The Impact of Storm-Induced SST Cooling on Storm Size and Destructiveness: Results from Atmosphere–Ocean Coupled Simulations

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ABSTRACT

In this study, both an atmospheric model [Weather Research and Forecasting (WRF) model] and an atmosphere–ocean (WRF)–ocean (Princeton Ocean Model; POM) coupled model are used to simulate the tropical cyclone (TC) Kaemi (2006). By comparing the simulation results of the models, effects of oceanic elements, especially the TC-induced sea surface temperature (SST) cooling, on the simulated TC size and destructiveness are identified and analyzed. The results show that there are no notable differences in the simulated TC track and its intensity between the uncoupled and coupled experiments; however, there are large differences in the TC size (i.e., the radius of gale-force wind) between the two experiments, and it is the TC-induced SST cooling that decreases the TC size. The SST cooling contributes to the decrease of air–sea moisture difference (ASMD) outside the TC eyewall, which subsequently leads to the decreases in surface enthalpy flux (SEF), radial sea-level pressure gradient, absolute vorticity advection, and wind speed outside the TC eyewall. As a result, the TC size and size-dependent TC destructive potential all decrease remarkably.

Key words: tropical cyclone (TC) size, destructiveness, sea surface temperature (SST) cooling

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1. Introduction

Tropical cyclone (TC) is one of the most severe weather systems on the earth, which brings about incalculable consequences to human life (Easterling et al., 2000; Peduzzi et al., 2012). Currently, TC forecasts generally focus on the TC track and intensity. Meanwhile, the potential wind damage near the TC center is also estimated. The TC intensity is expressed by either the minimum sea-level pressure or the maximum sustained wind speed (MWS) near the surface TC center. Because track and intensity are two important TC properties, many studies have been conducted to investigate the physical processes responsible for the TC intensity and track changes (Elsner et al., 2008; Kossin et al., 2014, 2016). Nonetheless, a broader understanding of TC properties should not be limited to only these two variables. The TC size is also an important metric, and it is actually associated with the changes of TC intensity and track (Chavas et al., 2016; Sun et al., 2017), and with the structure of TC wind field (Chan and Chan, 2013, 2014, 2015; Chavas et al., 2015; Wang et al., 2015; Chavas and Lin, 2016). The definition of TC size is different in many previous studies. For example, some used the azimuthally averaged radius of 17 m s\(^{-1}\) (\(R_{17}\)) or 15 m s\(^{-1}\) (\(R_{15}\)) of lower-tropospheric winds (Cocks and Gray, 2002; He et al., 2020). However, due to the lack of an objective, observation-based TC wind field, TC size has not been as widely studied as TC intensity and track (Chavas and Emanuel, 2010; Chan and Chan, 2012). So far, only the U.S. Department of Defense Joint Typhoon Warning Center (JTWC) and the Japan Meteorological Agency (JMA) provide some TC size information in their best-track datasets for the West Pacific typhoons, which has

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been used to investigate the TC wind structure (Lei and Chen, 2005; Lu et al., 2011). Recent works have begun to enhance our understanding of TC size, given more and more in situ and remote observations as well as improved numerical models of TCs (Wang, 2009; Xu and Wang, 2010; Chan and Chan, 2012; Chavas et al., 2015; Chavas and Lin, 2016; Chavas et al., 2016; Wang and Toumi, 2018a, 2019; Zhao et al., 2019; Sheng et al., 2020).

As an important indicator of ocean surface energy status, the sea surface temperature (SST) plays an important role in the formation and development of TCs (Webster et al., 2005; Pierce et al., 2006; Emanuel, 2007; Huang et al., 2015). Observational data and numerical model results have confirmed the sensitivity of TC characteristics (e.g., track, intensity, size, and destructiveness) to SST in the vicinity of the TC (Emanuel, 2005; Lin et al., 2014; Sun et al., 2014, 2017; Lau et al., 2016; Wang and Toumi, 2018b). As suggested by Sun et al. (2013, 2014), the simulated TC intensity and size are sensitive to SST changes over different radial extents; namely, both the TC intensity and size are more sensitive to SST changes within a radius of 300 km from the TC center, compared to the SST changes outside the 300-km radius from the center. It is reasonable to expect that the TC-induced SST cooling, which is mainly concentrated within a radius of 300 km from the TC center, plays an important role in determining the TC intensity and size. Specifically, the TC-induced SST cooling occurs when the Ekman pumping generated by the strong wind stress curl associated with TCs near the ocean surface, the heat exchange at the air–sea interface, and the mixing of the upper ocean all act together to lower the SSTs significantly after the passing-by TCs. The SST cooling observed at the TC wake is most commonly known as the cold wake (Price, 1981; Jacob et al., 2000; D’Asaro et al., 2007). The TC-induced SST cooling contributes to the reduction of TC–ocean heat flux, which is expected to be one of the several factors that determine the TC intensity and size (Emanuel et al., 2004; Wu et al., 2005).

