Exploratory Analysis of Upper-Ocean Heat Content and Sea Surface Temperature Underlying Tropical Cyclone Rapid Intensification in the Western North Pacific

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ABSTRACT

The statistical relationships between tropical cyclones (TCs) with rapid intensification (RI) and upper-ocean heat content (UOHC) and sea surface temperature (SST) from 1998 to 2016 in the western North Pacific are examined. RI is computed based on four best track datasets in the International Best Track Archive for Climate Stewardship (IBTrACS). The statistical analysis shows that the UOHC and SST are higher in the RI duration than in non-RI duration. However, TCs with high UOHC/SST do not necessarily experience RI. In addition, the UOHC and SST are lower in the storm inner-core region due to storm-induced ocean cooling, and the UOHC reduces more significantly than the SST along the passages of TCs in the lower-latitude regions. Moreover, most of the RI (non-RI) duration is associated with the higher (lower) UOHC, but this is not the case for the SST pattern. Meanwhile, the TC intensification rate during the RI period does not appear to be sensitive to the SST, but shows statistically significant differences in the UOHC. In addition, there is a statistically significant increasing trend in the UOHC underlying TCs from 1998 to 2016. It is also noted that the percentages of the TCs with RI show different polynomial and linear trends based on different calculations of the RI events and RI durations. Finally, it is shown that there is no statistically significant difference in the UOHC, SST, and the percentage of RI among the five categories of ENSO events (i.e., strong El Niño, weak El Niño, neutral, weak La Niña, and strong La Niña).

1. Introduction

Despite large improvement in the track forecast of tropical cyclones (TCs) in the past 30 years, intensity prediction remains a challenging task, due to the complicated processes involved in intensity change. As such, the internal dynamical processes, large-scale environmental conditions, and boundary processes regarding the atmosphere–ocean interaction are shown to be critical (Wang and Wu 2004; Elsberry et al. 2007; Knaff et al. 2013). Of particular interest in the issue of the TC intensity change is the rapid intensification (RI), defined as an increase of 30 kt (15.4 m s$^{-1}$; 1 kt $\sim$ 0.51 m s$^{-1}$) or more within 24 h (Kaplan and DeMaria 2003). Ito (2016) showed that the intensity forecast error is clearly connected to the intensification rate, and that the recent increase in RI events could have affected the intensity forecasting. Much of the previous work on RI has focused on the role of inner-core dynamics, such as the warm core, the eyewall potential vorticity (PV) ring, deep convection, hot towers, convective bursts, and so on (e.g., Hong et al. 2000; Schubert et al. 1999; Kossin and Schubert 2001; Nolan and Grasso 2003; Hendricks et al. 2010; Vigh and Schubert 2009; Rogers 2010; Guimond et al. 2010; Kieper and Jiang 2012; Chen and Zhang 2013; Chang and Wu 2017).

The literature has also focused on the favorable large-scale environmental conditions for RI. Kaplan and DeMaria (2003) found that RI preferentially occurs in storms that are significantly weaker than their maximum potential intensity and are located in environments with warmer sea surface temperatures (SSTs), higher lower-tropospheric relative humidity, easterly upper-level flow, and weaker vertical wind shear (VWS). Based on the climatology from 1989 to 2006 in the Atlantic (ATL) and eastern North Pacific (ENP), Kaplan et al. (2010) further demonstrated that the differences in certain environmental variables from the Statistical Hurricane Intensity Prediction Scheme (SHIPS) dataset (DeMaria and Kaplan 1994, 1999; DeMaria et al. 2006).

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The slow-moving intense TCs possess a deeper subsurface can intensify over a shallow subsurface warm layer while passage. In other words, the faster-moving intense TCs over the ocean can maintain higher UOHC and SST after TC translation speed would cause less SST cooling, and therefore the potential intensity (POT). Using these predictors, Kaplan et al. (2015) subsequently developed statistical models for RI. However, some unusual cases may still occur, such as under high VWS (Molinari and Vollaro 2010; Nguyen and Molinari 2012).

Shu et al. (2012) showed similar results in the western North Pacific (WNP), although they did not analyze the UOHC when comparing the RI and non-RI events. The majority of the aforementioned studies focused primarily on the role of atmospheric conditions in RI. In addition, Fudeyasu et al. (2018) indicated that the UOHC of TCs with RI are significantly greater than non-RI at tropical storm formation (TSF) and maturity, but SST of TCs with RI are significantly greater than non-RI at the maturity. However, the contribution of differences in SST and UOHC to RI has not previously been explored in detail. In contrast, the role of surface fluxes in TC intensification has long been understood. As an example, the wind-induced surface heat exchange (WISHE) theory of hurricane intensification (Emanuel 1986; Rotunno and Emanuel 1987; Emanuel 1989; Yano and Emanuel 1991) demonstrated a positive feedback process in which the ocean plays a critical role in supplying more surface heat flux under increasing swirling surface wind. However, a negative feedback exists due to the TC-induced SST cooling associated with the entrainment mixing and upwelling (e.g., Price 1981; Schade and Emanuel 1999; Lin et al. 2005; Wu et al. 2007; Cione and Uhlhorn 2003; Wu et al. 2016).

