An Investigation of the Influences of Mesoscale Ocean Eddies on Tropical Cyclone Intensities

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ABSTRACT

The impact of mesoscale ocean eddies on tropical cyclone intensities is investigated based on a combination of observations and atmosphere–ocean coupling simulations. A statistical analysis reveals that the tropical cyclone–eddy interactions occur at very high frequencies; over 90% of the recorded tropical cyclones over the western North Pacific have encountered ocean eddies from 2002 to 2011. The chances of confronting a cold core eddy (CCE) are slightly larger than confronting a warm core eddy (WCE). The observational sea surface temperature data have statistically evidenced that CCEs tend to promote the sea surface temperature decrease caused by tropical cyclones while WCEs tend to restrain such ocean responses. The roles of CCEs are statistically more significant than those of WCEs in modulating the sea surface temperature response. It is therefore proposed that CCEs should be paid no less attention than WCEs during the TC–ocean interaction process. The CCE-induced changes in sea surface temperature decreases are observed to be more remarkable for more intense and slower-moving tropical cyclones and for thinner depth of mixed layers. A set of numerical experiments reveal that the effects of ocean eddies are positively related to their strengths and storm intensities, and the eddy feedback is less pronounced when the eddy is located at one side of storm tracks than right below the tropical cyclone center. The eddy-induced moisture disequilibrium sooner vanishes after the departure of tropical cyclones. The intensity recoveries last for 1–2 days because of the dependence of surface enthalpy fluxes on surface winds.

1. Introduction

The warm ocean serves as the energy source for the development and maintenance of tropical cyclones (TCs), which are one of the most devastating natural disasters in the world (Malkus and Riehl 1960; Simpson et al. 2002). As evidenced in various observations, moving storms can inversely induce evident cooling of the sea surface temperature (SST), typically referred to as cold wake, by bringing cold subsurface water into the sea surface via upwelling, entrainment, and shear-induced vertical mixing (Stramma et al. 1986; Price 1981; Zedler et al. 2002; D’Asaro 2003). Recent studies on TC–ocean interactions have identified that such ocean responses put a strong control on the storm intensity by cutting down upward surface enthalpy fluxes, with the stronger ocean cooling being more influential (Bender and Ginis 2000; Cione and Uhlhorn 2003; Wu et al. 2005; Chen et al. 2010; Lloyd and Vecchi 2011; Halliwell et al. 2015). Nonetheless, a stronger storm does not necessarily lead to stronger SST cooling, since the SST response depends not only on TC properties but also ocean parameters, such as the mixed layer depth and the thermal stratification below the mixed layer (Schade and Emanuel 1999; Lloyd and Vecchi...
2011; Jaimes et al. 2015). Such a nonmonotonic relationship between the SST response and storm characteristics stresses the importance of fully incorporating the upper-ocean thermal information in anticipating the TC intensity evolution rather than taking into account only the SST (Leipper and Volgenau 1972; Gray 1979; Lin et al. 2013; Huang et al. 2015).

The mesoscale ocean eddies are one of the most energetic mesoscale features in the ocean. The subsurface temperatures in mesoscale ocean eddies are generally several degrees larger [warm core eddy (WCE)] or smaller [cold core eddy (CCE)] than those in surrounding water. The thicker and warmer water with downwelling motion in WCE regimes or the thinner and colder water with upwelling motion in CCE regimes can affect the rate of wind-induced turbulent entrainment and upwelling responses (Roemmich and Gilson 2001; Jacob and Shay 2003; Jaimes and Shay 2009, 2010, 2015; Jaimes et al. 2011, 2016). As a result, the WCEs are typically observed to reduce the storm-induced ocean response and the CCEs tend to amplify the ocean response (Shay et al. 2000; Lin et al. 2005; Jaimes and Shay 2009). Corresponding to the eddy feedback, the TCs tend to be more intensified when moving over WCEs while less intensified when confronting CCEs, which are accomplished primarily by modulating the transport of air–sea enthalpy fluxes (Ali et al. 2007; Jaimes and Shay 2009; Patnaik et al. 2014; Jaimes et al. 2016). Patnaik et al. (2014) investigated the interactions between ocean eddies and eight TC cases over Bay of Bengal. They found that surface enthalpy fluxes are evidently increased by WCEs and decreased by CCEs. The roles of WCEs have been especially noted as rapid intensifications occurred during the time the typhoons or hurricanes passed over WCEs (Hong et al. 2000; Shay et al. 2000; Lin et al. 2005; McTaggart-Cowan et al. 2007). Based on an idealized case study, Chan et al. (2001) depicted the intensity response of a TC to a preexisting WCE. The storm intensified as it reached the WCE edges, and as it exited the WCE regime the weakening process did not occur immediately. The simulations from a simple hurricane–ocean coupled model in Wu et al. (2007) also reflected that the storm intensified as it entered the WCE regime and recovered its intensity gradually after leaving the WCE (their Fig. 13). Lin et al. (2005) analyzed the rapid increase in the intensity of Typhoon Maemi (2003) by passing a giant WCE, and suggested that the WCE acts as an effective insulator between the TC and deeper ocean cold water. Jaimes et al. (2015, 2016) further proposed that the enhanced local buoyant forcing from the ocean could be an important intensification mechanism in TCs over WCE regimes. Although most of the existing literature records more intensified TCs after crossing over WCEs via the notably diminished SST decrease from observations or numerical modeling (e.g., Bao et al. 2000; Emanuel et al. 2004; Jaimes and Shay 2009; Vianna et al. 2010; Lin et al. 2011), Yablonsky and Ginis (2013) suggested that the circulation of a WCE can also produce nonnegligible impacts on the TC intensity, and that a WCE located to the right of the TC could even create an unfavorable condition by advecting the cold water into the TC inner core.