Previous studies have employed high-resolution ocean–atmosphere coupled models to study TCs, with conflicting results regarding the effect of the cold wake on TC development (e.g., Schade and Emanuel, 1999; Bender and Ginis, 2000; Bender et al., 2007; Lee and Chen, 2014). A coupled hurricane–ocean model was constructed to show that the cold wake effect can reduce the TC intensity by more than 50% (Schade and Emanuel 1999). However, Lee and Chen (2014) hold the opposite view and found, by using a coupled atmosphere–ocean model, that the cold wake can enhance the TC by increasing the atmospheric stability inside the TC and regulating the inflow of moist air. It is recognized that the underlying physical mechanisms responsible for the SST controlling TC intensity and size changes are not well understood (Sun et al., 2017; Bruneau et al., 2020).

Based on the above review and discussion, this paper intends to specifically examine the effect of the cold wake on TC size change and associated TC destructiveness. We take TC Kaemi (2006) as an example and use an atmosphere–ocean coupled model to investigate the physical mechanisms involved in the SST cooling impact on the TC size. The results are analyzed in comparison with a model simulation using an uncoupled atmospheric model only. The models and experimental design are described in Section 2. Sections 3 and 4 present the results from model simulations and illustrate possible influencing factors on the size and destructiveness changes of TC Kaemi, respectively. Finally, conclusions and discussion are given in Section 5.

2. Models and experimental design

To access the impact of SST cooling on the TC size, a case study of TC Kaemi is conducted here. TC Kaemi originated from the Northwest Pacific. Kaemi formed as a tropical depression on 18 July 2016, landing in Taiwan Island on 24 July, with the maximum wind force of 40 m s$^{-1}$ at the time of landing; it then weakened into an extratropical cyclone and dissipated on 26 July over southern China. It caused winds of 40 m s$^{-1}$ or higher over most parts of Taiwan Island.

The coupled model used in this study is based on the Weather Research and Forecasting (WRF) model (Powers et al., 2017) and Princeton Ocean Model (POM; Blumberg and Mellor, 1987). The WRF model is a fully compressible nonhydrostatic model; and the governing equation is realized in the flux form, which is integrated by using the Arakawa C grid. This model contains many physical options. In our experiments, the horizontal grid points of WRF model are 143 × 142 in the outermost simulation domain. The grid spacing is 15 km, and in the vertical there are 36 sigma layers. The inner nested moving grid spacing is 5 km. The Kain–Fritsch cumulus parameterization scheme (Kain and Fritsch, 1990; Kain, 2004), Monin–Obukhov surface-layer parameterization scheme (Webb, 1970; Beljaars, 1995), WRF single-moment (WSM) 6-class graupel microphysics scheme (Hong and Lim, 2006), Rapid Radiative Transfer Model longwave radiation scheme, Dudhia shortwave radiation scheme, and Yonsei University (YSU) planetary bound-
From Ma et al. (2009) were taken as the initial and open temperature and salinity, and climatological ocean data taken as the initial and open boundary conditions for the Simple Ocean Data Assimilation (SODA) data. To carry out a diagnosis run, the monthly averaged phase, two more steps need to be taken. The first step is suitable marine initial conditions before the coupling locations of ocean currents and eddies. In order to obtain salinity profile of the upper ocean as well as for accurate del is the model initialization for accurate temperature–ant factor to ensure good performance of an ocean mo-

The simulation domain is larger than that of the WRF, and the southern parts of Taiwan Island and Philippine Islands in the west of the simulation area are treated as a whole continent. The southern parts of Taiwan Island and Philippine Islands in the west of the model are taken as fixed boundary conditions, and the normal velocity perpendicular to the coast is set to zero. However, the bogus TC scheme will lead to an imbalance between the bogus vortex and large-scale forcing. The simulation of TC’s intensity and structure will produce a large deviation from the observation during the model spin-up. However, this imbalance will gradually decrease over time, and after a period (spin-up time), the bogus vortex and large-scale forcing will balance better. Therefore, the analysis time period we choose is after the spin-up time.