It has been shown that most of the major category 4 or 5 TCs in the Saffir-Simpson scale in various basins would experience RI in the region with high UOHC associated with either warm eddies or a thick warm mixed layer (e.g., Hong et al. 2000; Shay et al. 2000; Cione and Uhlhorn 2003; Lin et al. 2005, 2008, 2009a,b; Ali et al. 2007; Mainelli et al. 2008; Rozoff and Kossin 2011; Wada 2015). Moreover, TCs with faster translation speed would cause less SST cooling, and therefore the ocean can maintain higher UOHC and SST after TC passage. In other words, the faster-moving intense TCs can intensify over a shallow subsurface warm layer while the slow-moving intense TCs possess a deeper subsurface warm layer (Lin et al. 2009b). Wu et al. (2016) also indicated that the TC with lower translation speed would experience more upwelling and the subsequent cooling on the upper-ocean thermal structure, limiting the intensification of TC. Interestingly, Hendricks et al. (2010) indicated that both TCs with and without RI had similar translation speeds in WNP, but cases with RI had a greater translation speed in the ATL. In all, the translation speed of TCs in the ATL is faster than TCs in the WNP. Hendricks et al. (2010) also demonstrated that the SST varies more significantly in the meridional direction in the WNP, while more in the zonal direction in the ATL. Since most RI storms tend to move westward and west-northwestward and faster than other non-RI TCs, they can experience higher energy flux when moving toward the warmer ocean in the ATL. In other words, the impact of the upper ocean on RI is dependent on both the location and translational speed of a TC. In addition, the characteristics of TC response to the upper ocean are expected to vary among different basins (Knaff et al. 2013).

Recently, Balaguru et al. (2018) indicated that the RI magnitude has an increasing trend in the central and eastern tropical ATL. The correlation coefficient of SST with the time series of the 95th percentile of 24-h intensity changes is larger than UOHC in their analysis. This leads us to the question: What is the characteristic of UOHC/SST underlying RI and non-RI in the WNP? To address this, we examine the statistical relationships of the UOHC/SST between RI and non-RI. Moreover, we explore whether there is a trend for the UOHC/SST underlying RI and non-RI, and compare the trend of the RI percentage from the best track datasets with Advanced Dvorak Technique and Hurricane Satellite data (ADT-Hursat).

In general, category 4 or 5 TCs tend to form in the southeastern part of the WNP and move west-northwestward (Camargo and Sobel 2005; Camargo et al. 2007; Chan 2007a,b; Wada and Chan 2008). The higher TC intensity observed in the WNP occurs primarily as a result of the west-northwestward track with a longer lifespan (Wang and Chan 2002; Camargo and Sobel 2005). The location of TSF is greatly affected by El Niño–Southern Oscillation (ENSO), of which the warm phase has higher SSTs in the southeastern part of the WNP as well as strong convection. Wada and Chan (2008) also indicated that the number of super typhoons (i.e., $\geq 130$ kt) increased during the mature El Niño years. The associated higher UOHC in those years resulted in more intense and more persistent storms. Guo and Tan (2018) proposed the concept of short- and long-duration El Niño events, and suggested that the reduced VWS and increased midtropospheric humidity during short-duration El Niño events are caused by the westward
advection of the region with high UOHC. Similarly, Wang et al. (2015) indicated that UOHC increased over the region where the majority of RI events occurred, which is due to the anticyclonic wind anomaly in the subtropical gyre of WNP during the cold phase of Pacific decadal oscillation (PDO). Although these studies demonstrated that the ocean plays a critical role in modifying the ambient environmental conditions, the relative importance of atmospheric and oceanic conditions in RI has not previously been quantified.

There were also studies discussing the interactions between atmospheric and oceanic conditions associated with interannual and interdecadal variations of WNP TC activity (Ho et al. 2004; Chan 2007b; Yeh et al. 2010; Liu and Chan 2013). However, investigations of the long-term changes in UOHC and SST in the presence of TCs are lacking. Therefore, one of the goals of this study is to assess the statistical likelihood of changes in the UOHC and/or SST underlying RI/non-RI and RI occurrence under different ENSO scenarios.

In section 2, the data and the methodology used in this study are described. Results are presented in section 3, and the discussion and conclusions are shown in section 4.