Although there have been increasing endeavors in understanding the contributions of WCEs, the roles of CCEs in TC evolution have nonetheless drawn comparatively less attention. This is possibly because the rapid intensification of TCs has a quantized criterion (Kaplan and DeMaria 2003), but the slowed intensification phase caused by CCEs cannot be easily identified from the other environmental factors. Virtually the equal magnitudes of decrease in mixed layer depth would produce more pronounced influences on storm intensities than those of increase in mixed layer depth as indicated from Schade and Emanuel (1999, their Fig. 6), which has been also speculated by Lin et al. (2005). Additionally, the counts of CCEs are comparable to or even slightly larger than those of WCEs in the western North Pacific (not shown). Ma et al. (2013a) investigated the impact of a CCE on the TC intensity and structures based on idealized atmosphere–ocean coupling simulations. Contrary to the roles of WCEs, they found that the storm began to be weakened as it encountered the CCE, the intensity reduction attained a maximum shortly after passing over the CCE center, and the intensity underwent a time period of recovery after leaving the CCE.

The foregoing studies highlight the importance of WCEs and CCEs during TC evolutions. Most of the previous observational or modeling studies are based on single-case or limited-number eddies (e.g., Wu et al. 2007; Ma et al. 2013a; Patnaik et al. 2014). It remains unclear to what frequency the TCs interact with eddies in real-world situation, and to what an extent the findings based on limited case studies are applicable for most of the ocean eddies. Besides, little attention has been paid to investigate how the feedback of ocean eddies is sensitivity to the properties of TCs and ocean eddies. Therefore, this study attempts to further examine the impacts of ocean eddies on TC intensities under different situations, by using a combination of observation statistics and idealized simulations. Section 2 gives the statistically observational results of TC–eddy interactions. Section 3 introduces the atmosphere–ocean coupled model setup and
experimental design. The modeling results of TC–eddy interactions are presented in section 4, followed by the summary in section 5.

2. Observations

It has been a developing consensus that ocean eddies can effectively influence storm intensities through modulating the ocean response. Lin et al. (2005) proposed that abundant TC–eddy interactions may occur as storms cross over “two eddy-rich zones.” The advancement in remote sensing and in situ measurements renders it feasible to investigate various subsurface ocean systems (Pickard and Emery 1990; Qiu 1999). By merging the satellite altimeter data from the Ocean Topography Experiment (TOPEX)/Poseidon (T/P) (followed by Jason-1, and presently by Jason-2) and the successive European Remote-Sensing Satellites-1 and -2 (ERS-1 and ERS-2) altimeters, Chelton et al. (2007, 2011) have constructed high-resolution (approximately 40 km) global sea surface height fields at 7-day intervals to track the mesoscale ocean eddies, with the eddy information of property, center position, and radius incorporated. The eddy data are available from October 1992 to April 2012 (online at http://cioss.coas.oregonstate.edu/eddies/). The SST could be the most important indicator for ocean eddies to couple with the TCs. The SST data come from the NOAA 0.25° daily optimum interpolation SST (Reynolds et al. 2007) produced by the Advanced Very High Resolution Radiometer (AVHRR) and Advanced Microwave Scanning Radiometer (AMSR) (available online at https://www.ncdc.noaa.gov/oisst), which are capable of measuring the SST through clouds. It should be noted that the SST dataset is available since 2002, which covers a shorter time range compared to the eddy dataset. To keep the consistency, all the data used cover the same time range from 2002 to 2011. The ocean mixed layer depth data come from the NCEP Global Ocean Data Assimilation System (GODAS; http://www.esrl.noaa.gov/psd/). The storm best track data, including the information of storm center location, intensity, and size (denoted by the radius of the last closed isobar), come from the Joint Typhoon Warning Center (JTWC; http://www.usno.navy.mil/NOOC/nmfc-ph/RSS/jtwc/best_tracks/wpindex.php).