POM uses a second-order turbulence closure scheme to calculate the vertical mixing coefficient. It uses σ coordinates in the vertical, and orthogonally curvilinear grid in the horizontal to better match the lateral boundaries. It also uses the Arakawa C grid. The horizontal and temporal difference schemes are the explicit ones, and the vertical scheme is the implicit difference format. POM has both the internal and external modes, which are separately calculated. The external and internal modes are two-dimensional and three-dimensional, respectively. In our experiments, the number of regional POM grid points is 118 × 113, and the grid spacing is 1/6° × 1/6°. The simulation domain is larger than that of the WRF, with 16 sigma layers in the vertical. The maximum ocean depth is set to 3000 m, and the Philippine Islands are treated as a whole continent. The southern parts of Taiwan Island and Philippine Islands in the west of the model are taken as fixed boundary conditions, and the normal velocity perpendicular to the coast is set to zero.

Halliwell et al. (2011) suggested that the most important factor to ensure good performance of an ocean model is the model initialization for accurate temperature–salinity profile of the upper ocean as well as for accurate locations of ocean currents and eddies. In order to obtain suitable marine initial conditions before the coupling phase, two more steps need to be taken. The first step is to carry out a diagnosis run. The monthly averaged Simple Ocean Data Assimilation (SODA) data were taken as the initial and open boundary conditions for the temperature and salinity, and climatological ocean data from Ma et al. (2009) were taken as the initial and open boundary conditions for the sea level and velocity in POM. POM was run for one year, which provided the initial field suitable for the POM model in January 2006. The second step is to carry out a prediction run. Based on the initial field obtained from the first step, the climatological results of the regional ocean model (as mentioned above) were taken as the open boundary conditions of the sea level and velocity, and the monthly averaged SODA data were used as the open boundary condition of ther-mohaline. The upper boundary forcing field used the Quick Scatterometer (QuikSCAT) data. The upper boundary heat forcing assimilated the daily Advanced Mechanically Scanned Radiometer (AMSR) SST data. POM was run until 0000 UTC 20 July 2006 when TC Kaemi occurred, which provided the initial conditions for POM in the coupled experiment.

Both the uncoupled and coupled experiments were conducted in this study based on the above-mentioned configuration in WRF and POM. For convenience, the uncoupled experiment is referred to as CTRL. The coupled experiment is represented by COUP. We use suitable marine initial conditions obtained in the second step as the initial conditions of POM, and the FNL data combined with the bogus TC scheme as the initial conditions of the WRF model in COUP. The coupling phase of the model is 20–23 July, during which the WRF model drives POM through the atmospheric forcing [i.e., sensible and latent heat fluxes (SH and LH), and sea-surface wind stress at 10 m]. SST is the output of POM, which is used to estimate air–sea heat fluxes for the WRF model. The two models communicate with each other once an hour. The configuration of the WRF model is the same for both CTRL and COUP. The SST field used as the ocean forcing in CTRL is time-invariant, and is equal to the initial SST field in COUP.

3. Simulated results

3.1 Power dissipation index (PDI) and size-dependent TC destructive potential

The eye location of an TC is determined by the minimum wind speed near the minimum sea-level pressure. Figure 1 compares the TC tracks simulated in the two experiments with the best track with the 6-h interval from the Regional Specialized Meteorological Center (RSMC) Tokyo-Typhoon Center operated by the JMA (available at http://www.jma.go.jp/jma/jma-eng/jma-center/rsmc-hp-pub-eg/besttrack.html). Since the initial field of the mod-
PDI = \int_0^\tau V_{\text{max}}^3 \, dt, \quad (1)

where \( V_{\text{max}} \) is the MWS at 10 m, and \( \tau \) is the lifetime of TC.

Despite the wide use of PDI, it is not a perfect measurement of the TC destructive potential due to the lack of information on the TC size. In this study, we use the index that reflects the destructive potential over its lifetime; namely, the size-dependent TC destructive potential (PDS; Sun et al., 2017). The formula for calculating PDS is presented here,

\[
PDS = \int_0^{A_{17}} \int_0^\tau C_d \rho |V|^3 \, dA \, dt, \quad (2)
\]

where \( C_d \) is the surface drag coefficient, \( \rho \) is the average sea-level air density, \(|V|\) is the magnitude of the surface wind at 10 m, and \( A_{17} \) is the area of gale-force surface winds.

To compare the difference between the two indices in the two experiments, PDI and PDS calculated by Eqs. (1) and (2) are given in Table 1. PDI of CTRL is about 7.4\% (9.3 vs. 8.6; 10^{18} \text{ kg m}^{-2} \text{ s}^{-2}) more than that of COUP, which shows that even if considering the effect of SST cooling, PDI is essentially unchanged. However, there is a significant reduction from CTRL to COUP in terms of PDS, and the reduction ratio is more than 1/3 (2.1 vs 1.4; 10^{18} \text{ kg m}^{-2} \text{ s}^{-2}).