2. Data and methodology

a. Data

Four sources of best track datasets including Japan Meteorological Agency (JMA), Joint Typhoon Warning Center (JTWC), Hong Kong Observatory (HKO), and China Meteorological Administration (CMA) from the International Best Track Archive for Climate Stewardship (IBTrACS; Knapp et al. 2009; Kruk et al. 2009) database are used to identify RI cases based on the maximum sustained surface wind (MSSW) and to analyze the location and intensification rate of TCs for the period from 1977 to 2016 in the WNP. Data are available at four times per day (0000, 0600, 1200, and 1800 UTC). Every track point would be calculated to the next four track points for the identified RI samples. Note that the MSSW from JMA and HKO is the 10-min average, which is the World Meteorological Organization (WMO) standard. The MSSW from JTWC is the 1-min average, while the 2-min average is used at CMA. In this study, to avoid biases in wind conversion, the original wind values of JTWC and CMA are not converted. These best track intensity estimations in the WNP are derived from the original Dvorak enhanced infrared technique. Note that the data analyses may not be consistent over time due to the technological development and the change of analysis protocols. Because the ADT-Hursat (Kossin et al. 2013) from 1978 to 2009 contains a more consistent record of TC intensity, it is used for the calculation of RI from 1978 to 2009 for trend analyses. Table 1 lists the numbers of TC cases and track point basis identified as RI and non-RI for each best track dataset. The total numbers of cases are 716, 715, 731, 728, and 879 for JMA, JTWC, HKO, CMA, and ADT-Hursat, respectively. Note that nearly twice as many RI events (267) are found in the JTWC data, likely due to the usage of the 1-min average. The ADT-Hursat also comprises more TC cases, while it is based on a 1-min average and 3-hourly time interval record. The 1-min average MSSW speed is also used in Kaplan and DeMaria (2003) from the National Hurricane Center (NHC) Hurricane Data (HURDAT) file.

Sea surface height anomaly (SSHA) used to infer UOHC is obtained from the Archiving, Validation and Interpretation of Satellite Oceanographic database (AVISO; Ducet et al. 2000; Pascual et al. 2006, 2009) distributed by The Segment Sol multimissions d’ALTimétrie, d’Orbitographie et de localisation précise/Data Unification and Altimeter Combination System (Ssalto/Duacs). This dataset combines all available SSHA observations from multiple altimetry missions (e.g., TOPEX/Poseidon, Jason-1/2, and Envisat) on a daily mean basis, providing consistent, homogeneous,
and high-quality SSHA data for climate studies. The domain of this study is defined as 0°–40°N, 100°E–180°W. The horizontal spatial resolution is 0.25°, and the temporal resolution is daily. With the same temporal and spatial resolution, optimally interpolated (OI) SST version 5 product is obtained from Remote Sensing Systems (RSS; http://www.remss.com). This dataset combines the cloud-penetrating microwave SST observations from Special Sensor Microwave Imager Sounder (SSMIS), Advanced Microwave Scanning Radiometer (AMSR2), WindSat, and Advanced Scatterometer (ASCAT), providing high quality of daily SST maps.

The oceanic Niño index (ONI; Kousky and Higgins 2007) is taken from the Climate Prediction Center (CPC), which is based on 3-month running mean of the Extended Reconstructed SST version 5 product (ERSSTv.5; Huang et al. 2017). This product contains SST anomalies in the Niño-3.4 region (5°N–5°S, 120°–170°W) from 1950 to the present (http://www.cpc.ncep.noaa.gov). To remove the warming trend, the ONI values are based on centered 30-yr base periods updated every 5 years.

b. Methodology

Various definitions of RI have been used in previous studies (Holliway and Thompson 1979; Kaplan and DeMaria 2003; Kaplan et al. 2010; Lee et al. 2016). To highlight the differences in the lifetime maximum intensity (LMI) of TCs undergoing RI and those not undergoing RI, we follow the definition of RI used in Lee et al. (2016) as an increase of 35-kt MSSW or more with 24 h. Every track point would be calculated to the next four track points for identified RI samples from four best track datasets. TCs reaching an intensity of at least 35-kt MSSW can be considered. The onset of RI is defined as the first time a TC satisfies the definition of RI. The definition of RI events is the actual number of RIs, which means that each TC can only be recorded with one RI at most. The RI duration is defined as the period from the onset of RI to the final time the TC reaches the RI threshold, and the non-RI duration represents the whole lifespan of non-RI event. In other words, a TC with RI during the non-RI period would not be considered for the non-RI duration (group) in this study. To mitigate the influence of large-scale atmospheric flow and topographic effects, cases are excluded when 1) the latitude of a TC exceeds 30°N or 2) landfall occurs within 24 h or 3) the lifespan of the TC is less than 2 days. On the basis of the above thresholds, the average percentile underlying RI duration for the four sources of best-track datasets is the 86.5th percentile of 24-h over-ocean intensity changes of WNP (Table 1), with the lowest percentile being 73rd for JTWC (representing 1-min average MSSW). The definition of RI by Kaplan and DeMaria (2003) is the 95th percentile of 24-h intensity change. These results imply that the TCs in WNP reach RI more easily than the TCs in the North ATL.

