km. The standard product only provides temperature retrievals at 28 pressure levels between the surface and about 0.1 hPa with the same horizontal resolution as in the support product. In this study, we only use the temperature retrievals in the standard AIRS version 6 level 2 product with best or good quality at 12 pressure levels from 1000 to 100 hPa during August 2002–December 2014. The 12 pressure levels and their corresponding geopotential height values according the mean tropical sounding of Jordan (1958) and the mean moist tropical sounding of Dunion (2011) are given in Table 1. Specifically, best-quality data individually meet the designed accuracy requirements [i.e., absolute accuracy of 1 K in 1-km-thick layers in the troposphere; Aumann et al. (2003); Chahine et al. (2006)] and good-quality data meet the accuracy requirements only when temporally and/or spatially averaged.

b. TC best track data and selection of AIRS overpasses

The TC best track data during 2002–14 are obtained from the Joint Typhoon Warning Center (JTWC) and National Hurricane Center (NHC), including TC intensity (maximum sustained wind speed $V_{\text{max}}$) and storm center location. Both of them are linearly interpolated to match the observational time of AIRS. Six TC-prone basins are considered in this study: Atlantic (ATL), eastern and central Pacific (EPA), northwestern Pacific (NWP), northern Indian Ocean (NIO), southern Indian Ocean (SIO), and South Pacific (SPA). Not all TCs can be well observed by the AIRS with its swath width of 1650 km. Here we only select the AIRS overpasses that can capture the TC center. In addition, only those overpasses with available temperature profiles at all levels between 1000 and 100 hPa are used. A total of 7613 AIRS TC overpasses in 1061 TCs are identified based on these criteria (Table 2). We also stratify the overpasses into 6 TC intensity categories, including tropical storm (TS), category 1 (Cat1), category 2 (Cat2), category 3 (Cat3), category 4 (Cat4), and category 5 (Cat5) according to the Saffir–Simpson hurricane wind scale. The number of AIRS overpasses over each basin and each intensity category during 2002–14 is presented in Table 2. In general, the number of overpasses decreases as the TC intensity increases, which is consistent with the statistics in the best track data. The NWP basin has the greatest number of storms during the study period, therefore the greatest number of AIRS overpasses, while the NIO basin has the least number of storms and AIRS overpasses. Figure 1 shows the geographic distribution of storm centers covered by the 7613 selected AIRS overpasses during 2002–14. The strongest storms (Cat4 and Cat5) are mostly concentrated between the 5° and 25° latitude band.

To study the TC warm core, we need to calculate the temperature anomaly first. The temperature anomaly is defined as the difference between the observed
temperature in TCs and a reference environmental temperature profile. The question is complicated on how to choose the reference sounding because different choices can lead to different warm-core structures. There have been a number of choices used previously. Some studies (La Seur and Hawkins 1963; Hawkins and Rubsam 1968; Hawkins and Imbembo 1976) simply used the mean tropical sounding of Jordan (1958), while others used either the domain-averaged sounding with TC excluded or the mean sounding over an annulus within a specific range of distance from the storm center (KSDD; Halverson et al. 2006; Zhu and Weng 2013). As discussed extensively in SN12 and Munsell et al. (2018) and tested in Durden (2013) and Stern and Zhang (2016), the recommendation is to use an average taken at least several hundred kilometers from the TC center. For this study, the reference environmental temperature profile is calculated for each AIRS TC overpass by taking the average within 900–1400 km from the storm center.

3. Validation of AIRS temperature profiles in TCs

Previous studies have assessed and validated the AIRS temperature retrievals of different versions under clear-sky and cloudy conditions. For clear sky, the root-mean-square errors (RMSEs) are generally within 1 K km$^{-1}$ (the AIRS design expectation) as compared against collocated dropsonde (Gettelman 2004) and radiosonde data (Divakarla et al. 2006). Under cloudy regions, Yue et al. (2013) reported that the bias of AIRS version 5 temperature retrievals is about ±1 K from the surface to 200 hPa, with the largest biases from deep convective and nimbostratus clouds. More recently, Wong et al. (2015) quantified the uncertainties of the AIRS level 2 version 6 temperatures as a function of cloud types by using a global radiosonde dataset. They reported that temperature biases of all data are within ±1 K at altitudes above the 700-hPa level but increase with decreasing height. They also indicated that the AIRS GOOD quality RMSEs are about 1.5–2.5 K in thin clouds and slightly larger in thick clouds (reaching 3.5 K near the surface). However, these studies are for all atmospheric conditions. No studies are found at time of this writing on evaluating AIRS retrievals in TCs specifically.