Figure 1 exhibits a sample of distributions of TCs in the year of 2005 and ocean eddies in August 2005 over the western North Pacific. The ocean eddies are shown to be omnipresent in the vast ocean, with the WCEs and CCEs distributed almost evenly. The CCEs slightly outnumber the WCEs (not shown). Like the TCs, there are relatively few eddies generated in low latitudes (Chelton et al. 2007), possibly related with the small Coriolis force there. Since ocean eddies are quasi-permanent and quasi-stationary features (Chelton et al. 2007) relative to TCs, it is foreseeable that most TCs will inevitably confront ocean eddies during their lifetime. The percentiles of different ocean eddy radii confronted by TCs over the western North Pacific from 2002 to 2011 are summarized in Fig. 2. The eddy radius terms are defined by Chelton et al. (2007) to be the radius of a circle whose area is equal to that enclosed by the contour of maximum circum-average speed. The ocean eddies have radii ranging from less than 60 km to exceeding 360 km. Most eddies seem to have typical radii approximately from 90 to 180 km, which are comparable to the inner-core size of TCs.
Hence, the intensities of passing storms could be affected by the altered inner-core surface enthalpy fluxes associated with preexisting ocean eddies. Figure 3a displays the percentages of storms over the western North Pacific that encountered ocean eddies from 2002 to 2011 before weakening or landfall, classified by the number of eddies. Here the TC–eddy interaction is defined to occur as the TC center encounters the outer edges of ocean eddies, namely when the distance between the TC center and the eddy center is equal or less than the eddy radius. This criterion is relatively rigorous, and the interaction frequency should be higher if selecting a looser one. A somewhat unexpected result is that TCs interact with eddies at unusually high frequency, that over 90% of the TCs have encountered ocean eddies during their lifetimes, wherein over 70% have encountered more than one eddy. As for the WCE, the numbers of 1 and 2 account for over a half of the storms, while about 35% of the TCs did not cross over a WCE. The interactions between TCs and CCEs show a similar behavior, except that the TCs that did not encounter CCEs occupy a smaller portion (about 17%). This is probably because there are slightly more CCEs than WCEs in the western North Pacific (not shown). The statistical result reveals that most TCs will inevitably confront CCEs or WCEs during their lifetime, with the chances of confronting CCEs slightly larger than those of WCEs. It is therefore essential to identify how important the ocean eddies are for changing the TC-induced ocean responses. Figure 3b gives the relative locations between eddies and storms when interactions occurred. The TC is considered to pass over the eddy center when the closest distance between the eddy and storm centers is less than one-third of the eddy radius. By this division, the eddy diameters are equally separated into three parts. The right or left directions of eddies are obtained based on the eddy location and the storm translation direction at a 12-h interval. The proportions of left or right interactions are found to be relatively evenly distributed for either WCEs or CCEs, being approximate 40%. Of interest is that the occasion that storms encountered eddy centers occupies a fraction of only 20%, which is notably smaller than the fractions of the other occasions. Many of the present modeling studies have straightforwardly focused on the central interactions between storms and ocean eddies (e.g., Chan et al. 2001; Wu et al. 2007; Ma et al. 2013a). Hence, it may be revealing to examine the feedback of ocean eddies when they are located at one side of storms.

The SSTs along TC tracks are obtained by the bilinear interpolation of the SST data based on the 6-h-interval best track data. The SST changes are calculated by the difference of the averaged SST 12–2 days after the storm passage minus the SST 2 days before the storm passage as similarly done in Lloyd and Vecchi (2011). The storm translation speed is calculated based on the distance of storm centers at 6-h interval. The available data counts for different TC–ocean interaction conditions at 6-h interval are listed in Table 1. To examine the relation of SST response to the storm intensity and translation speed, the composite decreases of SST are displayed in Fig. 4. The SST data (Table 1) are interpolated as

![Fig. 3. Percentages (%) of (a) different number of the WCEs (red), CCEs (blue), and WCEs + CCEs (black) encountered by storms and (b) relative locations of eddies related to the storms from 2002 to 2011 over the western North Pacific.](image-url)
functions of maximum surface winds and translation speeds using the Cressman method. The WCE condition, CCE condition, and common water condition as well as their discrepancies are compared. There are fewer data when the translation speed is large (not shown), which may be less representable. The TCs are shown to cause SST decreases in all three conditions, and the SST anomalies overall decrease with the storm moving speed and increase with the storm intensity. This is consistent with the modeling results of Schade and Emanuel (1999) and observational results of Mei and Pasquero (2013). As in previous single-case observations or simulations, the WCEs have overall inhibited the SST response (Fig. 4d) while the CCEs have promoted the SST cooling relative to the common water (Fig. 4e). In addition, the roles of CCEs are shown to be more pronounced for slower and stronger TCs.

To further examine the influences of ocean eddies on SST responses, the box plots of SST decreases are shown in Fig. 5, with the dataset classified based on the storm translation speed and intensity. The storm translation speeds less than the 6 m s$^{-1}$ threshold are denoted as slow moving, and larger speeds than this threshold are denoted as fast moving. A threshold of 5 m s$^{-1}$ has given similar results (not shown). The category 0 data counts are much larger than the sum of the other categories, and the storms at category 0 are denoted as weak, while the other category 1–5 storms are...
together denoted as intense. The averaged SST decrease in the common water is near to that observed in North Atlantic basin [Fig. 4 in Lloyd and Vecchi (2011)]. The SST decrease is shown to be larger for slow-moving storms relative to fast-moving storms, and smaller for weak storms relative to intense storms in all three conditions. The SST decrease in CCE condition is uniformly larger than that in the common water condition independent of storm intensities and translation speeds. Such a relationship is significant above the 99% confidence level based on the Student’s t test for all conditions. The smaller SST decrease for the WCE than the common water is significant above the 95% and 98% confidence levels based on the Student’s t test for the slow-moving condition and intense storm condition, respectively.