To illustrate the significant difference in PDS between CTRL and COUP, the evolution of MWS, radius of maximum wind (RMW), and \( A_{17} \) in CTRL and COUP are shown in Fig. 2. Because the eyewall of TC can be asymmetric (Wang, 2007), we replace the location of overall eyewall with the azimuthal mean RMW. Throughout the simulation, MWS and RWM in CTRL show similar evolution to those in COUP. Changes of MWS go through two stages: from the start of simulation, MWS reached its first maximum of 33 m s\(^{-1}\) in about 11 h, and then the value reduced until the second enhancement stage began in 30 h. After a rapid intensification, the TC strengthened further and obtained its peak MWS of 38 m s\(^{-1}\) around 22 July, and then it began to weaken. The values of MWS show no significant difference between CTRL and COUP, which indicates that PDI did not change much during the entire simulation. The same is true for RMW; changes in the whole first stage are basically the same for both CTRL and COUP. In the second enhancement stage of TC, RMW reached a minimum of 40 km. This indicates that the TC was accompanied by the contraction of eyewall when it was enhanced. Nonetheless, the most striking feature is the change of \( A_{17} \). In the early stage of the simulation, changes of \( A_{17} \) from 20 to 21 July are basically the same in the two experiments; but from 21 July, that is, from the second enhancement stage of TC, \( A_{17} \)
showed a distinct trend in CTRL from that in COUP. $A_{17}$ in CTRL still continuously increased, but in COUP it began to decrease and reached a minimum near the moment when MWS reached its maximum value. At that time, the values of $A_{17}$ were about $2.1 \times 10^5$ and $4.4 \times 10^5$ km$^2$ in CTRL and COUP, respectively. It is noted that the wind disaster range in COUP was half of that in CTRL when TC was the strongest.

During the entire simulation, there was no notable difference in PDI but a large difference in PDS between the two experiments. Therefore, without considering the effect of the SST cooling on the TC size, the TC destructive potential could be largely overestimated.

The observed TC size in the JMA dataset was used for verification. Due to the lack of $R_{17}$, $R_{15}$ was used as a measurement of the TC size. We compared the values of $R_{15}$ from CTRL, COUP, and JMA. Figure 3 shows the temporal evolution of $R_{15}$ values during the simulation period. The results clearly show that the TC size obtained from the WRF–POM coupling experiment was reduced, compared to the uncoupled WRF model result, and was closer to the JMA observation.

3.2 Changes of the wind speed and SST

As mentioned above, the most significant difference between the two experiments is the change of $A_{17}$, which leads to the difference in the destructive potential change. To further investigate possible causes for the difference in the change of $A_{17}$ between the two experiments, we use Fig. 4a to show the azimuthal-mean wind speed distribution of COUP minus CTRL at 10 m throughout the simulation. At the beginning of the simulation, the azimuthal-mean wind speed change was small and concentrated near the maximum wind speed radius. It is obvious that the maximum wind speed difference occurred near the center of TC at 100–200 km on 22 July and reached a maximum of 5 m s$^{-1}$ (Fig. 4a).

Figure 4a also shows that PDS determined by $A_{17}$ has a significant difference in the mature TC stage for the two experiments. We focus on the time frame corresponding to the mature TC stage. In this study, the mature TC stage is defined as the period when the maximum wind speed at 10 m is close to its lifetime maximum wind speed at 10 m, and the difference between the two values is $\leq 3$ m s$^{-1}$. Thereby, the simulated TC mature stage is approximately from 1500 UTC 21 July to 0600 UTC 22 July 2006. A brief overview of the azimuthal-
mean wind speed–radius evolution within the TC mature stage is presented in Fig. 5. We can see that the wind speed of the mature TC in the two experiments increased first and then decreased with the radius, while the RMW values of the two were almost the same. A significant weakening of the wind speed in COUP occurred outside the eyewall, especially in the range of 100–200 km from the TC center. The wind speed was about 24 m s$^{-1}$ in CTRL, which dropped to 18 m s$^{-1}$ in COUP.

The size of TC is defined here as $R_{17}$. Consistent with Fig. 2c, there is a large difference in $R_{17}$ between the two experiments in the TC mature stage, as $R_{17}$ values of CTRL and COUP are approximately in the ranges of 320–340 and 240–260 km, respectively. Thereby, it is the decrease of wind speed outside the eyewall that leads to the reduction of TC wind disaster range.