The SSHA data are used to estimate the vertical ocean temperature profiles by regression models. This has been shown to outperform the widely used two-layer method (Pun et al. 2007, 2014, 2016). Following Pun et al. (2014), the linear regression model for the western North Pacific (REGWNP) provides reasonable estimates of UOHC and the depth of the 26°C isotherm (D26), which is used as a proxy for the thickness of the warm water layer. The UOHC is calculated via (Leipper and Volgenau 1972; Shay et al. 2000; Goni and Trinanes 2003; Goni et al. 2009; Shay and Brewster 2010; Pun et al. 2014, 2016):

$$\text{UOHC}_{(x,y)} = C_p \rho \int_{z=0}^{D_{26}} \Delta T_{(x,y,z)} \, dz,$$

where $C_p$ is the heat capacity of the ocean at a constant pressure (4178 J kg$^{-1}$C$^{-1}$), $\rho$ is the density of seawater (1026 kg m$^{-3}$), and $\Delta T_{(x,y,z)}$ is the temperature difference from 26°C, meaning only temperature equal to or greater than 26°C is considered. It is noted that UOHC estimates have relatively large uncertainty (error > 35%) in the northern part of the WNP (i.e., 30°–40°N, 140°E–180°W), due to the relatively small values (Pun et al. 2014). This is another reason why TC cases above 30°N are excluded from this analysis. The TC cases included in this analysis are located over a region with higher UOHC and D26, where the data quality is expected to be more reliable. Figure 1 shows the 19-yr climatological seasonal averages of UOHC and SST, and the RI track points from JMA, which is similar to other best track datasets (not shown). It should be noted that the best track four times per day would match the one daily mean UOHC and SST. In this study for each TC case, the corresponding UOHC and SST values are averaged within the radius from TC center to 300 km, and results are not affected when using different radii from 100 to 1000 km (not shown). We further examine the UOHC/SST comparison of TCs between the RI/non-RI duration and 7 or 14 days prior to that time (referred to as P7 or P14), as well as the characteristics of UOHC/SST of RI/non-RI at different latitudes.

Typically, an ENSO event begins in the boreal summer and persists into the next year. To avoid oversimplification by the El Niño/La Niña dichotomy and to further investigate the influence of ocean impacts on TC in ENSO, five ENSO scenarios are identified based on average ONI during the ENSO period: strong El Niño (SE; ONI ≥ 1.0), weak El Niño (WE; 1.0 > ONI ≥ 0.5), neutral (NE; 0.5 > ONI > −0.5), weak La Niña (WL;
and strong La Niña (SL; ONI ≤ −1.0). This classification is similar to that used by Wang and Chan (2002), who suggested that different ENSO categories based on the seasonal mean Niño-3.4 SST anomalies are meaningful for ENSO consideration. Based on the classification, the ENSO events included in this study are 6 SE events, 5 WE events, 6 WL events, and 4 SL events; the rest of the periods are NE cases. The categories of the ENSO scenarios are shown in Fig. S1 in the online supplemental material. Although different countries and organizations may define the ENSO index differently, all indices are similar, resulting in similar sensitivity of the indices to the statistics in this study (not shown).
Compared with the averaged UOHC and SST, the following analyses illustrate the Z-score using the standard normalized distribution formula by Kreyszig (1979). The formula of the Z-score is

$$Z = \frac{x - \mu}{\sigma},$$

where $\mu$ is the mean of the data and $\sigma$ is the standard deviation of the data. The value of the Z-score represents the distance between the data and the mean data in units of the standard deviation. Generally, the Z-score is used to analyze the relative variation of the data in different locations. Note that to compare the magnitude of UOHC with SST and to further describe their distribution in this study, $\mu$ ($\sigma$) is used as the mean (standard deviation) of the UOHC or SST data for the region $0^\circ$–$30^\circ$N and $130^\circ$E–$180^\circ$ from 1998 to 2016. In this study, $\mu$ and $\sigma$ of UOHC is 72.9 and 42.0 KJ cm$^{-2}$, respectively, and $\mu$ and $\sigma$ of SST is 27.6$^\circ$ and 1.67$^\circ$C, respectively. For statistical methods, the statistically significant difference is based on the Mann–Whitney rank sum test at the 95% level, which is a comparison of two means and a null hypothesis of no difference. The statistically significant trend is based on a t test at the 95% level.

