Impact of Sea Spray and Sea Surface Roughness on the Upper Ocean Response to Super Typhoon Haitang (2005)

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ABSTRACT: A coupled ocean–wave–sea spray model system is used to investigate the impacts of sea spray and sea surface roughness on the response of the upper ocean to the passage of the Super Typhoon Haitang. Sea spray–mediated heat and momentum fluxes are derived from an improved version of Fairall’s heat fluxes formulation and Andreas’s sea spray–mediated momentum flux models. For winds ranging from low to extremely high speeds, a new parameterization scheme for the sea surface roughness is developed, in which the effects of wave state and sea spray are introduced. In this formulation, the drag coefficient has minimal values over the right quadrant of the typhoon track, along which the typhoon-generated waves are longer, smoother, and older, compared to other quadrants. Using traditional interfacial air–sea turbulent (sensible, latent, and momentum) fluxes, the sea surface cooling response to Typhoon Haitang is overestimated by 1°C, which can be compensated by the effects of sea spray and ocean waves on the right side of the storm. Inclusion of sea spray–mediated turbulent fluxes and sea surface roughness, modulated by ocean waves, gives enhanced cooling along the left edges of the cooling area by 0.2°C, consistent with the upper ocean temperature observations.

KEYWORDS: Ocean dynamics; Atmosphere–ocean interaction; Hurricanes/typhoons; Oceanic mixed layer

1. Introduction

Typhoons are special atmosphere and ocean coupled systems with relatively complicated physical phenomena involving air–sea interactions. The upper ocean is rather sensitive to air–sea turbulent fluxes (sensible, latent, and momentum) affected by waves and sea spray droplets, associated with breaking waves, as induced by typhoons (Zhang et al. 2006; Zhang and Perrie 2008). Since direct measurements of air–sea turbulent exchanges are extremely challenging during the passage of a typhoon, estimates of the interfacial turbulent fluxes are usually constructed by using the Coupled Ocean–Atmosphere Response Experiment (COARE) bulk flux algorithm, in which the total effect of sea spray droplets and ocean waves are relatively absent (Andreas et al. 2008).

Sea spray droplets are generated from wave breaking induced by strong winds generated in severe weather systems. Once spray droplets are ejected into the air–sea interface, the balance of moisture and sensible heat in the boundary layer is changed by the evaporation and heat transfer due to the spray droplets (Riehl 1954). Sea spray observations collected in wind-wave tank (Wu 1974) measurements indicated that the

magnitude of the sea spray evaporation can contribute 13% to the total evaporation, in winds of 13.4 m s^{-1} . For normal-to-extreme high wind speeds, the contribution ratio of spray to total evaporation is able to reach 18%, to even 23%, in laboratory measurements (Komori et al. 2018; Jeong et al. 2012). Based on studies of sea spray generation (Zhang and Lou 1995; Andreas 1989, 1990) and evaporation (Hasse 1992; Andreas and Mahrt 2016), similar bulk heat fluxes parameter schemes have been developed to estimate the sensible and latent heat exchanges, mediated by sea spray (Andreas 1992; Andreas et al. 2015; Fairall et al. 1994). On the basis of a new sea spray generation function, including the effects of wind speed and whitecaps, Fairall et al. (1994) proposed a relatively reliable model for the calculation of sea spray heat fluxes in high wind conditions. By modifying the scale height of the sea spray evaporation zone and the whitecap areal fraction of Fairall’s sea spray algorithm, Zhang et al. (2019) considered the feedback mechanisms between sea spray and the air–sea interface. Peng and Richter (2019) also suggested that feedback mechanisms induced by spray evaporation should be considered in a spray-mediated flux model to modulate the corresponding interfacial fluxes.

When spray droplets are ejected into the air–sea interface at the height of the breaking wave crest, they are almost immediately accelerated by the local wind. By comparison, when they crash back to the sea surface, they also contribute to the momentum transfer from the atmosphere to the ocean (Edson 1990; Andreas and Emanuel 2001), significantly

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shifting the near-surface wind speed distribution (Kudryavtsev and Makin 2011). Thus, sea spray droplets affect the boundary layer dynamics via the direct impact of the droplets on the air–sea interfacial momentum exchange to form the so-called spray stress. Once the wind speed reaches about 30 m s^{-1} , spray stress provides about 10% of the total stress (Andreas 2004). Andreas (2004) also estimated that spray stress is able to carry all of the surface stress when the wind speed reaches about 60 m s^{-1} . In the latter wind speed conditions, Innocentini and Goncalves (2010) concluded that the sea spray-mediated stress is about 20% of the air–sea stress. When U_{10} reaches 50 m s^{-1} , Fairall et al. (1994) estimated that the sea spray carries approximately 10% of the interfacial stress. Mueller and Veron (2014) pointed that the value of spume stress is independent of the wind speed, providing about 10% of the momentum fluxes. Thus, assuming that the concentration of sea spray droplets is sufficiently high, spray droplets can carry a substantial amount of the total air–sea stress (Richter and Sullivan 2013).