The background SST and depth of the mixed layer are crucial for the tropical cyclones intensities, which can inversely bring changes in the SST response (Fisher 1958; Schade and Emanuel 1999; Mao et al. 2000). Figure 6 displays the SST decreases as functions of SST and mixed layer depth for WCE, CCE, and common water conditions as well as their differences. The background depth of mixed layer is calculated by averaging the GODAS data over a 2° × 2° area centered on the TC track location. On the whole the SST anomalies are shown to increase as the mixed layer depth becomes thicker, which can inhibit deeper cold water from being entrained upward (Price 1981). The discrepancy of SST decrease between WCE and common water conditions suggests that the WCEs have overall uniformly suppressing effects on the SST response independent of SST values and mixed layer depth. The discrepancies of SST decreases between CCE and common water conditions indicate that the CCEs tend to amplify the SST response, which is more pronounced when the depth of mixed layer is thin or the SST is low. The boxplots shown in Fig. 7 have given consistent results. The larger SST decreases for CCE condition than common water condition are all significant above the 99% confidence level. Their difference of mean SST decrease is larger when the mixed layer depth is thinner. The relationship between the SST decrease and the specific SST value is not as evident as the depth of mixed layer. The WCEs are also capable of modulating the SST response, which has passed the
significance test, but to a lesser extent than the CCEs as reflected in Fig. 5.

Figure 8 shows the evolution of mean composite SST decrease for WCE, CCE, and common water conditions. The SST decrease is calculated by the SST on each specific day minus the SST 12 day earlier. The day 0 denotes the day when TCs pass over the location. The SST evolves slowly from −12 to −2 days, and when the TC approaches to the location, there appears a rapid cooling, reaching a minimum 2 days after the storm passage; afterward, the SST recovers slowly. This behavior of SST evolution is very similar to that in Lloyd and Vecchi (2011) in other ocean basins. By comparing the WCE, CCE, and common water conditions, the CCEs have given the largest mean SST decrease, exceeding 1.2°C, followed by the common water, less than 0.9°C, and the WCEs have produced the weakest SST decrease, about 0.7°C. Such a relationship evidences that the CCE tends to boost the TC-induced SST response and the WCE tends to inhibit the SST response, while the modulating effects of CCEs are overall more salient than those of WCEs.

3. Coupled model and experimental design

a. Model setup

The above statistical observation analysis does not incorporate subsurface eddy information, and the coarse spatial and temporal resolutions have prevented a more in-depth investigation of the eddy feedback. In this section the impact of ocean eddies on TC intensities and structures is further explored by conducting numerical experiments. The model setup basically follows that in Ma et al. (2013a), and a brief introduction is given below. The atmosphere–ocean coupled model is an upgraded version, which consists of the Weather Research and Forecasting
WRF Model, version 3.5 (Skamarock et al. 2008), as the atmospheric component and the Princeton Ocean Model (POM; Meller 2004) as the oceanic component. These two model components transfer SST, surface wind stress, surface sensible and latent heat fluxes, and shortwave and longwave radiation fluxes through the Model Coupling Toolkit (MCT; Larson et al. 2005). The coupling interval keeps the same as the time step of atmospheric (outmost domain) and oceanic models, specified as 90 s in this study. Both the atmospheric and oceanic models are set on the $\beta$ plane with the domain centered at 20°N.

For the WRF configuration, the Yonsei State University (YSU) scheme (Hong et al. 2006) and the corresponding Monin–Obukhov scheme are used to parameterize the boundary layer and surface layer processes. The Lin scheme (Lin et al. 1983) is chosen to handle the microphysical processes. The outmost domain utilizes the Betts–Miller–Janjić scheme (Betts and Miller 1986) to parameterize the cumulus convections. There are two meshes with dimensions of $220 \times 220$ and $202 \times 202$, and resolutions of 15 and 5 km, respectively. Because a large number of simulations will be conducted, a relatively coarse resolution of 5 km for the inner domain has been applied as a compromise with the computation efficiency. Although smaller-scale internal processes and high-number asymmetries could not be well resolved at such a resolution, it is expected that this resolution has the capability of properly reproducing the ocean feedback on the evolution of TC intensities (Gentry and Lackmann 2010).

The POM is configured with a single domain of $219 \times 219$ grid points at a resolution of $0.165^\circ \times 0.165^\circ$, covering a region slightly larger than WRF. A total of 23 vertical levels are used, with 11 levels being placed in the upper 100 m. The Mellor–Yamada turbulence closure scheme (Mellor and Yamada 1982) is embedded in the POM model to provide vertical mixing coefficients. An Arakawa C grid is utilized as the horizontal finite-difference scheme. The horizontally homogeneous temperature and salinity profiles for POM initialization and boundary conditions are obtained from the monthly averaged (August) Simple Ocean Data Assimilation (SODA) profiles at (20.25°N, 176.25°E), except that the temperature of the upper ocean mixed layer and the SST

![Fig. 7](https://example.com/fig7.png)  
As in Fig. 5, but for (a) thin and deep mixed layers and (b) low and high SST. The thresholds for the mixed layer depth and SST are 40 m and 29°C, respectively. The larger SST decrease for the CCE than the common water is significant above the 99% confidence level based on the Student’s $t$ test for all conditions. The smaller SST decrease for the WCE than the common water is significant above the 90% and 99% confidence levels for the thin and deep mixed layers, and significant above the 90% and 95% confidence levels for the low and high SSTs based on the Student’s $t$ test.