3. Results

a. UOHC and SST underlying RI and non-RI

We first present the results of the different probability density function (PDF) of the UOHC and SST underlying RI and non-RI duration based on the average of IBTrACS. The peaks of SST PDFs with RI and non-RI are similar (Z-score from 0.5 to 1), but those of the UOHC PDFs are different (Fig. 2). This implies that UOHC can be relatively more important than SST for RI. The UOHC/SST PDFs indicate that the TC underlying non-RI duration can sometimes have high UOHC/SST, and vice versa, even though the mean UOHC/SST of TC underlying RI duration is statistically more significant than those of non-RI duration cases. As mentioned earlier, the SST varies more significantly in the meridional direction in WNP. Since the locations of TCs have a considerable impact on the meridional temperature distribution of the ocean, further investigation of UOHC and SST regarding their latitudinal variations in WNP is necessary. We therefore examine how the TC-induced ocean cooling affects the UOHC and SST by dividing $0^\circ$–$30^\circ$N into six regions every 5$^\circ$. Figure 3 demonstrates the UOHC and SST of TC underlying RI and non-RI duration (Figs. 3a,d), P7 (Figs. 3b,e), and P14 (Figs. 3c,f) based on JMA in different latitude bands. The patterns of the other best track datasets are similar (see Table S1 in the online supplemental material). Differences in the UOHC and SST between the TC underlying RI and non-RI duration are statistically significant. The Z-score of the UOHC (SST) underlying RI duration from $25^\circ$ to $30^\circ$N is $-0.97$ (0.23). For the TC underlying non-RI duration, the UOHC (SST) value is $-1.1$ ($-0.5$). The difference of the Z-scores between RI and non-RI of the SST (0.73) is greater than for the UOHC (0.13) in the higher-latitude regions from $25^\circ$ to $30^\circ$N. On the other hand, the difference between the UOHC (0.28) of the TC underlying RI and non-RI duration is larger than the SST (0.26) in the lower-latitude regions from $5^\circ$ to $10^\circ$N. The result also shows the mean values of P7 and P14. Similar to the above result, the difference in SST between all TCs and P7/P14 in the higher-latitude regions is greater than UOHC, and vice versa. The difference of Z-score between TC and P7 for RI (non-RI) duration in the lower-latitude regions from $5^\circ$ to $10^\circ$N is 0.31 (0.41) in UOHC and 0.23 (0.25) in SST. The overall SST variations are shallower than the UOHC variations from the tropics to the subtropics, which is consistent with Lin et al. (2009b), who indicated that the SST is generally around 29$^\circ$C for category 5 TCs. Figures 4a and 4b shows the comparison between the UOHC and SST averaged from the TC center to 1000 km in the RI, non-RI duration, P7, and P14 by lower- and higher-latitude regions. The clear TC-induced ocean cooling is shown near the center. Meanwhile, the UOHC of TC underlying RI/non-RI duration is significantly less than the SST for all regions, and in particular near TC center in the higher-latitude regions from $25^\circ$ to $30^\circ$N. Figures 4c and 4d further demonstrate that the difference between UOHC (SST) of TC underlying RI and non-RI duration, UOHC (SST) of
RI P7 and TC underlying RI duration, and the UOHC (SST) of non-RI P7 and TC underlying non-RI duration such as Figs. 4a and 4b, respectively. The result also shows that the difference between the SST of TC underlying RI and non-RI duration is larger than the UOHC, especially in the higher-latitude regions. Meanwhile, the difference between SST of P7 and TCs (RI and non-RI duration) is larger than the UOHC in the higher-latitude regions, but on the other hand, the opposite pattern is shown in the lower-latitude regions. This implies that TCs need warmer SST to survive in the higher-latitude regions. Because the UOHCs in the higher-latitude regions always remain low for the whole year (Fig. S2), they may not be appropriate for investigating TC-induced ocean cooling as compared with SST. Note that the track points of the RI events/duration in the higher latitudes are relatively less than the lower-latitude regions. As a consequence, the evidence suggests that UOHC is more essential than SST for TC-induced ocean cooling in the lower-latitude regions.

The ocean composite temperature for cases that undergo RI based on JMA are also shown in Fig. 5. Again, patterns of the other best track datasets are similar (Figs. S3–S5). In this study, a TC is listed as a tropical...
storm from the first time when it reaches the 35-kt intensity, which is called TSF. In addition to the TC-induced ocean cooling, similar meridional asymmetric patterns have been found for TSF, the underlying RI duration of TCs, and LMI stages, in particular the UOHC in the higher-latitude regions, which is associated with the meridional variation of ocean temperature in WNP. For LMI, we only consider the TC that reach LMI for the first time if there are two or more LMI points in one TC. The UOHC and SST both decrease from TSF to LMI, which represent the negative feedback between TC intensity and ocean temperature (Figs. 5a–e,k–o). The lowering UOHC in the TC center is shown in TC underlying RI duration in the lower-latitude regions, and the amount of decrease is more significant than the SST (Fig. 5j). The UOHCs remain low in the higher-latitude regions (Figs. 5f–i), which is consistent with the above analyses. The SST cooling at the TC center is clearly shown in LMI from 10° to 30°N.

Meanwhile, the UOHC cooling in LMI is shown in a larger area from the TC center, in particular in the northern part of the TC (Figs. 5k–n), which resembles a TC underlying RI duration. This is also because the positions of LMI are usually located more to the north than those of TSF. The composite ocean temperature in RI groups demonstrate the overall decreasing UOHC in the northern part of TC. In comparison to the composite ocean temperature patterns, the UOHC and SST of P7 overall show larger values without clear TC-induced ocean cooling (Fig. 6; Figs. S6–S8). This pattern is similar to the composite ocean temperature in non-RI groups with different magnitude of Z-score (not shown). Note that there is only one case for TSF in the higher regions from 25° to 30°N. In addition, to distinguish the TC-induced ocean cooling in the northern part from the negative meridional gradient of UOHC/SST, we also investigate the difference between RI duration (also including TSF and LMI) and P7 and show that the result (Fig. S9) is consistent with Fig. 4. These results nevertheless suggest that oceanic conditions may play some roles in RI.