To find out why SST cooling has such a large impact on the TC gale-force surface wind speed in the mature stage, we show the SST horizontal distribution after the spin-up time every 12 h from 1200 UTC 20 July to 0000 UTC 23 July 2006 in Fig. 6. In the initial stage after the TC adjustment, an obvious cold zone, which was not caused by TC Kaemi, appeared near the east side of Taiwan Island (Fig. 6a). By analyzing the Remote Sensing Systems data (available at ftp://ftp.remss.com/sst/daily/mw/v05.0/netcdf/2006), we found that TC Bilis (2006) just passed the significant SST cooling area one week ago and produced a strong cooling; and the SST off eastern Taiwan Island dropped to 24°C before TC
Kaemi’s arrival. TC Kaemi formed around 18 July and entered our simulation area around 20 July. As the TC moved northwestward, the model-simulated SST showed a significant decrease and reached 25–26°C near the TC center on 22 July, consistent with the observation reported in Subrahmanyan (2015). This indicates that the WRF–POM coupled model can reproduce the TC-induced SST cooling.

In addition, the cold pool off eastern Taiwan Island became cooler in our simulation (Fig. 6). In general, the SST change beneath TC is sensitive to the surface ocean mixed layer depth (MLD; D’Asaro et al., 2007). Thus, we use the climatology monthly MLD data from French Research Institute for Exploration of the Sea (IFREMER; available at http://www.ifremer.fr/cerweb/deboyer/mld/Surface_Mixed_Layer_Depth.php) for a further analysis. It is found that the MLD off eastern Taiwan Island is shallow in July, generally between 10 and 20 m (figure omitted). When a TC (e.g., Bilis) passed through our simulation domain, strong winds caused large SST cooling in the shallow MLD areas. When TC Kaemi reached its mature stage, the cold pool in the upper left corner of the study domain was still far from the center of Kaemi (>300 km). In Sun et al. (2014), the region outside the 300-km radius was defined as the remote region, and it is noted that the SST change in the remote region has a relatively small effect on the size and intensity of TC Kaemi.

Similar to Fig. 4a, Fig. 4b shows the azimuthal-mean SST distribution of COUP minus CTRL throughout the simulation period. Figures 4a and b together reveal a relationship between the SST cooling and wind speed reduction. In the early stage of the simulation, SST did not change much (about 0.5°C) from 20 to 21 July. However, a significant cooling (more than 3°C) happened at 0600 UTC 22 July. Moreover, the maximum SST cooling occurred mainly near the center of the TC, and the cooling decreased away from the TC center. Next, we will try to explore the mechanisms involved with the effect of the TC-induced SST cooling on the wind speed changes outside the TC eyewall.

4. Physical mechanisms

4.1 Sea surface entropy

Previous studies have shown that the change in SST can affect the TC activity by changing the distribution of surface enthalpy flux (SEF), such as the TC intensity and size (Emanuel, 1986; Rotunno and Emanuel, 1987; Xu and Wang, 2010). SEF is the sum of LH and SH. LH is determined by the sea-surface wind speed and air–sea moisture difference (ASMD), where ASMD is defined as the sea-surface saturation specific humidity minus air specific humidity at 2 m. SH is determined by surface wind speed and air–sea temperature difference, where the air–sea temperature difference is defined as the SST minus air temperature at 2 m. LH and SH are calculated by Eqs. (3) and (4):

\[ Q_{LH} = \rho L_e C_v U (q_s - q_a), \]  
\[ Q_{SH} = \rho C_p C_h U (T_s - T_a), \]

where \( \rho \) is the density; \( L_e \) is the evaporation latent heat; \( C_p \) is the specific heat at constant pressure; \( C_v \) and \( C_h \) are turbulent exchange coefficients of the latent heat and sensible heat, respectively; \( U \) is the wind speed at 10 m; \( q_s \) and \( q_a \) are the sea-surface saturation specific humidity and air specific humidity at 2 m, respectively; \( T_s \) and \( T_a \) are the SST and air temperature at 2 m, respectively; and \( q_s \) is only a function of \( T_s \).

Generally speaking, evolutions and distributions of LH and SH are similar. According to the formulas above, SST determines \( q_s \), which determines ASMD. Therefore, ASMD is the key to understand the SEF response to SST changes. Although the underlying SST determines the TC size and intensity by inputting SEF into the TC, the change in SST may be inconsistent with the change in SEF caused by the SST change.