Therefore, in this study, we first validate the AIRS temperature retrievals in TCs using aircraft dropsonde

![Fig. 1](image_url). The geographic distribution of the storm center covered by the 7613 selected AIRS overpasses during 2002–14. The color represents storm intensity.
data. Dropsondes are critical observational instruments used by NOAA and NASA aircraft missions into hurricanes to observe the temperature soundings from the altitude of the aircraft to the surface. In 2014, there were a series of aircraft missions into hurricanes that included flights with the NASA unmanned Global Hawk (GH) aircraft as part of the Hurricane Severe Storm Sentinel (HS3) field campaign (Braun et al. 2016; Zawislak et al. 2016). The GH is a high-altitude aircraft that reached altitudes up to 16.8–19.9 km. Dropsonde data from HS3 GH missions into TCs during September 2014 are collected and collocated with AIRS observations. The collocation criteria are defined as no more than a 50-km spatial distance and 3-h temporal difference. A total of 120 pairs of collocated dropsonde/AIRS temperature profiles are identified. Out of the 120 AIRS profiles (hereafter ALL), 61 profiles meet the AIRS BEST quality criteria (hereafter BEST and 108 meet the BEST or GOOD quality criteria (hereafter BEST+GOOD).

Figure 2 shows the profiles of biases (mean differences) and RMSEs between dropsonde measured and AIRS-retrieved temperatures for different AIRS data quality groups. The best agreement with dropsonde measurements is found for the AIRS BEST group between 200- and 700-hPa levels (where warm cores are typically located) with biases within ±0.5 K and RMSEs less than 1 K. Similar agreement for the AIRS BEST+GOOD group is found with slightly higher RMSEs. There is a cold bias near surface (peak at 850 hPa) for the AIRS ALL group, consistent with Wong et al.’s (2015) finding in cloudy conditions. The cloud-induced near-surface cold bias is the largest (about 1 K) for the AIRS ALL group, but is about a factor of 2 smaller for the BEST and BEST+GOOD profiles. At levels above 200 hPa, a cold bias of <1.2 K of AIRS retrievals is seen. Both of the upper-level biases and RMSEs are increasing with height above 200 hPa, with RMSEs nearly 1.6 K at 100hPa. A similar cold bias of −0.5 to −1.0 K of global AIRS temperature retrievals against radiosonde data was reported by Wong et al. (2015, their Fig. 8).

Next, we will present the case of Hurricane Edouard (2014) to compare the AIRS-derived temperature anomalies with those derived from dropsondes. Note that AIRS retrievals of all qualities are included in Figs. 3–4 below in order to show the entire TC region of Edouard. Therefore, some cloud-induced cold bias might be seen at lower levels. (However, only GOOD and BEST quality data points will be used in Figs. 5–14 to ensure no substantial cold bias problem in the climatological data analysis in this study.) Figure 3 presents the longitude–height cross section of AIRS-retrieved temperature anomaly of Hurricane Edouard (2014) on three different days. On 12 September, the storm was at tropical storm strength with intensity around 20 m s⁻¹. The AIRS temperature anomaly shows a weak warm-core structure with a maximum anomaly around 6 K that peaked between 300 and 500 hPa (Fig. 3a). As the storm intensity increased to 41 m s⁻¹ at 1648 UTC 15 September (Fig. 3b), the AIRS-retrieved warm core became stronger with a maximum anomaly of 12 K at 300 hPa. The 2-K warm anomaly extended from the surface to 150 hPa. Around Edouard’s peak intensity of 54 m s⁻¹ at 1736 UTC 16 September (Fig. 3c), the 2-K warm anomaly kept the same vertical extent as in 15 September but became wider horizontally. The maximum warm anomaly of 12 K is located between 300 and 400 hPa.

Munsell et al. (2018) showed the radius–height cross section of azimuthal-mean perturbation temperature for Edouard’s inner-core dropsondes deployed during the 16–17 September HS3 GH flight (their Fig. 6b) and the CIMSS-processed AMSU-A data from the 2025 UTC 16 September Edouard overpass (their Fig. 6d). To directly compare the warm-core structure of Edouard retrieved by AIRS with that observed by the dropsondes as in Munsell et al.’s (2018) Fig. 6b (shown in Fig. 4b here), Fig. 4a presents the radius–height cross section of azimuthal-mean perturbation temperature for Edouard’s inner core as derived from AIRS observations on 16–17 September.
2018. Temperature anomalies in Fig. 4a are defined differently as in the rest of the paper. Here the reference environmental temperature profile is calculated by taking the average within 300–700 km from the storm center, same as in Munsell et al. (2018), whereas in all other figures of this study (Figs. 3, 5–14) the reference profile is averaged within 900–1400 km from the storm center. However, the two reference profiles are similar for Edouard case.