b. Long-term trend of UOHC and SST

Another interesting question is: What is the relationship between the UOHC/SST and the percentage of RI? Figure 7 shows the linear trend of the UOHC/SST underlying RI/non-RI and TC as well as the tendency of the UOHC/SST for P7, which is different from the overall ocean tendency without TC passage. Based on the t test, the UOHC underlying RI for 1998–2016 is the only one that shows a statistically significant increasing trend with 95% confidence level, while the others only increase slightly and have no significance (P > 0.05) from the JTWC, JMA, and HKO datasets. Note that all of the tendencies for CMA are nonsignificant. A similar trend is also shown for the UOHC of TCs. Meanwhile, the UOHC of P7 shows that the statistically significant trend and P value is smaller than the UOHC of TC underlying RI duration. Nevertheless, the percentage of RI events and TC underlying RI duration trend from different datasets for polynomial and linear trend for 1978–2016 show different results in Fig. 8. The only difference between Figs. 8a,c and Figs. 8b,d is the polynomial and linear trend line. The percentages of RI
events from JMA, HKO, and CMA show an increasing trend for polynomial trend line after 2000. As opposed to them, JTWC and ADT-Hursat show a decreasing polynomial trend after 2006, with $R^2$ around 0.42 (Fig. 8a). In contrast, Fig. 8b shows that the statistically significant increasing percentage of RI events is only found in JTWC, while the others do not have the statistically significant linear trend. In addition, the percentage of TC underlying RI duration from ADT-Hursat shows an increasing polynomial trend (Fig. 8c). By analyzing the linear trend for TC underlying RI duration, JTWC and ADT-Hursat demonstrate an increasing pattern, but the $R^2$ value of JTWC is larger than those in ADT-Hursat. Again, there is no significant trend ($P > 0.05$) from the JMA, HKO, and CMA datasets. Note that Fig. 7 shows a plateau during 2012–14. However, Fig. 8 does not show similar high percentage of RI events and RI durations during these years. Since the occurrence of non-RI is accompanied by high UOHC as shown in Fig. 2, low UOHC is less prone to RI, but high UOHC is not a necessary condition for RI. There is insufficient evidence to show that the percentage of RI increases is due to the UOHC in an increasing trend.

c. ENSO scenarios

To investigate how ENSO strength may affect TC activity, five ONI indices are considered as the five ENSO scenarios in this study. This study also examined the results of the five ENSO scenarios based on TC locations from four sources of best track data used in JTWC, JMA, CMA, HKO, respectively. Figure 9 shows the average locations underlying RI and non-RI duration in the five ENSO scenarios from 1977 to 2016, stratified by RI and non-RI, respectively. The standard deviation is represented by the oval area. Overall, both RI and non-RI show a geographical distribution from southeast to northwest for TCs in El Niño and La Niña conditions.
cases from all four best track datasets. The analyses show that the pattern of TSF for El Niño (La Niña) resemble the warm (cold) PDO phase, which is consistent with Wang et al. (2015) (Fig. S10). Likewise, the pattern of average locations of LMI shows similar results to Fig. 9 and Fig. S10 (Fig. S11). The number of cases for LMI is a little different from TSF since the locations of LMI can be beyond the selected latitude region (e.g., TSF and LMI of Typhoon Fifteen 2015 were at 22.6°N, 147.2°E, and 31.1°N, 141.8°E, respectively) or be affected by topography, while the TCs have a long lifespan covering a period with different ONI indices (e.g., Typhoon Parma from 0600 UTC 29 Sep 2009 to 0000 UTC 1 Oct 2009). It is interesting to note that the tracks of TCs in La Niña from TSF to LMI tend to propagate northward, whereas the tracks of TCs in El Niño tend to propagate west-northwestward. Although the percentage of RI in El Niño is greater than in La Niña from JTWC and JMA, there is no systematic difference among five ENSO scenarios by four sources of best track datasets (Fig. 10), which is not consistent with Fudeyasu et al. (2018), who suggested that RI occurrence rates of TCs in ENSO are El Niño 28%, NE 23%, and La Niña (15%). It is noted that RI percentages calculated based on JTWC are higher than those from other best track datasets (Fig. 10a).

Figure 11 shows the boxplots of the averaged UOHC and SST of TC underlying RI and non-RI duration from JMA in the five different ENSO scenarios from 1998 to 2016. The similar results are based on other best track datasets (not shown). Differences in the UOHC and SST among the five ENSO scenarios for TC underlying RI and non-RI duration are not statistically significant. This is possibly due to the regions in which TCs occurred among the five ENSO scenarios largely overlapping, even though the average locations of TCs are different.
This also implies that in both El Niño and La Niña, the energy provided by the ocean to TCs with RI/non-RI is similar. Nevertheless, it appears that the RI in El Niño/NE is likely to occur when warm UOHC is present rather than based on SST in the middle region (10°–15°N, 125°–150°E) of the WNP (Figs. 12a–c). The overall average locations of RI are more southeastern than non-RI even under different ENSO scenarios (Fig. 12), which is consistent with Fudeyasu et al. (2018).