Figure 7 is the Hovmöller diagram of the azimuthal-mean ASMD and LH in CTRL, COUP, and their differences in the range of 0–300 km from the TC center. From Figs. 7a, b, the maximum LH occurred near the eyewall due to the largest wind speed at the eyewall. Before 21 July, both LH and ASMD distributions were similar in the two experiments. However, the situation became different after 21 July when LH and ASMD distributions were different outside the eyewall. In CTRL, LH showed an outward expansion outside the eyewall, while the radial distribution of ASMD showed the relatively small change. In COUP, the low-value LH contracted from the outer region to the eyewall, and the low-value ASMD expanded from the inner region to the outer. They are consistent with the results of Sun et al. (2013), as the low-level inflow approached to the eyewall, the surface moisture was gradually lost due to the negative effect of the decreased SST in the outer region of TC. As a result, ASMD was relatively smaller in the outer region of TC and relatively larger in the inner region of TC (Fig. 7c).

4.2 Radial and tangential wind speeds

As suggested by recent studies, changes in the radial distribution of SEF greatly affect the variation of tangen-
The azimuthal wind, which affects the size of TC (Miyamoto and Takemi, 2010; Xu and Wang, 2010; Sun et al., 2013, 2014). To investigate how surface winds outside the eyewall (and thus the storm size) respond to changes in the underlying SEF, which is associated with SST, we discuss the TC momentum budget equations for the azimuthal-mean radial and tangential winds proposed by Xu and Wang (2010), which are given as follows:

\[
\frac{\partial \bar{u}}{\partial t} = -\frac{1}{\bar{\rho}} \frac{\partial \bar{\rho}}{\partial r} + \frac{\bar{V}^2}{r} + f \bar{V} + \bar{F}_u + \bar{D}_u, \tag{5}
\]

\[
\frac{\partial \bar{V}}{\partial t} = -\bar{u}\bar{\zeta}_u - \bar{w} \frac{\partial \bar{V}}{\partial z} + \bar{F}_V - \bar{u}'\bar{\zeta}' - \bar{w}' \frac{\partial \bar{V}'}{\partial z} + \bar{D}_V, \tag{6}
\]

where \(\bar{u}, \bar{V},\) and \(\bar{w}\) are the azimuthal-mean radial and tangential wind speeds and vertical velocity, respectively; \(\bar{\zeta}_u\) is the azimuthal-mean vertical absolute vorticity; \(r\) is the distance to TC center; \(z\) is the height; \(\bar{\rho}\) and \(\bar{\rho}\) are the density and pressure, respectively; \(\bar{F}_u, \bar{F}_V, \bar{D}_u\) and \(\bar{D}_V\) are the parameterized sub-grid vertical diffusion, including the surface friction of radial and tangential winds, and horizontal diffusion of radial and tangential winds, respectively; \(u', V', w',\) and \(\zeta'\) are the radial, tangential and vertical winds, as well as vertical relative vorticity deviation, respectively, from their corresponding azimuthally-averaged values. On the basis of Eqs. (5) and (6), Xu and Wang (2010) suggested that \(\bar{u}\) (i.e., low-level inflow) is largely affected by the pressure gradient term \(\frac{\partial \bar{\rho}}{\partial r}\), which further impacts the distribution of tangential wind (\(\bar{V}\)) through the absolute vorticity radial advection (\(-\bar{u}\bar{\zeta}_u\)).

Figure 8 shows the Hovmöller diagram of the azi-
muthal-mean SEF and sea-level pressure gradient in CTRL and COUP as well as their differences in the range of 0–300 km from the TC center. From Figs. 8a, b, we can see that the radial pressure gradient near the eyewall reached its maximum due to the wind speed near the eyewall being the largest. Because LH was about one order of magnitude greater than SH. The distribution of SEF was basically the same as that of LH. For the two experiments, the pressure gradient did not change significantly near the eyewall; but outside TC, the low-pressure gradient contracted from the outer region to the eyewall. For the simulated TC in COUP, compared with that in CTRL, both the SEF and radial pressure gradient outside the eyewall decreased significantly in the mature stage (Fig. 8c). The change in pressure gradient was related to the hydrostatic adjustment of adiabatic heating in the convection, which was contributed by changes in SEF, namely, the suppression of SEF outside the eyewall could lead to reduction of the convection. This is consistent with the results obtained by Bister (2001), which suggested that the TC convection will be suppressed as SEF decreases. The reduction in convection will further result in a reduction in TC radial pressure gradient outside the eyewall. This is the mechanism that explains how the relatively larger reduction of SEF contributes to the reduction of sea-level pressure gradient outside the eyewall (Fig. 8c).