The dropsonde-derived warm-core structure of Edouard (Fig. 4b) was very similar to what is derived from the AIRS retrievals (Fig. 4a). The dropsonde-derived maximum warm anomaly (Fig. 4b) was 12 K, the same as shown in Fig. 4a. The AIRS-derived warm-core height of 300–400 hPa corresponds to 7.6–9.7 km according to Table 1, which is consistent with the 8–10-km warm-core height derived from the dropsondes. In contrast, as shown in Munsell et al.’s (2018) Fig. 6d, the AMSU-A-derived temperature anomaly showed two distinct maxima: one slightly higher (~9–11 km) than in dropsonde or AIRS data, and one significantly lower in the atmosphere (near the surface). The magnitudes of these maxima were much weaker (~4 K for the upper-level maximum and ~6 K for the lower maximum). This indicated that the AIRS temperature retrievals are comparable with the dropsonde observations for Edouard, whereas the AMSU-A temperature retrievals are inadequate for observing the inner-core temperature structure of Edouard. There are some more detailed structures in Fig. 4b that are not captured by the AIRS retrievals because of lower vertical resolution of AIRS relative to the dropsondes.
Based on results from Figs. 2–4, AIRS can retrieve temperature profiles in TCs reasonably well. The AIRS bias relative to dropsonde data is within 1–2 K on average for multiple TCs during September 2014 when using GOOD+BEST quality data. More importantly, the TC warm core derived from AIRS extends from 150 hPa to the surface in Figs. 3b,c and 4. This is an improvement to the AMSU-derived TC temperature anomalies, which showed an unrealistic cold core structure below 400 hPa as in KSDD and Kidder et al.’s (2000) Fig. 6. For the case of Hurricane Edouard (2014), although the dropsonde-derived warm-core structure is more detailed because of its higher vertical resolution, AIRS can adequately resolve the warm-core strength and height, producing a warm-core structure comparable with that derived from the dropsondes, while the AMSU-A failed to do so. Therefore, for the purpose of this study, AIRS data are suitable for determining the climatology of the height and strength of the TC warm-core structure.

4. Results

Figure 5 shows the radial–height composites of azimuthally averaged temperature anomalies retrieved by AIRS in TCs for six different TC-prone basins. In general, the height of warm core is located around 300–400 hPa for all TCs in global basins. The warm anomaly of ≥1 K extends vertically from the surface (near 1000 hPa) to the upper troposphere around 150–200 hPa and horizontally from the TC center to beyond the 300-km radius. The warm core is the strongest for TCs in the NWP basin (8 K), followed by TCs in SPA and SIO (6 K), ATL (5 K), EPA (4 K), and NIO (3 K) basins in decreasing order. The basin-dependent variation of the warm-core strength is primarily due to the TC intensity difference in different basins. The mean TC maximum sustained wind intensity $V_{\text{max}}$ is the highest in the NWP basin, while it is the lowest in the NIO basin (Table 3). However, the warm-core strength differences in the ATL, EPA, SPA, and SIO basins are not perfectly correlated with the $V_{\text{max}}$ intensity differences in these basins. For example, the mean $V_{\text{max}}$s in SIO and EPA are similar (i.e., both are around 31 m s$^{-1}$), but there is a 2-K difference in the composite warm-core strength for TCs in these two basins. Instead, the warm-core strength is better correlated with the minimum sea level pressure ($P_{\text{min}}$, Table 3 and Fig. 5) in different basins. The lower warm-core strength in EPA relative to SIO corresponds to weaker TC intensity defined by $P_{\text{min}}$. Zhang and Chen (2012) and Kieu et al. (2016)
demonstrated that the warm-core structure is more associated with TC intensity defined by $P_{\text{min}}$ through the hydrostatic relationship.

The AIRS+AMSU level 2 standard products also provide cloud fraction, total precipitable water (TPW), and sea surface temperature (SST) retrievals. Table 3 presents the mean SST, cloud fraction, and TPW averaged within the 500-km radius from the TC center for AIRS TC overpasses in each basin. In Table 3, the mean environmental vertical wind shear, averaged between 200–850 hPa and the 500–750-km radius from the TC center (Zagrodnik and Jiang 2014), is derived from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) data.