4. Discussion

Different results will be obtained based on four sources of best track datasets in this study. Therefore,
more caution is necessary due to the data issue. The UOHC underlying RI shows a significant increasing trend from 1998 to 2016 based on the JTWC, JMA, and HKO best track datasets. However, the percentage of TCs with RI shows an ambiguous trend affected by different analyses and datasets. Only JTWC presents a statistically significant ($F$ test) increasing pattern for both RI events and TC underlying RI duration by polynomial or linear trend. However, ADT-Hursat shows the decreasing RI events, and the increasing TC underlying RI duration from 2006 to 2009. Meanwhile, there is no statistically significant increasing RI events and TC underlying RI duration from the JMA, HKO, and CMA. Further investigations are needed for decadal changes of TC underlying RI duration and the actual number of RI in the WNP.

More recently, Xu et al. (2016) indicated that a positive relationship between intensification rate of TC and SST exists in the ATL. To understand the characteristics of ocean underlying the different intensification rates during an RI period, Fig. 13 shows the boxplots of UOHC/SST based on the threshold of the RI definition with 6, 12, 18, and 24 h, respectively. The Kruskal–Wallis one-way analysis of variance is used to evaluate the statistical difference among different intensification rates (Kruskal and Wallis 1952). Statistically significant difference is found in the UOHC from JTWC and JMA. Moreover, all pairwise multiple comparison procedures through the Holm–Sidak method are applied to UOHC from JTWC and JMA to evaluate the comparison with different intensification rate groups (Holm 1979; Table 2). As a result, the UOHC of TC underlying RI duration occurring within 6 h is greater than that of the 12-, 18-, and 24-h groups from JTWC. Based on the result from JMA, the UOHC of TC underlying RI duration is also greater than the other two groups in the 12-h RI cases, but not in the cases of SST of TC underlying RI duration. This result resembles Hendricks et al. (2010), who suggested that the environment of RI and intensifying TCs is similar, but their analyses did not involve UOHC. However, Xu and Wang (2018) indicated that the relationship between the TC maximum potential intensification rate (MPIR) and SST exists in WNP and ATL. A higher SST environment generally possesses a higher MPIR, since MPIR is a function of MPI as indicated in Xu et al. (2016). Zhang and Emanuel (2016) also mentioned that the theoretical intensification rate might be dependent on the difference between the squared MPI and the squared current intensity. Therefore, we can conclude that the ocean...
FIG. 9. The plots and average area of (left) RI and (right) non-RI duration for different ENSO scenarios from 1977 to 2016 by (a) JTWC, (b) JMA, (c) HKO, and (d) JMA best track datasets. The oval areas indicate the standard deviation regions. The red, yellow, black, light blue, and blue colors indicate SE, WE, NE, WL, and SL, respectively.
provides an important environmental condition, but the contributions from other factors remain to be evaluated. We also demonstrate the TSF (Fig. S10), LMI (Fig. S11), and TC underlying (Fig. 9) RI duration associated with the ENSO in the WNP, and suggest that despite the different TC geographical distributions found among the five ENSO scenarios (Lander 1994; Wang and Chan 2002; Camargo and Sobel 2005; Chan 2007b; Zhan et al. 2011; Wang et al. 2015; Mei et al. 2015), no statistical difference is identified in the percentage of RI among the five ENSO scenarios from four sources of best track datasets, but the average percentage of RI is greater in El Niño than in La Niña from JTWC and JMA. This result is a little different from Fudeyasu et al. (2018), who showed a systematical different percentage of RI by JMA, which is probably due to usage of a different ENSO classification. Note that the higher percentage of RI presented in JTWC is probably a result of the use of maximum 1-min sustained wind speed as compared to 10-min sustained wind used in other analyses. Moreover, the El Niño years appear to have more TC cases than La Niña years (Fig. 9). In addition, the UOHC and SST underlying RI or non-RI duration are

![Image](https://example.com/image1.png)

**Fig. 10.** The 100% stacked column chart of the percentages of RI (red) and non-RI duration (light blue) for different ENSO scenarios in WNP from 1977 to 2016 by (a) JTWC, (b) JMA, (c) HKO, and (d) CMA best track datasets.