![Fig. 8](https://example.com/fig8.png)  
Lagrangian composite SST decrease ($^\circ$C) relative to 12 days before the storm passage day for WCE, CCE, and common water.
is modified slightly to be 29°C so as to keep the same as the initial SST in WRF Model. The entire domain is set to be ocean with a uniform depth of 2500 m. For simplicity, the POM model is initialized with quiescent current to keep the constructed eddy quasi stationary. The atmospheric and oceanic model output is saved every 1 h.

**b. Initialization and experimental design**

For the atmospheric model, the uniform easterly winds of 3 m s$^{-1}$ are specified at all levels, with geopotential heights adjusted according to the geostrophic wind balance. A Rankine vortex in hydrostatic and gradient wind balance is then implanted in the environment with a maximum wind speed of 24 m s$^{-1}$ at a radius of 125 km. The atmospheric model is integrated individually for 24 h as the model spinup. The radius of 15 m s$^{-1}$ surface wind for the spunup storm gives a value of 212 km (not shown), which is in the medium-size range (Lee et al. 2010). The ocean temperature anomalies between eddy center and the common water are generally several degrees large (Roemmich and Gilson 2001; Lin et al. 2005) and the eddy radii range typically from less than 90 to 180 km (Fig. 2). Because of the lack of subsurface observations, the ocean eddies are constructed by artificially assigning specified initial departures from the surrounding ocean temperature referring to the feature-based methodology (Yablonsky and Ginis 2008) as in Ma et al. (2013a). The temperature departure is set to be maximized at approximately a depth of 400 m and decreases in the upper and lower levels gradually, to keep the temperature horizontally homogeneous at the sea surface and below the depth of 1700 m approximately. The salinity profile in the ocean eddies stays the same as that in the common water. The ocean eddy is then initialized by integrating the POM individually for 96 h without external forcing. During the spinup period of the ocean model, the density and current are adjusted geotropically to a steady state progressively.

A total of 11 experiments have been conducted, with their descriptions summarized in Table 2. All coupling simulations were integrated for a total of 96 h. The experiment with the horizontally homogeneous ocean state is considered as the control run (CTRL). Since the ocean eddies could possess different strengths and the TC interacts with ocean eddies in complex ways (Figs. 1 and 3), different conditions of the storm–eddy interaction have been considered. This includes the eddies with different thermal properties (WCE or CCE) and strengths (strong CCE and weak CCE), the storms at different intensification phases (24 and 48 h), different relative locations (eddies in central, right, and left sides of the storm track), and interactions with two eddies during the storm lifetime (CCE at 24 h and WCE at 60 h as well as WCE at 24 h and CCE at 60 h). Two experiments are conducted to examine the effects of different eddy locations, with one the WCE being located at the right of the storm track and the other one the CCE being located at the left of the TC track. To compare the relative importance of ocean eddies and other ocean parameters, one experiment with a deep mixed layer as in WCE center have been conducted.

**Figure 9** shows the initial upper-ocean temperature profiles in the ocean environment and center of ocean eddies for the coupling simulation. They are obtained by individually integrating the POM for 96-h spinup.
to a steady state before the coupling simulation. The maximal temperature anomaly between the eddy center and the common water reaches 6.5°C for the CCE. The CCE has shallower mixed layer than the surrounding water, and the weak CCE possesses a temperature profile between the CCE and the common water. The mixed layer in the WCE center is about 60 m in depth, which is approximately twice of that in common water. The DEEPML run has the same mixed layer as the WCE center but a sharper thermocline, so that its low-level temperature stays the same as that in the common water. This run is somewhat similar to the “perpetual ocean eddy” simulation in Wu et al. (2007) since the storm is always in a thick mixed layer environment. The initial horizontal temperature and current fields of the steady-state ocean eddies at 45 m in

Fig. 10. The initial horizontal temperature (°C, shaded) and sea current (m s⁻¹, vector) at 45-m depth for the steady-state (a) WCE, (b) CCE, and (c) weak CCE. (d) Schematic for the location of eddies relative to the storm track.
depth are displayed in Fig. 10. As shown in Fig. 9, the WCE possesses evidently positive temperature anomalies than the surrounding water (Fig. 10a), while the CCE shows negative anomalies (Fig. 10b). The current velocities for both eddies are on the order of $\sim 1 \text{ ms}^{-1}$, which is significantly larger than the eddy translation speed (Chelton et al. 2011). The constructed eddies in Figs. 10a and 10b are shown to possess relatively intense strengths. The temperature anomalies in the weak CCE (Fig. 10c) are smaller than those in the strong CCE, suggestive of its weaker strength, and correspondingly the current fields are less vigorous. The radius of constructed eddies is about 100 km, which is within the typical range of eddy sizes in the western North Pacific (Fig. 2; Roemmich and Gilson 2001). Figure 10d shows a schematic of the relative locations between the storm and the eddy during their interactions, which could provide implications for the discrepancies of ocean–eddy feedback caused by relative locations between TCs and eddies.