The first item on the right-hand side in Eq. (5) (i.e., \( \frac{1}{\rho} \frac{\partial \rho}{\partial r} \)) will determine the change of radial wind (i.e., \( \frac{du}{dt} \)). To verify the rationality of the formula, Fig. 9 displays the Hovmöller diagram of the azimuthal-mean radial wind speed (\( \bar{u} \)) and absolute vorticity radial advection (\( -\vec{u} \zeta_u \)) in CTRL and COUP as well as their differences in

\[ \begin{align*}
\text{Volume} & \quad 34
\end{align*} \]

![Fig. 9. As in Fig. 7, but for the azimuthal-mean radial wind speed (m s\(^{-1}\); shading; outward is positive) and absolute vorticity radial advection (10\(^{-2}\) m s\(^{-2}\); black contours).](image-url)

4.3 Systematic examination

We carried out a systematic examination on the relationship between the SST drop and TC size change based on our model results. Since the model was run for 3 days (72 h) in total, we removed the first 12 h as this was the spin-up time. We correlated the time series of SST drop with that of the TC size change by using the remaining
60-h data. According to Sun et al. (2013, 2014), the SST beyond 300 km from the TC center has little effect on the TC size. Therefore, we hereby define the SST drop as the average SST within 300 km from the TC center in COUP minus that in CTRL. The TC size change is defined as \( R_{17} \) in COUP minus that in CTRL. Figure 10 shows temporal evolution of the TC size change and SST drop after the spin-up time. It can be seen from Fig. 10 that the SST drop and TC size change have similar trends. Their correlation coefficient is 0.76, significant at the 99.99% confidence level. It is thus concluded that the TC size change is strongly associated with the SST drop. This result can partly explain the effect of cold wake on the TC size. It is consistent with the results of some previous studies based on numerical models, indicating that the TC size would increase (decrease) under the SST warming (cooling; Xu and Wang, 2010; Sun et al., 2013, 2014, 2017).

In this study, a statistical analysis based on observational data is also made to validate the simulation results on the relationship between the SST and TC size. Because it is hard to separate the SST-cooling effect from the SST effect on the observed change of the TC size based on observations, we focus on the relationship between the SST and TC size, rather than that between the SST cooling and TC size change. We select the best-track data at 6-h intervals from the RSMC Tokyo-Typhoon Center operated by the JMA (available at http://www.jma.go.jp/jma/jma-eng/jma-center/rsmc-hp-pub-eg/besttrack.html), which have the longest data period with TC size information (from 1977 to 2018) over the western North Pacific (WNP), and the NOAA Extended Reconstruced SST version 5 (ERSSTv5; available at https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.ersst.v5.html). Due to the lack of radius of 17 m \( s^{-1} \) wind \( (R_{17}) \) in JMA, the radius of 15 m \( s^{-1} \) wind \( (R_{15}) \) is used as a measure of the TC size. The statistical analysis includes 1019 TCs from 1977 to 2018. The TC-related SST is defined as the average SST within a radius of 1000 km from the TC center in the life cycle of TC (i.e., \( R_{\text{lifetime}} \)). The corresponding TC size is defined as the average \( R_{15} \) in the life cycle of TC (i.e., \( R_{\text{lifetime}} \)). The correlation coefficient of SST\(_{\text{lifetime}}\) and \( R_{\text{lifetime}} \) is −0.051. It seems that there is no significant relationship between the TC size and SST. This may be an important reason why there are few studies on the relationship between the SST and TC size.

Nevertheless, the situation become substantially different when we divide the life cycle of TC into two stages, i.e., the intensifying stage before TC reaches its lifetime maximum intensity (LMI) and the decaying stage after TC reaches its LMI. SST of the two periods (i.e., stages before and after TC reaches its LMI) are defined as SST\(_{\text{pre-LMI}}\) and SST\(_{\text{post-LMI}}\), respectively. The TC sizes of the two stages are defined as \( R_{\text{pre-LMI}} \) and \( R_{\text{post-LMI}} \), respectively. On the one hand, the correlation coefficient of SST\(_{\text{pre-LMI}}\) and \( R_{\text{pre-LMI}} \) is 0.197, which is significant above the 99% confidence level. It can be concluded that the TC size is closely associated with SST before TC reaches its LMI. On the other hand, unexpectedly, the correlation coefficient of SST\(_{\text{post-LMI}}\) and \( R_{\text{post-LMI}} \) is −0.125, which is also significant above the 99% confidence level. Generally speaking, more than 45% of all recurving storms have coincident recurvature and LMI over WNP (Evans and McKinley, 1998). Namely, many WNP TCs turn northward after reaching its LMI. Note that TCs with larger size \( (R_{15}) \) tend to move northward farther as larger TCs are often of higher intensity and longer duration. As a result, the larger TCs are inclined to move to the higher latitudes with lower SST. This explains why SST\(_{\text{post-LMI}}\) and \( R_{\text{post-LMI}} \) have a significant negative relationship. Overall, without the TC’s northward moving, the SST warming (cooling) contributes significantly to the increase (decrease) of TC size. The issue on the observed relationship between the SST and TC size will be further investigated in future as it is more complicated and somewhat out of the scope of the present study.