The environmental vertical wind shear is lower in NWP and EPA basins (6–7 m s$^{-1}$) and higher in other basins, especially in SPA and NIO (≥9 m s$^{-1}$, Table 3). The TPW is the highest for the NWP (53 mm), followed by NIO, SPA, SIO, and EPA (around 50 mm), and ATL (48 mm) as the lowest. The SST is greater than 28°C for TCs in the NIO and NWP; between 27°C and 28°C in ATL, SPA, and SIO; and below 27°C for the EPA basin. The mean cloud fraction of TCs in the EPA basin is also in the low end. We speculate that the relatively low SST in EPA contribute to low cloud fraction and weaker environmental wind shear magnitude decreases as TC intensity increases, while the SST and cloud fraction increases as storm intensity increases (Table 5). From Figs. 6–7 and Table 5, we speculate that the warm-core strength and height might be systematically dependent on the shear and SST. KSDD found a negative relationship between the warm-core height and the magnitude of environmental shear using AMSU data, while Durden (2013) using limited dropsonde and radiosonde data and Stern and Zhang (2016) using numerical simulations of Hurricane Earl (2010) found no such relationship. A further study is under way to examine possible correlations between the level of the AIRS-derived maximum anomaly and a series of storm parameters (including thermodynamic and environmental parameters) with a goal to investigate if such relationships exist.

Figure 8 presents the composite temperature anomaly profiles within 30 km of the storm center for different TC intensity categories and basins. Consistent with Fig. 5, the NWP profile for all TCs features higher warm-core height and stronger warm-core strength than those for all other basins (Fig. 8a). NIO profiles stratified into different intensity groups are not shown in Figs. 8b–d due to the small sample size in this basin (Table 4). For TSs, the strongest warm-core anomaly is at 400 hPa, but there is a secondary warm anomaly in the lower troposphere (between 800–900 hPa) for storms in the SPA basin (Fig. 8b). No significant warm-core differences are seen in different TC basins for the Cat1–Cat2 intensity group (Fig. 8c). The warm-core strengths of major hurricanes (Cat3–Cat5, Fig. 8d) in the ATL and NWP basins are stronger than those of SPA, SIO, and EPA basins.

The composite results in Figs. 5–8 represent the overall average warm-core structure in TCs. To examine

<table>
<thead>
<tr>
<th>Basin</th>
<th>$V_{\text{max}}$ (m s$^{-1}$)</th>
<th>$P_{\text{min}}$ (hPa)</th>
<th>Shear (m s$^{-1}$)</th>
<th>SST (°C)</th>
<th>Cloud fraction</th>
<th>TPW (mm)</th>
</tr>
</thead>
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<tr>
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</table>
the detailed distributions of warm-core strength and height in TCs, probability density functions (PDFs) of the warm-core strength and warm-core height for six different basins and six TC intensities are shown in Figs. 9 and 10, respectively. Here the warm-core strength is defined as the maximum temperature anomaly within 30 km from the storm center using the data samples in Table 4. Traditionally, the warm-core height is defined as the height of the maximum temperature anomaly (SN12; Durden 2013; Gao et al. 2017). However, as pointed out by Ohno et al. (2016), “this is not a clear indicator of the height of the warm core in cases with less clear peaks (p. 4295).” We therefore follow Ohno et al. (2016) to define a weighted height of the warm core by the temperature anomaly within 30 km from the storm center using the formula given by Eq. (1) of Ohno et al. (2016). The warm-core strength ranges from 1 to 20 K with the strongest warm core (18–20 K) seen in the NWP and ATL basins (Fig. 9a). As in Figs. 5 and 8a, the median warm-core strength is the highest for TCs in the NWP basin, followed by SPA, SIO, ATL/EPA, and NIO basins in decreasing order. The most frequent warm-core strength is between 2 and 8 K (the mode of the PDFs in Fig. 9a).

The weighted warm-core height ranges from 850 to 150 hPa (Fig. 9b). Using limited cases measured by dropsondes and radiosondes, Durden (2013) showed that the warm-core height varies between 760 and 250 hPa. But the long-term satellite-based AIRS climatology in Fig. 9b indicates that there are about 20% of TC cases with warm-core height higher than 250 hPa and about 3% cases with warm-core height as low as 850 hPa. Although the AIRS-derived warm-core height has a wide range, a distinct mode of 300–400 hPa is seen for TCs in all basins. A total of 60%–70% TC cases have a warm-core height falling between 300 and 400 hPa. This clearly indicates that the typical warm-core height is located at the upper troposphere around 300–400 hPa, consistent with Fig. 7.

Figure 10 displays PDFs of the warm-core strength and height for six different TC intensity categories. As seen in Figs. 6–7, the warm-core strength is well correlated with TC intensity, as the median maximum temperature anomaly increases from 5 K for TS, 8 K for Cat1, 10 K for Cat2, 11 K for Cat3, to 12 K for Cat4 and 13 K for Cat5 storms. The mode of the PDF of warm-core strength increases as the TC intensity increases, from 5 K for TSs to 15 K for Cat4–Cat5 hurricanes (Fig. 10a).