To a certain extent, the air–sea exchange coefficient (drag coefficient C_d) or sea surface roughness, represents the ocean and atmosphere interfacial momentum exchange. Previous field studies (Johnson et al. 1998) and numerous measurements (Large and Pond 1981; Taylor and Yelland 2001) suggest that C_d increases with U_{10} at “normal” nontyphoon extremal wind conditions ($U_{10} \leq 30 \text{ m s}^{-1}$). Generally, an increase in C_d is associated with a *broadening* of the surface wave spectrum with increasing wind speed (Makin et al. 1995; Janssen 1989). When wind speeds continue to increase ($U_{10} > 30\text{--}35 \text{ m s}^{-1}$), C_d appears to level off (Takagaki et al. 2016) and eventually decrease (Jarosz et al. 2007; Zijlema et al. 2012). However, the variation of drag coefficient has significant scatter, if parameterized as only dependent on wind speed (Zhang et al. 2017, hereinafter Z17), suggesting that other physical processes (ocean waves, wave breaking, and sea spray) may possibly play important roles in determining the drag coefficient or sea surface roughness. At wind speeds exceeding 33 m s^{-1} , a thin part of the surface layer adjacent to the surface becomes a regime of limited saturation with the spume droplets (Kudryavtsev and Makin 2011), which is responsible for a decrease of the drag coefficient (Powell et al. 2003; Makin 2005); this further dampens the turbulent mixing and enhances energy dissipation (Bao et al. 2011). In addition, the wave breaking associated whitecapping also induces a rapid increase in the heat exchange coefficient for increasingly extreme high wind speeds, as observed through laboratory experiments (Troitskaya et al. 2020). Soloviev et al. (2014) also investigated the formation of large numbers of sea spume droplets, suggesting that they can be generated from the Kelvin–Helmholtz shear instability, thereby forming a thin foam layer, reducing the drag coefficient by suppressing the short capillary–gravity waves. Andreas (2004) pointed out that the mass flux of spray droplets falling back to the ocean is comparable to a “heavy rain,” capable of dampening the short waves in high wind speed conditions, making the sea surface visually and aerodynamically smoother. Following these results, Z17 proposed a parameterization for sea surface roughness that includes the influence of sea spray and ocean waves, based on the

assumption that the spray droplets have an effect on the stratification of the air–sea boundary layer. They were able to demonstrate a reduction in the surface drag at high wind speeds that is in fair agreement with available observations. However, a field study by Bell et al. (2012) suggested that C_d remains approximately constant for extremely high wind speeds ($U_{10} > 50 \text{ m s}^{-1}$), based on experimental data from Black et al. (2007). The constancy of C_d in extremely high wind speeds ($U_{10} > 60 \text{ m s}^{-1}$) remains consistent with recent field and laboratory observations (Donelan 2018; Zweers et al. 2010; Soloviev et al. 2014).

Although Z17 made progress in investigating the respective impact of sea spray and wave induced mixing processes on the upper ocean, the collective impact of these processes has received less attention. In a companion paper (Zhang et al. 2018), we have investigated the roles of Langmuir turbulence and surface wave breaking during Super Typhoon Haitang, showing that the Langmuir turbulence-induced mixing can strengthen the sea surface cooling by more than 0.5°C in most typhoon-affected regions. Here, as a follow-on study, the parameterizations of sea spray and wave stress are again included in the coupled wave–ocean model to understand their effect in the response of the upper ocean to a super typhoon, now in addition to the typhoon-induced Langmuir turbulence and wave breaking processes discussed by Zhang et al. (2018). The paper is organized as follows: Ocean and wave model configurations, air–sea dynamics roughness and sea spray parameterization schemes are given in section 2. Experimental design and typhoon cases are described in section 3. The effects of sea spray and wave roughness on the upper ocean are presented in section 4. Discussion and conclusions are given in section 5.

2. Model description

a. POMgcs model

The Princeton Ocean Model with the generalized coordinate system (POMgcs) is a primitive equation, time-dependent, free surface, coastal ocean circulation model, developed by Ezer and Mellor (2004), based on Blumberg and Mellor (1987) (<http://www.ccpo.odu.edu/POMWEB/>). The ocean model is set up for the domain covering seas adjacent to Taiwan ($116^\circ\text{--}127.5^\circ\text{E}$, $18^\circ\text{--}29^\circ\text{N}$), with a horizontal resolution of $1/20^\circ$ and 35 vertical levels. The maximum value of the depth is set to 5035 m in this domain. The hybrid coordinate system is used in the vertical levels configuration, following the topography with a sigma-level grid near the sloping bottom areas, whereas z levels are used above the sigma levels with the upmost five levels at 0.0-, 2.5-, 5.0-, 10.0-, and 15.0-m depths. This resolution is able to resolve the main dynamic and thermodynamic processes under the typhoon, especially the mesoscale eddy induced quasigeostrophic and typhoon-induced upwelling/downwelling processes (Chen et al. 2013). The model has a free surface and uses a split time step technique (Madala and Piasek 1977) to separate the vertically integrated equations (external mode) from the vertical structure equations (internal mode). The external mode portion of the model is