5. Conclusions and discussion

To better understand the feedback effect of TC-induced SST cooling on the TC size, a WRF–POM coupled model is used to investigate the associated possible mechanisms in comparison with an uncoupled WRF
model simulation. Previous studies have elaborated on the physical processes involved in SST changes over different radial extents on the TC intensity but have seldom paid attention to the effect of TC-induced SST cooling on the TC size. Our results indicate that due to the TC-induced SST cooling in the TC mature stage, the wind speed profile changes substantially outside the eyewall (100–200 km from the TC center), resulting in a significant decrease in the TC size, which eventually leads to less destructiveness (e.g., PDS).

Our results further indicate that TC could induce a significant SST cooling (1–3°C drop) within a radius of 0–300 km from the TC center; and the closer to the TC center, the greater the SST cooling. Previous studies have shown that SEF (i.e., LH + SH), which is largely determined by SST, plays an important role in the TC intensity and size. The physical mechanism of SST cooling affecting the TC size is summarized in Fig. 11. As the low-level air approaches the TC eyewall, surface moisture is gradually lost because a significant SST cooling happens during the TC mature stage. As a result, ASMD becomes smaller outside the eyewall, which leads to a smaller SEF outside the eyewall. The adiabatic heating associated with the reduced SEF outside the eyewall has two effects. On the one hand, it leads to a smaller absolute vorticity in the outer region. On the other hand, it causes a weakening of the outer pressure gradient, resulting in a weakening of the low-level inflow outside the eyewall. According to the TC momentum budget, both these effects contribute to the weakening of tangential wind by $\bar{u}_a$ outside the eyewall. The weakening of tangential wind in turn leads to further weakening of SEF in the outer region. The feedback between the SEF and tangential wind results in further weakening of the tangential wind. Weakening of the tangential wind, combined with that of the low-level inflow, reduces the wind speed profile outside the eyewall. The TC size (i.e., $R_{17}$) and thus TC size-dependent destructiveness (i.e., PDS) all decrease. If the feedback mechanism of SST to TC is not considered, the size and destructiveness of TC will be seriously overestimated.

Generally speaking, the cold wake may weaken the TC intensity by inhibiting the development of TC itself via reducing the energy supply (e.g., Schade and Emanuel, 1999; Chen et al., 2017). However, Lee and Chen (2014) suggested that the cold wake of the storm and stable boundary layer in the right-rear quadrant of the TC center can offset part of the negative impact on the storm intensity by increasing the atmospheric stability inside TC and regulating the moist air inflow. We conclude that the TC-induced cold wake contributes little to the TC intensity in the mature stage, based on our experiments. Maintenance of the TC intensity between CTRL and COUP can be explained by Eq. (6) and Fig. 9c. Despite of the decreased SST, the increased inflow can lead to an increase of the tangential wind speed when $-\bar{u}_a > 0$ near the eyewall in the TC mature stage. The increase of tangential wind speed will offset part of the negative impact on TC by the cold wake.

A significant increase in the TC size can be seen in

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**Fig. 11.** The schematic diagram summarizing possible mechanisms responsible for the impact of TC-induced SST cooling on the TC size.
Fig. 3, because of the use of the air–sea coupled model. However, gaps still exist between observations and COUP outputs. The TC not only caused SST cooling but also strong winds, which produced large sea-surface waves affecting the air–sea momentum, heat, and moisture fluxes. It is foreseeable that if a wave model is added to the air–sea coupled model, the prediction of TC size and destructiveness will be more accurate. In addition, Bruneau et al. (2018) showed that the TC-induced cold wake can be more evident in the presence of continental shelf. This means that the strong decreases of the TC size and destructiveness, as found in the present study, may be more observable when a TC approaches land, due to the stronger SST cooling (cold wake) effect. These issues will be studied in future.

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