Unlike Fig. 9b where the mode of warm-core height concentrates at between 300 and 400 hPa for all different basins, the modes of warm-core height are different for different TC intensity categories (Fig. 10b). The typical warm-core height is 400 hPa for TSs, 300 hPa for Cat1–Cat3 hurricanes, 250–300 hPa for Cat4 hurricanes, and
150 hPa for Cat5 hurricanes. A majority of the cases with lower-troposphere (between 400 and 850 hPa) warm-core height are TSs. The rest are Cat1–Cat3 hurricanes. There are 50% of Cat5 hurricanes whose warm-core height is at 150 hPa, 12% of them at 200 hPa, 30% at 250 hPa, and only 8% at 300 hPa. No Cat4–Cat5 hurricanes have warm-core height lower than 300 hPa. The minimum warm-core height of Cat3 hurricanes is 400 hPa.

To examine how good the relationships are between TC intensity in terms of $V_{\text{max}}$ and $P_{\text{min}}$ and warm-core strength and height as seen in Figs. 6–7 and 10, scatterplots of $V_{\text{max}}$ and $P_{\text{min}}$ versus the warm-core strength and height within 30 km of the storm center for each TC-prone basin are presented in Figs. 11–14, respectively. The warm-core strength is well correlated with $V_{\text{max}}$ with correlation coefficients ranging between 0.59 and 0.84 (Fig. 11). The highest correlation is seen in the ATL basin, followed by NWP, EPA, SIO, and NIO in decreasing order. Similar correlations are seen between $P_{\text{min}}$ and warm-core strength in the EPA, NWP, SIO, and SPA basins, but higher correlations between $P_{\text{min}}$ and warm-core strength (Fig. 12) are seen in the ATL ($-0.89$) and NIO ($-0.65$) basins than those between $V_{\text{max}}$ and warm-core strength (Fig. 11). This is consistent with Zhang and Chen (2012) and Kieu et al. (2016), which showed that the warm core is better correlated with $P_{\text{min}}$ than with $V_{\text{max}}$.

All the correlations shown in Figs. 11–12 are significant at the 99.99% level, except for the NIO basin, which is at the 97% significant level, mainly due to the small sample size in this basin. Figures 11–12’s result is consistent with Gao et al. (2017), which showed a positive correlation of 0.89 between $V_{\text{max}}$ and maximum temperature anomaly within the radius of maximum wind of TCs in the NWP basin. Durden (2013) also indicated that the maximum temperature anomaly in the eye is correlated with the TC intensity indicated by $P_{\text{min}}$ at the 99.5% significant level, although the correlation coefficient was not specified. Similarly, Zhu and Weng (2013) examined the ATMS-derived temperature anomalies for 53 observations in 10 TCs and found correlation coefficients of 0.78 and 0.83 between the maximum warm-core temperature anomaly and $V_{\text{max}}$ and $P_{\text{min}}$, respectively, at the 99% confidence level.

Similar correlations are seen between the weighted warm-core height and $V_{\text{max}}$ (Fig. 13) and between the weighted warm-core height and $P_{\text{min}}$ (Fig. 14). Overall the warm-core height increases as the TC intensity increases. When considering the full TC intensity range in this study with $V_{\text{max}} \geq 17$ m s$^{-1}$, the correlation coefficients (shown in blue in Figs. 13–14) are lower than those in Figs. 11–12, ranging between $-0.33$ and $-0.57$. However, the lowest correlation is from the NIO basin, which is not significant (confidence level 50%), whereas the correlations shown in blue in Figs. 13–14 for all other basins are significant at the 99.99% level. When considering only those samples with $V_{\text{max}}$ greater than 32 m s$^{-1}$ (hurricane strength, shown in red in Figs. 13–14), the NIO correlations are the highest ($-0.98$ and 0.9) and significant at the 95% confidence level. The correlations for hurricanes in all other basins as shown in red in Figs. 13–14 are significant at the 99.99% level. When considering samples reaching hurricane strength only, the correlations between the warm-core height and $V_{\text{max}}$ improve for EPA (from $-0.45$ to $-0.68$), NWP (from $-0.48$ to $-0.51$), and SPA (from $-0.57$ to $-0.70$) basins (Fig. 13). The correlation coefficients for ATL (from $-0.52$ to $-0.49$), and SIO (from $-0.41$ to $-0.31$) basins decrease instead.
Similarly, the correlations between the warm-core height and $P_{\text{min}}$ improve for EPA (from 0.45 to 0.48), NWP (from 0.47 to 0.48, and SPA (from 0.57 to 0.77) basins (Fig. 14). The correlation coefficients for ATL (from 0.51 to 0.41) and SIO (from 0.40 to 0.38) basins decrease instead.