two-dimensional in updating for surface elevation and vertically averaged velocities, with a short time step 3 s. The internal model is three-dimensional in updating for ocean current, temperature, salinity, and the turbulence quantities, using a long time step 180 s in this study. The tidal open boundary values are obtained from the Oregon State University (OSU) TOPEX/Poseidon Global Tidal Model version 7 (TPX07) with eight primary tidal constituents defined as M_2 , S_2 , K_1 , O_1 , N_2 , K_2 , P_1 , Q_1 (<http://people.oregonstate.edu/~erofeevs/>). These tidal harmonics parameters provide elevation and depth-averaged current ellipses at the open boundaries of the ocean model. The model initial and lateral boundary fields are derived from the China Ocean Reanalysis (CORA) dataset (www.odinwestpac.org.cn) for the coastal waters of China and the northwest Pacific (Han et al. 2011). The model forcing fields for the air–sea turbulent fluxes are calculated by using air–sea parameters, including sea surface wind speed, sea surface temperature, sea level pressure, air temperature and humidity, etc. The 10-m wind fields are reconstructed (see section 3a) based on the Joint Typhoon Warning Center and the Cross-Calibrated Multiplatform (CCMP) winds (<http://www.remss.com/measurements/ccmp/>). The other sea–air parameters are obtained from the Climate Forecast System Reanalysis (CFSR) 6-hourly products (<https://climatedataguide.ucar.edu/climate-data>), whose spatial resolution is 0.312° . The bottom topography is derived from NOAA one arc-minute Gridded Global Relief Data (ETOPO1) (<https://ngdc.noaa.gov/mgg/global/global.html>). The Coriolis–Stokes force (CSF), the Craik–Leibovich vortex force (CLVF), and Langmuir turbulence (Harcourt 2015) have been introduced into the circulation model (Zhang et al. 2018) to improve the typhoon-induced turbulent mixing and the advective transports. Detailed descriptions of these components can be seen in the companion paper (Zhang et al. 2018).

b. SWAN model

The Simulating Waves Nearshore (SWAN) wave model (Booij et al. 1999) is implemented to obtain wave parameters, with the same domain as the POMgcs. SWAN simulates directional wave spectra in terms of wavenumber direction bands by solving the spectral action balance equation with source and sink terms, such as wind forcing and wave dissipation. A detailed description of the SWAN model can be found on SWAN’s website (<http://www.swan.tudelft.nl>). The time step in SWAN is 180 s, consistent with the internal mode time step in POMgcs. The wave spectrum is discretized by 36 evenly spaced directions (directional resolution) and 32 logarithmically spaced frequencies (starting at 0.03 Hz). The bottom topography, horizontal resolution, and forcing fields are the same as those of POMgcs.

c. Air–sea fluxes algorithm

1) COARE VERSION 2.6 BULK MODEL

The COARE version 2.6 bulk model (Fairall et al. 1996) is introduced into the POMgcs to calculate air–sea interfacial turbulent fluxes, which represent the feedback processes between the ocean model and air–sea turbulent fluxes model. The

air–sea interfacial sensible heat (H_s), latent heat (H_L), and momentum (M) fluxes are estimated by

$$H_s = \rho_a c_{pa} C_h U_{z_l} (\theta_0 - \theta_{z_l}), \quad (1a)$$

$$H_L = \rho_a L_v C_k U_{z_l} (q_0 - q_{z_l}), \quad (1b)$$

$$M = \rho_a C_d U_{z_l}^2, \quad (1c)$$

where the subscript z_l refers to the atmospheric lowest level, and “0” denotes the water surface. The variables U , θ , and q are the mean horizontal wind speed, potential temperature, and specific humidity, respectively; ρ_a is the air density, c_{pa} represents the specific heat of air at constant pressure, and L_v is the latent heat of water vaporization. The Monin–Obukhov similarity (MOS) is used to determine the heat (C_h), moisture transfer coefficient (C_k), and momentum (C_d , drag coefficient)

$$C_h = c_{hn}^{1/2} c_{dn}^{1/2} \left/ \left\{ \left[1 - \frac{c_{hn}^{1/2}}{a\kappa} \psi_h(\xi) \right] \left[1 - \frac{c_{dn}^{1/2}}{\kappa} \psi_u(\xi) \right] \right\} \right., \quad (2a)$$

$$C_k = c_{kn}^{1/2} c_{dn}^{1/2} \left/ \left\{ \left[1 - \frac{c_{kn}^{1/2}}{a\kappa} \psi_h(\xi) \right] \left[1 - \frac{c_{dn}^{1/2}}{\kappa} \psi_u(\xi) \right] \right\} \right., \quad (2b)$$

$$C_d = c_{dn} \left/ \left[1 - \frac{c_{dn}^{1/2}}{\kappa} \psi_u(\xi) \right]^2 \right., \quad (2c)$$