4. Model results

a. TC intensity

The simulated storm intensities in terms of the minimum sea level pressure ($P_{\text{min}}$) and maximum azimuthally averaged surface winds are summarized in Fig. 11 to compare the intensity discrepancies for WCEs, CCEs, and other experiments. The $P_{\text{min}}$ and maximum surface wind speeds basically give consistent discrepancies among all runs, but the surface winds undergo more severe fluctuations. All storms experience an overall intensifying trend in the first 48 h, and afterward show small intensity variations in the later stage except CCE24WCE60 and WCE24CCE60. Since TC intensity is sensitive to the ocean state (e.g., Torn 2016), there are large discrepancies among these runs. Therefore, the differences between CTRL and the sensitivity experiments can give implications for the impact of ocean eddies under different conditions as well as the relative influences of ocean parameters.

From the central interaction case—namely, the TC center crosses over the eddy center—the storm intensification begins to be hindered when encountering a CCE (CCE24 and CCE48), and be amplified when encountering a WCE (WCE24 and WCE48). After
departure from the eddy, the storm experiences a period of recovery, and eventually it basically returns to its steady-state intensity as in CTRL. Such an intensity response coincides with previous modeling studies (Wu et al. 2007). Lin et al. (2005) have proposed a concern as to whether the ocean eddy can contribute to intensity maintenance, which is important for evaluating the potential risk of active TCs. These simulations seem to indicate that ocean eddies impose transient influences on TC intensities, since the TC interacts with the eddy transiently. The intensity changes caused by the CCE are shown to be greater than that those by the WCE, as a result of larger temperature anomalies for the CCE. A comparison of CCE24 with CCE48, or WCE24 with WCE48, reveals that the more intensified the storm is, the more pronounced the modulation effects of eddies are. This agrees well with the observational results (Fig. 5b), which is because stronger storms tend to induce more severe ocean response (Schade and Emanuel 1999), so that the roles of CCEs or WCEs become more salient. Another common feature of the eddy experiments is that the changes in storm intensities become most significant slightly after the storm moves over the eddy center. Table 3 lists the largest positive or negative differences of $P_{\text{min}}$ between the sensitivity experiments and CTRL as well as the corresponding simulation time. The largest intensity deviations are shown to occur about 9–12 h delayed.
after the storm passes over the eddy center as similarly found in Ma et al. (2013a).

With smaller temperature anomalies and weaker currents, the intensity change caused by the weak CCE is much reduced relative to that by the strong CCE. This evidences a simple guideline that the modulation effects of eddies on the ocean response are proportionally related to the eddy intensities. The largest $P_{\text{min}}$ deviation for the weak CCE occurs much later after the storm moves over the eddy, which may be a consequence of the storm structural fluctuations caused by the eddy. The cooling effect of CCEs could also be similar as that of the cold wake in thermally stabilizing the boundary layer, which has a positive impact on TC intensity and partly offsets their weakening roles (Lee and Chen 2014). The experiments with ocean eddies on one side of the TC also show that the WCE is beneficial while the CCE is detrimental for the storm intensification, but the effect of WCE or CCE is weaker relative to that right blow the TC center (Fig. 11). For real-world TC–eddy interactions, these occasions have taken place more frequently than the central interaction cases (Fig. 3b).

Since most TCs experience more than one eddy during their lifetime (Fig. 3a), two runs with the storm passing over a WCE after a CCE, and a CCE after a WCE are conducted. Agreeing well with the central interaction cases, the storm becomes weakened first and intensifies afterward in CCE24WCE60, and becomes intensified first and weakens afterward in WCE24CCE60. The changes in $P_{\text{min}}$ caused by the ocean eddy can be as large as 9 hPa. Therefore, strong ocean eddies can effectively modulate the intensification of storms, for instance, boosting or suppressing the phase of rapid intensifications (e.g., Lin et al. 2005). Among all the runs, the DEEPML experiment has produced the most intense storm throughout the simulation, with its largest $P_{\text{min}}$ difference with CTRL exceeding 10 hPa. This indicates that the WCE could be less determinative than the deep background mixed layer in TC intensification, since the WCE only influences the TC for a short time period.

b. Ocean response

The storm in CTRL moves northwestward slowly at approximately translation speeds of 2–4 m s$^{-1}$, and leaves behind an asymmetric SST decrease with a width of roughly 100 km (Fig. 12). The other experiments simulate similar storm tracks (not shown). From a sample of the plan views of SST difference when the storm passes over the eddy center, the WCE is shown to be capable of suppressing the ocean response while the CCE is prone to boost the SST cooling (Figs. 12b,c), which concurs well with the observational results (Fig. 8). The SST change is more effectively affected by the CCE associated with its larger temperature anomalies. The simulated SST has shown some small-scale noisiness, which may cause small-scale structural changes...
of TCs. Nonetheless these asymmetric features may not bring much influence on TC intensities since they play a minor role in TC intensification (Nolan et al. 2007). The simulated TC intensity responses to the ocean features are reasonable as expected (Fig. 11) and basically consistent with previous simulations (e.g., Hong et al. 2000; Chan et al. 2001). In responses to the SST, the upward surface latent heat fluxes are increased by the WCE and decreased by the CCE as well (Fig. 13). Compared with the cold wake, the ocean eddy can locally change the inner-core surface latent heat fluxes. The changes in surface latent heat fluxes are again larger by the CCE than the WCE, which is responsible for the correspondingly larger storm intensity deviations (Fig. 11 and Table 3).