Comparing with the high correlations between the warm-core strength and TC intensity (Figs. 11–12), the relatively low correlations between the warm-core height and TC intensity (Figs. 13–14) is primarily due to the relatively high warm-core heights in some TSs and Cat1 hurricanes as seen in both Figs. 13 and 14. Similarly low but statistically significant correlations between TC intensity and warm-core height were found in Durden (2013) and Gao et al. (2017). However, in the modeling case study of Hurricane Earl (2010), Stern and Zhang (2016) found no obvious relationship between the warm-core height and either intensity or intensity change, similar to those found in the modeling study of Hurricane Edouard (2014) in Munsell et al. (2018).

5. Discussion

This section examines the results of section 4 in light of previous work regarding the typical warm-core height

<table>
<thead>
<tr>
<th>Category</th>
<th>$V_{\text{max}}$ (m s$^{-1}$)</th>
<th>$P_{\text{min}}$ (hPa)</th>
<th>Shear (m s$^{-1}$)</th>
<th>SST (°C)</th>
<th>Cloud fraction</th>
<th>TPW (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TS</td>
<td>23.12</td>
<td>991.48</td>
<td>8.45</td>
<td>27.96</td>
<td>0.54</td>
<td>50.60</td>
</tr>
<tr>
<td>Cat1</td>
<td>36.39</td>
<td>985.44</td>
<td>8.27</td>
<td>27.93</td>
<td>0.58</td>
<td>50.30</td>
</tr>
<tr>
<td>Cat2</td>
<td>45.57</td>
<td>971.61</td>
<td>7.55</td>
<td>27.98</td>
<td>0.59</td>
<td>50.44</td>
</tr>
<tr>
<td>Cat3</td>
<td>52.81</td>
<td>955.68</td>
<td>7.57</td>
<td>27.99</td>
<td>0.59</td>
<td>50.24</td>
</tr>
<tr>
<td>Cat4</td>
<td>61.77</td>
<td>934.57</td>
<td>6.28</td>
<td>28.28</td>
<td>0.60</td>
<td>50.30</td>
</tr>
<tr>
<td>Cat5</td>
<td>69.77</td>
<td>909.28</td>
<td>5.33</td>
<td>28.59</td>
<td>0.64</td>
<td>50.92</td>
</tr>
</tbody>
</table>

Fig. 8. As in Fig. 7, but for different basins for (a) all TCs, and (b) TS, (c) Cat1–2, (d) Cat3–5 TCs.
of TCs. As noted in the introduction section, various warm-core heights were found by previous studies, which led to SN12’s argument that the typical warm-core height in TCs is unknown. However, using the long-term AIRS temperature retrievals in over 1000 TCs, this study confirmed the theory that the typical warm-core height is actually dependent on storm intensity. Although for all TCs in general, the composites (Fig. 7) and PDFs (Fig. 9b) of the height of maximum temperature anomaly within 30 km of the storms indicate that the typical warm-core height is at 300–400 hPa, the typical warm-core height varies as TC intensity varies (Fig. 10b). The warm-core height is 400 hPa for TSs, 300 hPa for Cat1–3 hurricanes, 250–300 hPa for Cat4 hurricanes, and 150 hPa for Cat5 hurricanes (Fig. 10b).

Besides these typical values, it is also important to note that the range of warm-core heights also varies with TC intensity. Therefore, it is expected to have some TC cases whose warm-core heights fall beyond the typical warm-core height values. Table 6 summarizes the typical values and ranges of warm-core height for different TC intensities as seen in Figs. 10b, 12, and 14. Next we will give examples from the current study and previous studies showing that these existing observational studies (mainly from dropsonde or radiosonde data) are generally consistent with the ranges of warm-core heights derived from the AIRS data for different TC intensities (Table 6).

The height of AIRS-derived warm core of TSs varies between 850 and 150 hPa as found in this study (Table 6). For the Hurricane Edouard (2014) case when it was in TS intensity on 12 September, the dropsonde data indicated a warm-core height of 300–400 hPa (Fig. 3a), which falls within the range for TSs in Table 6. Durden (2013) examined 18 eye soundings (for 8 storms) from the aircraft dropsonde data and 9 eye soundings (for 9 storms) from the ground-based radiosonde data. The corresponding storm intensities of these soundings are at hurricane strength or above (see his Table 4 and Fig. 19) after applying the wind–pressure relationship used by the Dvorak technique in the Atlantic (Kossin 2015). Durden (2013) found that the warm-core height varied between 760 and 250 hPa, which is generally consistent with the range of warm-core heights of 700–150 hPa for all Cat1–Cat5 hurricanes found in this study. Note that in this study we defined the warm-core...
height as the weighted height as in Ohno et al. (2016), which is different with Durden (2013) who defined it as the height of the maximum temperature anomaly. This definition difference might contribute to the difference of the lowest height found between this study and Durden (2013). In addition, the highest level analyzed in Durden (2013) was 250 hPa. This is probably why there was no case with a warm-core height higher than 250 hPa in his study.