where a accounts for the difference in scalar and velocity von Kármán constants. The variables κ and ψ are the von Kármán constant (0.4) and the MOS profile function, respectively. The c_{hn} , c_{kn} , and c_{dn} variables are the transfer coefficients in the neutral conditions

$$c_{hn} = \frac{\kappa^2}{\ln(z_l/z_0) \ln(z_l/z_{0t})}, \quad (3a)$$

$$c_{kn} = \frac{\kappa^2}{\ln(z_l/z_0) \ln(z_l/z_{0q})}, \quad (3b)$$

$$c_{dn} = \frac{\kappa^2}{\ln^2(z_l/z_0)}, \quad (3c)$$

where z_{0t} , z_{0q} , and z_0 are the roughness lengths for temperature, humidity, and velocity (sea surface dynamics roughness), respectively.

2) SEA SURFACE ROUGHNESS

The sea surface roughness is expressed as the effective ocean surface roughness $z_{0\text{form}}$ depending on ocean waves through the friction velocity (Charnock 1955), plus a smooth flow limit $z_{0\text{visc}}$ (Smith 1988; Fairall et al. 2003):

$$z_0 = z_{0\text{form}} + z_{0\text{visc}} = \alpha \frac{u_*^2}{g} + \frac{0.11\nu}{u_*}, \quad (4)$$

where α is the “Charnock” parameter, setting to 0.011 in COARE version 2.6 bulk model, and ν is the kinematic viscosity. Ocean waves affect the effective roughness of the ocean

FIG. 1. Comparison of drag coefficient for adjusted Z17 parameterization scheme (black symbols for different wave ages) with observations, as indicated.

surface, for conditions ranging from low to moderate wind speeds, which leads to an increase in C_d with increases in U_{10} (Donelan et al. 2004; Edson et al. 2013). In strong wind speed conditions ($U_{10} > 30 \text{ m s}^{-1}$), sea spray droplets are blown off the crests of waves by appreciably intense breaking, forming a stable layer near the sea surface which dampens the air–sea interface turbulence (Makin 2005). In these conditions, the drag coefficient levels off, and even decreases with increasing wind speeds, as shown in laboratory studies (Takagaki et al. 2012; Curcic and Haus 2020) and ocean field campaigns (Holthuijsen et al. 2012), at very high wind speed conditions.

Therefore, we introduced the effects of wave age (e.g., Drennan et al. 2003; Oost et al. 2002), wave height, and sea spray into a new sea surface dynamics roughness parameterization (Z17), for the full range of wind speeds,

$$\alpha = 0.0847^{(1-1/\omega)} \beta_*^{3(1-1/\omega)/2} n^{1/\omega} \beta_*^{m/\omega}, \quad (5)$$

where n and m are the wave relative parameters (0.42 and -1.03 , respectively; Donelan 1990); wave age β_* is equal to the wave spectral peak velocity divided by friction velocity u_* ; ω is a correction parameter indicating the impact of sea spray on the logarithmic wind profile

$$\omega = \min(1, a_{cr}/\kappa u_*), \quad (6)$$

where a_{cr} represents a critical value for the terminal velocity of falling sea spray droplets (0.64 m s^{-1} ; Makin 2005). Based on Eqs. (5) and (6), the effect of sea spray droplets restrains the air–sea momentum transfer and C_d to level off for high winds (Fig. 2 in Z17). However, recent studies suggest that the values of C_d remain constant (Bell et al. 2012), or even slightly increase (Donelan 2018) for extremely high winds ($>60 \text{ m s}^{-1}$), because the associated two-phase environment suppresses short gravity–capillary waves and alters

the aerodynamic properties of the sea surface (Soloviev et al. 2014). This is still somewhat controversial, due to the very scattered C_d observations in extremely high wind speed conditions. According to these conclusions, the drag coefficient should be approximately constant for extremely high wind conditions. Therefore, Eq. (5) is adjusted in the Z17 parameterization schemes Eq. (7),

$$\begin{cases} \alpha = 0.0847^{(1-1/\omega)} \beta_*^{3(1-1/\omega)/2} n^{1/\omega} \beta_*^{m/\omega}, & U_{10} < 60, \\ \alpha = \alpha_{U_{10}=60}, & U_{10} \geq 60. \end{cases} \quad (7)$$

Hence, the sea surface dynamics roughness in Eq. (7) is introduced into COARE to modify the air–sea turbulent fluxes in our simulations.