Figure 14 shows the radial distributions of SST decrease and surface latent and sensible heat fluxes for CTRL, CCE24, and CCEWeak at 24h. As the storm moves over the eddy, the maximal SST decrease occurs in the storm center, with magnitudes of 2.5°C for the strong CCE and 1.5°C for the weak CCE. The difference of SST decrease between CCE24 and CTRL is several times larger than that between CCEWeak and CTRL. Consistent with the SST relationships, the decrease of surface latent and sensible heat fluxes is remarkably alleviated in the weak CCE run. This leads to a much less weakened storm, implying that for the ocean eddies with weak intensities, their modulations on TC intensifications could be less significant.

The SST decreases are azimuthally averaged in four quadrants relative to the storm translation direction (Fig. 15). The difference of SST change between WCERight and CTRL is overall larger in the two right quadrants than the two left quadrants, but WCERight gives overall diminished SST decrease in all quadrants. Therefore, the WCE located at the right of the storm track is instrumental for the storm intensification in WCERight (Fig. 11). Comparing WCERight with WCE24, the roles of WCEs are shown to be less notable when the storm crosses over the eddy edges.
than the eddy center. As the storm encounters the CCE located at the left side, the SST decrease is amplified in all quadrants. In the left-rear and left-front quadrants the CCE even causes greatly larger SST cooling outside the storm center than the CCE24 run. This nonetheless does not result in a weaker storm relative to CCE24, which is because the symmetric cooling plays a decisive role in weakening the storm intensity (Wu et al. 2005). These results reveal that when an ocean eddy is located at one side of the storm track, the eddy effects on the TC intensity could be reduced, but the role of its current field may not be necessarily more significant than its temperature field as in Yablonsky and Ginis (2013).

The Hovmöller diagram of the SST decrease for CTRL and WCE24CCE60 as well as their difference is shown in Fig. 16 to demonstrate the impacts of multiple eddies on ocean responses. The SST cooling in CTRL increases progressively as the storm strengthens. In WCE24CCE60, the SST in the storm inner core does not show evident decrease before 24 h when a WCE is present, but decreases rapidly as the storm enters the CCE region. This is more apparently shown from the difference between WCE24CCE60 and CTRL (Fig. 16c). After departure from the WCE, the storm begins to recover its intensity gradually, but this recovery shortly ceases because the storm confronts a preexisting CCE afterward (Figs. 11 and 16c).

c. Surface heat fluxes

The surface latent and sensible heat fluxes in the WRF Model are computed based on bulk formulas:

\[ \text{LH} = \rho L_v C_q U_a (q_s - q_a) , \quad \text{(1)} \]

\[ \text{SHX} = \rho c_p C_h U_a (\theta_g - \theta_a) , \quad \text{(2)} \]

where LH denote the surface latent heat fluxes and SHX denote the surface sensible heat fluxes; \( \rho \) is the air density in surface layer; \( L_v \) is the latent heat of vaporization; \( C_h \) and \( C_q \) are the surface exchange coefficients for heat and...
moisture, respectively; $U$ is the horizontal wind speed; $q$ is the mixing ratio of water vapor; and $\theta$ is the potential temperature. The subscripts $a$ and $g$ signify the lowest model level and bottom surface, respectively. To investigate how the ocean eddies influence the surface enthalpy fluxes, both surface latent heat fluxes and sensible heat fluxes, as well as the primary components composing surface latent heat fluxes, including $U_a$, $q_s - q_a$, $q_a$, and $q_w$, are shown in Figs. 17 and 18 for CTRL and its difference with CCE48 and WCE48, respectively. The surface latent heat fluxes are linked to the SST via the saturated water vapor mixing ratio $q_s$. The surface latent and sensible heat fluxes in CTRL increase as surface winds strengthen roughly in the first 48 h, and fluctuate in the later stage of the simulation. As the storm approaches the CCE regime, the $q_s$ begins to be diminished accompanying the SST decrease (Figs. 17e and 12c), with a minimum in the storm center. As a consequence, the $q_s - q_a$ is reduced in the storm region. It is the thermodynamic disequilibrium between the sea surface and surface air and corresponding moisture disequilibrium that fundamentally determine the surface enthalpy fluxes, rather than the specific values of SST (Jaimes et al. 2015). The changes in $q_s - q_a$ have led to evident reduction in surface latent heat fluxes, and eventually the storm intensification is suppressed. The correlation between surface sensible heat fluxes and thermodynamic disequilibrium gives similar behaviors (not shown). The CCE also diminishes the surface layer water vapor (Fig. 17f) by enhancing SST cooling, which inversely has functions of offsetting the CCE-induced changes in surface latent heat fluxes [Eq. (1)]. Since the cooled surface air is continuously mixing with surrounding air parcels (Ma et al. 2013b), the decrease of surface layer water vapor vanishes sooner after leaving the cooling region of the CCE. A remarkable feature reflected by Fig. 17 and also by Fig. 18 is that the surface latent heat flux changes do not recover instantaneously after the moisture disequilibrium changes vanish. This is shown to be caused by the mutual dependence between surface latent heat fluxes and surface winds (Figs. 17a,c and 18a,c). When a storm has left the CCE, the background sea surface is capable of supplying larger surface enthalpy fluxes to sustain greater intensity (Emanuel 1995), but the surface winds remain weak and hence a time period of readjustment is required for the storm to fully recover. Similar results can also be found for the difference between WCE48 and CTRL, except that the WCE-induced changes in surface enthalpy fluxes and storm intensity are less significant associated with its comparatively weaker strength relative to the CCE (Figs. 10a,b).