La Seur and Hawkins (1963) demonstrated an upper-level warm core at ~240 hPa of Hurricane Cleo (1958) when it was a Cat1 hurricane ($V_{\text{max}}$ around 40 m s$^{-1}$). This falls within the range of AIRS-derived warm-core height for Cat1 hurricanes, between 700 and 150 hPa as in Table 6. Another example of Cat1 hurricane with eye dropsonde data available is Hurricane Erin (2001). Halverson et al. (2006) showed that the warm-core height of Erin was 500 hPa when its $V_{\text{max}}$ was about 40 m s$^{-1}$. Both Cleo and Erin’s warm-core heights are not the typical value for Cat1 storms (300 hPa), but they are within the range of between 700 and 150 hPa.

On the first day of Hurricane Inez (1966) flight (27 September), the storm’s maximum winds were around 53 m s$^{-1}$, which qualified it as a Cat3 hurricane on the Saffir–Simpson hurricane wind scale (Hawkins and Imbembo 1976). Two temperature anomaly maxima were found by the flight level data: one at 300 hPa and one at 600 hPa. The upper-level warm-core height falls within the range of AIRS-derived warm-core height for Cat3 storms, between 400 and 150 hPa, but the lower-level warm-core height does not. Hawkins and Imbembo (1976) mentioned that the lower-level warm core was thought to be “rather unusual.” However, in this study, only one temperature anomaly maximum is identified for each AIRS-derived temperature profile within 30 km of the storm center. As shown in Hawkins and Imbembo’s (1976) Fig. 6, the lower-level maximum was smaller in size. Therefore, this maximum would be secondary to the larger upper-level maximum because of the AIRS’s resolution and the methodology used in this study. This is probably why no warm-core height lower than 400 hPa was identified in this study for Cat3 storms and none lower than 300 hPa was identified for Cat4–Cat5 storms. From aircraft dropsonde data, Munsell et al. (2018) showed that Hurricane Edouard’s (2014) warm-core height was 300–400 hPa on 16–17 September when the storm was a Cat3 hurricane ($V_{\text{max}}$ around 54 m s$^{-1}$). This warm-core height falls within the range in Table 6 for Cat3 storms as well.
Hurricane Hilda (1964) was a Cat4 hurricane when the flight level data were obtained on 1 October. As constructed by Hawkins and Rubsam (1968), the warm-core height of Hilda was ~275 hPa, which falls within the range of AIRS-derived warm-core height for Cat4 storms, between 300 and 150 hPa. On the second flight day (28 September) of Hurricane Inez (1966), the storm was a Cat5 hurricane with maximum winds of 81 m s\(^{-1}\). The warm-core height constructed from flight level data by Hawkins and Imbembo (1976) was 240 hPa. This also falls within the range of the AIRS-derived warm-core height for Cat5 storms, between 300 and 150 hPa as in Table 6.

In their idealized simulations of TCs, SN12 found that the warm-core height is generally at 4–8 km over a wide
range of storm intensities. This is much lower than the typical warm-core height found in this study (8–10 km). They did find an upper-level maximum near 13–14 km but its magnitude was always weaker than the midlevel maximum. One may question about the possibility that the AIRS retrievals might be biased to better detect an upper-level warm-core anomaly because of the cloud/precipitation contamination problem at lower levels as seen in Figs. 3 and 4. However, it is important to note that Figs. 3 and 4 were generated using AIRS all-quality data, which may contain some profiles with the cloud/precipitation contamination problem. After removing

![Graphs showing scatterplots of weighted height vs Vmax and Pmin for different basins.](image_url)

**Fig. 13.** Scatterplots of the weighted height of maximum temperature anomaly within the 30 km of the TC center vs $V_{\text{max}}$ TC intensity over 6 different basins: (a) ATL, (b) EPA, (c) NWP, (d) NIO, (e) SIO, and (f) SPA. The correlation coefficients of linear regression fitting for all TC intensity range (TS and above) are shown in blue, and for TC intensities $>32$ m s$^{-1}$ (hurricane strength) are in red.

**Fig. 14.** As Fig. 13, but for scatterplots of the weighted height of maximum temperature anomaly within the 30 km of the TC center vs $P_{\text{min}}$. 

![Graphs showing scatterplots of weighted height vs Pmin for different basins.](image_url)
those problematic profiles, the GOOD and BEST quality data shown in Fig. 2 indicate an agreement between AIRS and dropsondes-derived temperature profiles of within 1–2 K for all vertical levels below 150 hPa. All data analysis in Figs. 5–14 used only GOOD and BEST quality AIRS data. Therefore, based on the consistency of the AIRS results with the previous observational studies discussed above, we argue that the typical warm-core height is at the upper level around 8–10 km in general and increases as storm intensity increases. This is consistent with the traditional view of the typical warm-core structure. The structure of the warm core in SN12’s simulations might be representative of some special cases of TCs in lower-intensity ranges, but not for all TCs in general. Another theoretical work by Emanuel (1986) presented a calculated temperature anomaly in the eye of a mature TC, which peaks just above 12 km. Zhang and Chen (2012) simulated Hurricane Wilma (2005) using the Weather Research and Forecasting (WRF) Model and found the warm-core height to be above 12-km hPa at Wilma’s peak intensity (Cat5 hurricane).