Figure 1 shows the variations in drag coefficient according to the adjusted Z17 parameterization Eq. (7). It is clear that C_d may be characterized by peak values at about 32 m s^{-1} , and thereafter declines to approximately constant values for extremely high wind speeds. Moreover, C_d decreases with increases in wave age, which may be explained by differences in the fetch and the distinctive peak enhancement factor dampening (Takagaki et al. 2016). Thus, on the basis of the adjusted Z17 parameterization, C_d is in agreement with the estimated observations, based on direct in situ measurements (Powell et al. 2003; Black et al. 2007; Edson et al. 2013), numerical simulations (Soloviev et al. 2014; Donelan 2018), and laboratory tank measurements (Troitskaya et al. 2012; Takagaki et al. 2012; Curcic and Haus 2020).

d. Sea spray-mediated air–sea turbulent fluxes

1) SEA SPRAY PARAMETERIZATION

The sea spray-mediated momentum flux is calculated using the formulation of Andreas and Emanuel (2001) and Andreas (2004):

$$\tau_{\text{sp}} = \frac{4\pi}{3} \rho_w \int_{r_{\text{lo}}}^{r_{\text{hi}}} u_{\text{sp}}(r) r^3 \frac{dF}{dr} dr, \quad (8)$$

where ρ_w is seawater density, and r represents the initial radius of the sea spray droplets, whose lower and upper radius limits are r_{lo} and r_{hi} , respectively. The variable $u_{\text{sp}}(r)$ indicates the horizontal speeds of the sea spray droplets at the effective height z_{sp} , which is equal to a fraction of the significant wave height $z_{\text{sp}} = 0.63H_s$ for all droplets (Iida et al. 1992). Since the wave amplitude grows with increasing wind speeds, then the corresponding radii of droplets that increase to $500 \mu\text{m}$ will take even longer to accelerate to the local wind speed (Andreas 2004). The $u_{\text{sp}}(r)$ is set independent of r (Andreas 2004); u_{sp} is estimated via the logarithmic profile

$$u_{\text{sp}} = \frac{u_*}{\kappa} \ln \left(\frac{z_{\text{sp}}}{z_0} \right). \quad (9)$$

It is noted that although most of the sea spray droplets up to $500 \mu\text{m}$ in radius are able to have enough time to accelerate to the local wind speed; however, some of the very small spray droplets remain suspended indefinitely (Andreas and Emanuel. 2001). In other words, not all sea spray droplets are able to reach the effective height and fall back to ocean surface to form a sea spray momentum flux. Hence, in terms of the sea spray-mediated momentum flux that we are considering, 30 and $500 \mu\text{m}$ are the droplets lower and upper radius limits, respectively (Zhao et al. 2006). The term dF/dr is spray generation function (SGF), representing the number of droplets produced per square meter per second per micrometer r increment at the ocean surface. Based on available observational data from field and laboratories, Zhao et al. (2006) introduced the effect of wind-sea Reynolds number R_B , wind forcing and wind wave sea state into the spray generation function,

$$dF/dr = \begin{cases} 7.84 \times 10^{-3} R_B^{1.5} r^{-1}, & 30 < r < 75 \mu\text{m}, \\ 4.41 \times 10^1 R_B^{1.5} r^{-3}, & 75 < r < 200 \mu\text{m}, \\ 1.41 \times 10^{13} R_B^{1.5} r^{-8}, & 200 < r < 500 \mu\text{m}, \end{cases} \quad (10)$$

where R_B represents the coupling effect of wind forcing and wind wave sea state, regarded as a measure of the fluid dynamical conditions at the air-sea boundary layer.

Furthermore, an improved parameterization for the sea spray-mediated heat fluxes (Fairall et al. 1994; Z17) is used in this study. It is based on the earlier work on sea spray droplet microphysics and the associated time scales of sea spray droplets (Andreas 1989, 1992). The sea spray-mediated sensible (Q_s) and latent (Q_l) heat fluxes are proportional to the spray droplets mass flux and the area of sea surface, respectively (Fairall et al. 1994):

$$Q_s = 0.5S_v W \gamma(u_{10}) \rho_w c_{pw} (T_s - T_a), \quad (11a)$$

$$Q_l = 0.5S_a h_d W \gamma(u_{10}) \beta(T_a) \rho_a L_e [q_s(T_a) - q], \quad (11b)$$

where S_v and W are the relevant whitecap normalized droplet volume fluxes and the whitecap areal fraction, respectively.

The term $T_s - T_a$ indicates the air-ocean temperature difference; c_{pw} and h_d represent the specific heat of liquid water and evaporation zone scale height, respectively. The term S_a is set to 0.125 s^{-1} ; $q_s(T_a)$ and q are the saturation mixing ratio and the ambient specific humidity, respectively. The variable L_e is the latent heat of condensation; $\gamma(u)$ and $\beta(T_a)$ represent the exchange term and correction factor. These parameterizations are modified by introducing a new spray evaporation zone height and a whitecap areal fraction parameter (Z17). The spray evaporation zone height is higher (lower) than that of the breaking (significant) wave height (Makin 2005); therefore, it is calculated in terms of 1/10 of the significant wave height. Moreover, the whitecap coverage parameters, including the breaking-wave effect R_B from Zhao and Toba (2001) and Guan et al. (2007), is introduced to improve the parameterization for the sea spray effect,

$$W = 3.88 \times 10^{-7} R_B^{1.09}. \quad (12)$$

Details regarding the introduction of the modified Fairall parameterization are given by Z17.