5. Summary

This study has investigated the roles of mesoscale ocean eddies during TC–ocean interactions by a combination of statistical analysis of observations and a set of numerical experiments. Unlike previous studies focusing on single-case or limited-case TC–eddy interactions, the best track data and eddy observations are combined to obtain hundreds of TC–ocean interactions from 2002 to 2011 over the western North Pacific. The SST decreases caused by TCs are found to become noticeable with TCs approaching, be maximized 2 days after the TC passage (Lloyd and Vecchi 2011), and be relatively larger for slow-moving or intense TCs than fast-moving or weak TCs (Mei and Pasquero 2013). Since the observations are not enough for a more

Fig. 16. Hovmöller diagram of the azimuthally averaged SST decrease (°C) for (a) CTRL, (b) WCE24CCE60, and (c) the difference between WCE24CCE60 and CTRL.
in-depth exploration due to lack of subsurface ocean information and coarse spatial and temporal resolutions, a series of simulations are carried out to compare the changes in TC intensities caused by ocean eddies with various properties. These involve different eddy properties (WCE and CCE), eddies with different strengths, TCs confronting eddies at different intensities, eddies located at different locations, and confronting more than one eddy.

The statistics of TC–ocean eddy interactions from 2002 to 2011 have quantitatively demonstrated that TCs over the western North Pacific interact with mesoscale ocean eddies at very high frequencies. Over 90% of historical TCs have encountered ocean eddies during their lifetimes, and over 70% of them have encountered more than one eddy. The chances of confronting a CCE are slightly larger than those of confronting a WCE. The observational SST data have statistically evidenced the dependence of storm-induced SST cooling on the ocean eddies, in that CCEs are overall beneficial for amplifying the TC-induced SST cooling, while WCEs tend to suppress the TC-induced SST response. These are reconfirmed by the numerical modeling results. The modulation effects of CCEs on SST responses to TCs are overall more prominent than those of WCEs. The roles of ocean eddies are also sensitive to the characteristics of TCs and the upper ocean. The CCE effects are evidently more significant for slow-moving TCs than fast-moving TCs, intense TCs than weak TCs, and thin mixed layer than deep mixed layer.

Numerical simulations suggest that the influences of ocean eddies are positively related to their strengths, in that the changes in SST response as well as the intensity change caused by a weak CCE are much smaller than an
intense CCE. The feedback of ocean eddies is also less pronounced when they are located at one side of the TC track than right below the TC center. Mesoscale ocean eddies merely impose transient influences on TC intensities. After departure from the eddy, the TC experiences a recovery period of 1–2 days. This is because although the eddy-induced changes in air–sea moisture disequilibrium sooner vanish after leaving the eddy, the surface heat fluxes remain reduced so that a time period of readjustment is required for the surface winds to fully recover.

Previous studies on interactions between ocean eddies and TCs have illustrated the physical mechanisms of ocean eddy responses to TCs (e.g., Jaimes and Shay 2010, 2015; Jaimes et al. 2011), stated the notable eddy feedback caused by severe TCs (e.g., Patnaik et al. 2014), suggested the positive roles of WCEs and negative roles of CCEs in TC intensification (e.g., Chan et al. 2001; Ali et al. 2007; Jaimes and Shay 2009; Patnaik et al. 2014; Jaimes et al. 2016), proposed an intensification mechanism in TCs over WCEs (e.g., Jaimes et al. 2015, 2016), and recorded the rapid intensification of TCs after passage over WCEs (e.g., Lin et al. 2005; McTaggart-Cowan et al. 2007). In this study a large amount of TC-eddy interaction data and a series of numerical simulations have been considered. The primary new findings of this study are summarized as follows: 1) the interactions between eddies and TCs have been quantified and give a very high frequency over the western North Pacific, 2) the roles of CCEs are statistically more significant than those of WCEs in modulating the TC-induced SST response, 3) the effect of CCEs is observed to be more salient for slow-moving or intense TCs and for thin mixed layer background, and 4) the eddy feedback is positively related to the eddy strength and the TC intensity. It is recommended that

![Fig. 18. As in Fig. 17, but for the difference between WCE48 and CTRL.](image-url)
CCEs should be paid more or no less attention than WCEs during TC–ocean interaction processes. Besides, the conventional notion of the positive impact of WCEs and negative impact of CCEs on influencing SST responses, which are primarily obtained by single-case or limited-case studies, has been evidenced to be applicable to most of the mesoscale ocean eddies in the real world.

Finally, this study has statistically stressed the importance of mesoscale eddies on modulating the ocean response caused by TCs. Observations and simulations demonstrate that the modulating effects of eddies are also sensitive to the characteristics of TCs and ocean eddies. More high-resolution observations and simulations are expected for a further in-depth investigation of the TC structural changes caused by ocean eddies.

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