![Image](https://example.com/image2.png)

**Fig. 11.** Boxplots of the average (a) UOHC and (b) SST underlying RI and non-RI duration for different ENSO scenarios in WNP from 1998 to 2016 from JMA. The red, yellow, gray, light blue, and blue colors indicate SE, WE, NE, WL, and SL, respectively. The box represents the lower quartile ($p = 25\%$), the median, and the upper quartile ($p = 75\%$). The mean value is plotted as a cross mark, and the short horizontal lines indicate the maximum and minimum values of data. The unit is the $Z$-score.
FIG. 12. The 19-yr climatological composites of UOHC (kJ cm$^{-2}$) and SST (°C) underlying RI duration with the average location of RI (black) and non-RI (gray) duration for (a) SE, (b) WE, (c) NE, (d) WL, and (e) SL ENSO scenarios from 1998 to 2016. The oval areas indicate the standard deviation regions.
FIG. 13. Boxplots of UOHC and SST of different intensification rates of RI duration by (a) JTWC, (b) JMA, (c) HKO, and (d) CMA best track datasets in WNP from 1998 to 2016. The box represents the lower quartile ($p = 25\%$), the median, and the upper quartile ($p = 75\%$). The mean value is plotted as a cross mark, and the short horizontal lines indicate the maximum and minimum values of data. The unit is the $Z$-score.
also similar among the five ENSO scenarios. This indicates that in both El Niño and La Niña environments, the ocean can provide enough energy for TC development, in particular for RI. Note that if the analyses of the UOHC and SST follow the ENSO classification by Fudeyasu et al. (2018), the UOHC underlying RI has a systematic difference in its pattern (Fig. S12). Theoretically, following the ENSO classification by ONI means that the weight of the ocean impact would be increased. Therefore, the different results from Fudeyasu et al. (2018) may be associated with influence from the atmospheric processes.

5. Conclusions

RI occurs in TCs mainly due to atmospheric internal dynamical processes and environmental conditions, as well as ocean temperature. Previous studies have demonstrated the importance of atmosphere–ocean interaction in RI. For example, it has been pointed out that the TC-induced ocean cooling provides a negative feedback that can limit the storm intensification rate (Lin et al. 2008, 2009a,b; Wu et al. 2016), and that the UOHC in RI cases is significantly higher than non-RI cases in the ATL (Kaplan and DeMaria 2003; Kaplan et al. 2010, 2015). However, these studies did not focus on the relative importance of the UOHC and SST. One of the objectives of this study is to identify oceanic conditions, in particular the UOHC, that impact RI in the WNP. To compare the quantitative UOHC and SST, the Z-scores applied in this study are based on the standard normalized distribution. Results show that the UOHC/SST underlying RI duration is statistically significantly greater than the UOHC/SST underlying non-RI duration in Fig. 3, which is consistent with previous studies. Moreover, the difference between TC underlying RI and non-RI duration is statistically significantly greater in UOHC PDF than in SST PDF. Meanwhile, there are greater PDFs in non-RI than RI at higher UOHC, while most TCs underlying RI duration are associated with higher UOHC in the WNP. This indicates that TC-induced ocean cooling is more significant in UOHC than in SST. One possible reason is that the entrainment mixing and upwelling processes can affect the UOHC more directly than the SST.

TC-induced ocean cooling is well known to be caused by entrainment mixing and upwelling. Further detailed analyses for different latitudes show that the UOHC of TCs decreases more than SST in lower-latitude regions (5°–10°N) between TC underlying RI and non-RI duration, and vice versa (Fig. 3). The magnitude of TC-induced ocean cooling generally increases with higher latitude due to a stronger Coriolis force, and the other important reason may be the cold subsurface water in the higher latitudes, which is consistent with the momentum budget in Wu et al. (2007). The effect on the momentum budget of turbulent entrainment from the upper thermocline into the mixed layer would be increased when the Coriolis term becomes large. This study shows that the UOHC and SST are lower in the storm inner-core region. Unfortunately, the low value of the UOHC in the higher-latitude regions from 25° to 30°N may not be appropriate for investigating TC-induced ocean cooling, resulting in the differences between TC underlying RI and non-RI duration in UOHC being less than in SST. In addition, the comparisons of P7/P14 and TC underlying RI/non-RI duration show a similar result with the comparisons of TC underlying RI and non-RI duration. The above results imply that UOHC can better reflect the air–sea interaction than SST in the tropics but SST reflects better in the subtropics. Our results indicate that TC-induced ocean cooling is greater in TC underlying non-RI duration than in RI duration, which is consistent with Lin et al. (2008), who suggested that the deep warm layer can restrain TC-induced ocean cooling during intensification. This also implies that the depth of ocean is more important than the upwelling by wind-driven forcing in the WNP, even though the most intense TCs are those that undergo RI (Lee et al. 2016). Most notably, this is the first study to our knowledge to investigate the different normalized impacts of TC-induced ocean cooling on UOHC and SST. It is noteworthy that TC with RI can still occur when a TC passes the asymmetric UOHC and SST pattern, which
may affect the structure of the TC since some asymmetrically structured TCs can intensify. Further work is needed to analyze the relationship among the asymmetric ocean pattern, ocean depth, ocean upwelling, and the structure of the TC.

In all, future work should include ocean sensitivity experiments designed to assess the relative roles of the UOHC and SST and also the interaction between the atmosphere and ocean by a full-physics high-resolution coupled atmosphere–ocean model. Finally, how the tendency of ocean affects the TC intensity under extreme global warming conditions is also an important scientific issue requiring further research.

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