2) FEEDBACK OF SPRAY EVAPORATION AND ENVIRONMENT

Previous studies have suggested that the air-sea interfacial fluxes and the sea spray-mediated fluxes constitute nearly all of the total air-sea fluxes (Peng and Richter 2019). Total sensible $H_{S,T}$, latent $H_{L,T}$, and momentum M_T fluxes are obtained by adding the spray ($Q_{S,\text{sp}}$, $Q_{L,\text{sp}}$, τ_{sp}) and the corresponding bulk interfacial (H_S , H_L , M) fluxes (Andreas and Decosmo 1999):

$$H_{S,T} = H_S + Q_{S,\text{sp}}, \quad (13a)$$

$$H_{L,T} = H_L + Q_{L,\text{sp}}, \quad (13b)$$

$$M_T = M + \tau_{\text{sp}}. \quad (13c)$$

Not all sea spray droplets are able to provide heat exchange at the top of the sea spray evaporation layer. The average temperature and humidity profile are also changed by the existence of sea spray, within the sea spray evaporate layer. In an attempt to account for the complicated processes between the sea spray and the environment at the air-sea interface, Andreas and Decosmo (1999) suggested that the sea spray-mediated net sensible ($Q_{S,\text{sp}}$) and latent heat fluxes ($Q_{L,\text{sp}}$) may be calculated as

$$Q_{S,\text{sp}} = \beta Q_s - (\alpha - \gamma) Q_l, \quad (14a)$$

$$Q_{L,\text{sp}} = \alpha Q_l. \quad (14b)$$

Here, αQ_l is sea spray-mediated latent heat fluxes at the top of the spray evaporation layer in Eq. (14b). In Eq. (14a), $-\alpha Q_l$ represents the atmospheric heat supply for sea spray evaporation. The term βQ_s is the sea spray-mediated sensible heat flux, due to the cooling processes of the sea spray droplets, resulting from their production and subsequent return to the ocean. The evaporation of sea spray changes the air-sea temperature difference, which adds the relative sensible heat flux

FIG. 2. Maximum sustained wind speed (VMAX) and minimum sea level pressure (MSLP) from JTWC, and the reconstructed 10-m wind speeds (m s^{-1}) at (a)–(l) 0000 UTC 16 Jul–1800 UTC 18 Jul 2005. Moving tracks of Typhoon Haitang are depicted with star symbols.

λQ_i . The values of α , β , and γ are given by 2.46, 15.15, and 1.77, respectively (Andreas et al. 2015), obtained from a statistical fitting of the observations from several experiments.

3. Methodology

a. Typhoon cases

Shortcomings in model forcing can constitute an important cause for differences between observations and simulations. To improve the model wind forcing in Super Typhoon Haitang, the reconstructed 10-m wind is used to force the ocean current and wave coupled models in this study, based on the CCMP ocean surface wind analysis and the TC's characteristic parameters from the Joint Typhoon Warning Center (JTWC). First, a TC wind profile model (Carr and Elsberry 1997) is used to produce high-resolution surface winds, based on features of the parameters that characterize the TC, such as estimates of position, maximum sustained surface wind speed (VMAX), minimum sea level pressure (MSLP), and maximum and zero wind speed radii, from the Joint Typhoon Warning Center. This model has been successfully used to produce high resolution surface winds to investigate the thermal and wave responses in the South China Sea for Tropical Cyclones Ernie (1996) and Muifa (2004) (Chu et al. 2000; Chu and Cheng 2008). Thereafter, the constructed 10-m wind fields for Typhoon Haitang are derived through assimilating these high-resolution surface

winds into the CCMP ocean surface wind, via a 3D variational data assimilation approach (Li et al. 2008; Zhang et al. 2018).

Figure 2 shows several examples of the distributions of the reconstructed 10-m horizontal winds during Typhoon Haitang, which formed as a depression just to the east of 150°E in the northwest Pacific on 10 July. It developed to an intense typhoon ($>51 \text{ m s}^{-1}$), which began to enter our region of interest on 16 July. On 17 July, Haitang strengthened to a super typhoon with peak wind speeds exceeding 64 m s^{-1} , and significant wave heights over 10 m. An intense vortex structure developed with a distinct typhoon eye, which extended to the east of the Luzon Strait. The intensity of Haitang gradually reduced on 18 July, as it approached Taiwan and subsequently made landfall.

b. Experimental design

The goal in this study is to understand the response of the upper ocean to forcing by typhoon conditions, as modulated by the effects of sea spray and ocean waves. The model-coupling toolkit (MCT) is used to couple the ocean and wave models in a two-way data exchange. Wave spectrum and parameters from the wave model are the input to the ocean model to obtain sea surface roughness, drag coefficient, and sea spray-mediated turbulent fluxes. In addition, Doppler shift is estimated in the wave model via the wave dispersion relation on the basis of the real-time currents and water levels from the ocean model. When the one-way coupled

TABLE 1. Experimental setups.