### 6. Conclusions

In this study, a total of 120 pairs of collocated dropsonde/AIRS temperature profiles in TCs are identified during the month of September 2014 under the collocation criteria of 50-km spatial distance and 3-h temporal difference to evaluate the AIRS-retrieved temperature profiles. It is found that, in general, the BEST or GOOD quality AIRS temperature retrieval is accurate enough to capture the TC warm-core structures. Then, the global climatology of TC warm-core structures is constructed using AIRS observations from 2002 to 2014. The main findings are summarized as follows:

### Table 6. A summary of the range and typical value of the AIRS-derived warm-core height for different TC intensities. Examples from this study and previous studies that fall into the range of the AIRS-derived warm-core height under different TC intensity categories are also indicated.

<table>
<thead>
<tr>
<th>TC intensity category</th>
<th>$V_{\text{max}}$ range by Saffir–Simpson hurricane wind scale (m s$^{-1}$)</th>
<th>Typical AIRS-derived warm-core height</th>
<th>Range of AIRS-derived warm-core height</th>
<th>Examples of dropsonde/radiosonde observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>TS</td>
<td>17–32</td>
<td>400 hPa, ~8 km</td>
<td>850–150 hPa, ~1.5–14 km</td>
<td>This study: Hurricane Edouard (2014) 12 Sep; warm-core height: 300–400 hPa</td>
</tr>
<tr>
<td>Cat1</td>
<td>33–42</td>
<td>300 hPa, ~10 km</td>
<td>700–150 hPa, ~3–14 km</td>
<td>La Seur and Hawkins (1963): Hurricane Cleo (1958) 12 Sep; warm-core height: 240 hPa</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Halverson et al. (2006): Hurricane Erin (2001); warm-core height: 500 hPa</td>
</tr>
<tr>
<td>Cat2</td>
<td>43–49</td>
<td>300 hPa, ~10 km</td>
<td>700–150 hPa, ~3–14 km</td>
<td>Hawkins and Imbembo (1976): Hurricane Inez (1966) 27 Sep; warm-core height: 300 hPa</td>
</tr>
<tr>
<td>Cat3</td>
<td>49–58</td>
<td>300 hPa, ~10 km</td>
<td>400–150 hPa, ~8–14 km</td>
<td>Munsell et al. (2018): Hurricane Edouard (2014) 16–17 Sep; warm-core height: 8–10 km (~300–400 hPa)</td>
</tr>
<tr>
<td>Cat4</td>
<td>58–70</td>
<td>250–300 hPa, ~10–11 km</td>
<td>300–150 hPa, ~10–14 km</td>
<td>Hawkins and Rubsam (1968): Hurricane Hilda (1964); warm-core height: 275 hPa</td>
</tr>
<tr>
<td>Cat5</td>
<td>≥70</td>
<td>150 hPa, ~14 km</td>
<td>300–150 hPa, ~10–14 km</td>
<td>Hawkins and Imbembo (1976): Hurricane Inez (1966) 28 Sep; warm-core height: 240 hPa</td>
</tr>
</tbody>
</table>
1) The composite warm-core strength is the strongest (−8 K) for TCs in the NWP basin, and weakest (−3 K) in the NIO basin, while the heights of the warm core are very similar among different basins.

2) Both the warm-core strength and height increase as the storm intensity increases. The typical warm-core height is at the upper level around 300–400 hPa for all TCs, but varies with TC intensity: 400 hPa for TSs, 300 hPa for Cat1–3 hurricanes, 250–300 hPa for Cat4 hurricanes, and 150 hPa for Cat5 hurricanes.

3) The range of warm-core heights also varies with TC intensity: 850–150 hPa for TSs, 700–150 hPa for Cat1–Cat2 hurricanes, 400–150 hPa for Cat3 hurricanes, and 300–150 hPa for Cat4–Cat5 hurricanes.

4) The warm-core strength shows a significantly positive correlation with TC intensity and correlation coefficients ranging between 0.59 (NIO) and 0.89 (ATL) for different basins. The warm-core height has a weaker but still significant correlation with TC intensity for most of the TC-prone basins when considering the full TC intensity range. The correlations between TC intensity and warm-core height improve for EPA, NWP, and SPA basins when considering samples reaching hurricane strength only.

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