Experiment name	New sea surface roughness [Eq. (7)]	Sea spray-mediated heat fluxes [Eq. (14)]	Sea spray-mediated momentum fluxes [Eq. (8)]	Description
CTRL	No	No	No	COARE version 2.6 bulk model [Eq. (1)] is used to estimate air–sea interfacial turbulent fluxes, with the traditional sea surface roughness [Eq. (4), $\alpha = 0.011$] to calculate z_0 and C_d
PS_R	Yes	No	No	As in CTRL, except that the traditional sea surface roughness is substituted for sea surface roughness [Eq. (7)]
PS_S	No	Yes	Yes	As in CTRL, except that sea spray-mediated heat [Eq. (14)] and momentum fluxes [Eq. (8)] are introduced via Eq. (13)
FULL	Yes	Yes	Yes	As in CTRL, except that we use the total air–sea heat and momentum fluxes Eq. (13) via the sea surface roughness dynamics Eq. (7) and the sea spray-mediated turbulent fluxes Eqs. (14) and (8)

system is used between the wave model to the ocean model, the effect of the ocean model on the wave model is disabled, similar to Reichl et al. (2016) and Zhang et al. (2018), at every internal-mode time step.

Four experiments are designed (see Table 1) to investigate the influence of sea spray and sea surface roughness on the response of the upper ocean, based on POMgcs, the SWAN model, the new sea surface roughness Eq. (7), the sea spray-mediated air–sea heat Eq. (14), and momentum Eq. (8) fluxes parameterizations. Tropical cyclones are special atmosphere–ocean coupled systems, where atmosphere and ocean interactions play a key role in the generation and feedback of sea spray and ocean wave (Moon et al. 2004). However, the atmosphere (Zhang et al. 2006; Bao et al. 2011) and ocean (Zhang et al. 2017) physical processes are very complicated for the exploration of the influence of sea spray and ocean wave. To simply and clearly illustrate the effects of sea spray and sea surface roughness on the upper ocean to avoid uncertainties to some degree, the high-resolution ocean model coupled only with the wave model is used to investigate the response of the upper ocean to Super Typhoon Haitang, without including atmospheric numerical model. Hence, the study can be regarded as an initial study to investigate the impact of sea spray and sea surface roughness during the passage of a super typhoon. In addition, authors will have worked in an atmosphere–ocean–wave coupled model by the end of the year to further explore the influence of sea spray and ocean wave on the atmosphere.

The experiments are as follows:

- 1) The control run (CTRL) with the air–sea interfacial turbulent fluxes in Eq. (1), where the traditional sea surface roughness [Eq. (4), $\alpha = 0.011$] is used to calculate z_0 in the COARE version 2.6 bulk model, without effects from sea spray or ocean waves.
- 2) Sea surface roughness Eq. (7) is introduced into the COARE bulk model, taking the place of the traditional sea surface roughness Eq. (4), thereby allowing the

calculation of the new air–sea interfacial turbulent fluxes, denoted PS_R.

- 3) Thermal and dynamic effects of sea spray are included to allow the calculation of the total air–sea heat and momentum fluxes in Eq. (13) via the COARE version 2.6 bulk model and the sea spray-mediated turbulent fluxes, denoted PS_S.
- 4) Effects of sea surface roughness (young ocean waves) and sea spray are introduced into the COARE model to allow the calculation of the total air–sea heat and momentum fluxes Eq. (13) via the new sea surface roughness dynamics Eq. (7) and the sea spray-mediated turbulent fluxes Eqs. (14) and (8), given in the FULL test.
- 5) JTWC dataset are used to provide the Typhoon Haitang's best track information (wind, pressure, track, etc.), in which the wind forcing fields have been optimized by using the CCMP wind analysis and TC's characteristic parameters. The wind, pressure, and track of Haitang remain the same in all model simulations (CTRL, PS_R, PS_S, and FULL).

Detailed experimental setups are described in Table 1.

FIG. 3. Drag coefficient calculated by the sea surface roughness relation [Eq. (7)], as a function of wind speed (m s^{-1}) and wave age.

FIG. 4. (a) The right (21°–150°), rear (151°–240°), and left (241°–20°) three azimuthal sectors along the typhoon's track direction are represented by black lines [following the notation of Black et al. (2007)]. The white line indicates the direction of the typhoon propagation. Particular right (30°, 150°), rear (151°, 235°), and left (260°, 300°) locations are respectively indicated by green, white, and black stars, at 100 km from the typhoon center. Contour lines are the sea surface wind speed (m s^{-1}) on 17 Jul. (b) Model-generated directional wave spectra by SWAN at the right (30°, 150°), rear (151°, 235°), and left (260°, 300°) locations, as indicated in (a). The solid circles indicate wave lengths of 100, 200, 300, and 500 m (outer to inner). The dashed circles correspond to wave lengths of 150, 250, and 350 m. The contours indicate the spectral energy density multiplied by 100 (m Hz^{-1} per degree) on 17 Jul.

FIG. 5. Spatial distribution of (a) drag coefficient C_d by CTRL, (b) drag coefficient C_d by FULL, (c) significant wave height (SWH) (m) by CTRL, and (d) sea surface wind speed (m s^{-1}) by CTRL on 17 Jul. Red line indicates the direction of the typhoon translation. Black lines represent the right (21° – 150°), rear (151° – 240°), and left (241° – 20°) three azimuthal sectors of typhoon's propagation direction.

4. Results

a. Drag coefficient characteristics

To analyze the spatial distribution characteristics of the drag coefficient under Super Typhoon Haitang, the variation of the drag coefficient with sea surface winds and wave age is shown in Fig. 3. The drag coefficient increases with increasing wind speeds, reaching limiting values when wind speeds are about 30 – 35 m s^{-1} . At these wind speeds, the drag coefficient reaches maximal values with wave age values decreasing to 8 – 10 (Liu et al. 2012). When wind speeds continue to increase beyond these values, the corresponding drag coefficient begins to decrease, despite wave ages that are less than 5. For wind speeds reaching 60 m s^{-1} , the drag coefficient remains approximately constant, as affected by the sea spray droplets.

Analyses of Holthuijsen et al. (2012) proposed a dependence of drag coefficient on different azimuthal angles in tropical cyclones, which generally confirmed the findings by Powell et al. (2003). Thus, the distribution characteristics of the drag coefficient should be combined with the directional properties of ocean waves and wind speeds in Typhoon Haitang. Accordingly, in our study, we define the two-dimensional wave spectra as right (21° – 150°), rear (151° – 240°), and left (241° – 20°)

azimuthal sectors of the typhoon's track direction, following Black et al. (2007) and Moon et al. (2004).

Figure 4 shows the directional wave spectrum obtained from SWAN wave model, as Haitang moves northwest, passing to

FIG. 6. (top) Drag coefficient and (bottom) wave age change with wind speeds (m s^{-1}) in right, rear, and left directional sectors on 17 Jul.

FIG. 7. Air-sea fluxes on 17 Jul: (a) sea spray induced momentum fluxes (N m^{-2}), (b) sea spray induced latent heat fluxes (W m^{-2}), (c) sea spray induced sensible heat fluxes (W m^{-2}), (d) interfacial momentum fluxes (N m^{-2}), (e) interfacial latent heat fluxes (W m^{-2}), and (f) interfacial sensible heat fluxes (W m^{-2}) in the FULL results.

the east of the Luzon Strait on 17 July 2005. In the Haitang's right sector, at 30° , the wave field is unimodal with a peak wavelength of 350 m and wave height of 14.5 m. In the right

direction, to the rear, at 150° , the wavelength shortens to about 260 m and wave height decreases to 7.3 m. Thereafter, the spectrum broadens to become bimodal. To the rear, in

FIG. 8. (left) The air-sea momentum (N m^{-2}) and (right) heat fluxes (W m^{-2}) vs wind speed (m s^{-1}). Black circles are interfacial turbulent fluxes; red circles are sea spray induced turbulent fluxes.

FIG. 9. Daily sea surface temperature ($^{\circ}\text{C}$) from AVHRR and AMSRE observations on (a) 15 and (b) 17 Jul, and comparisons with model results of (c) FULL and (d) CTRL. Black dots and blue stars indicate Haitang's track and ARGO floats (1–4), respectively.

directions 155° and 220° , the wave height decreases to 6.2 and 7.5 m, respectively, and particularly at direction 220° begins to recover unimodal characteristics. In the left sectors, at 260° and 300° , the wave spectra exhibit unimodal shapes due to resonance effects, and wave height increases to 12.0 and 13.2 m, respectively. The highest waves propagate on the right side of the storm, because these waves are exposed to prolonged forcing from wind, becoming “trapped” within the typhoon (Moon et al. 2004). By comparison, the fetch and duration of the wind forcing in the rear sector is limited, and the shortest waves are propagated, relatively. Hence, in the rear quadrant, the waves are steeper, shorter, and lower, whereas in the right and left quadrants, the waves are flatter, longer, and higher. We also plotted the wave spectra at two additional distances (150 and 200 km away) from typhoon center to verify that our conclusions are valid in different directional areas (not shown).

Depending on directional properties of the ocean waves and wind speeds, the drag coefficient C_d from Eq. (7) exhibits a variability that is different from results for C_d in the CTRL run. This is seen (below) in comparing Figs. 4 and 5 with Fig. 6. The maximum areas for C_d correspond to the locus of wind speeds (approximately a circle) with values $30\text{--}35\text{ m s}^{-1}$. From the maximum C_d locus to the center of the typhoon eye, the surface wind reaches super speeds ($>60\text{ m s}^{-1}$) in the right quadrants (Fig. 5b). The “trapped fetch” effect produces higher, longer, and older waves, which further decrease the values of sea surface roughness and C_d . In the meantime, the corresponding C_d tends to approximately level off, for super wind speeds

($>60\text{ m s}^{-1}$), and even decline, consistent with Soloviev et al. (2014). Therefore, C_d reaches minimal values in the right forward sector of Typhoon Haitang's track, compared with other directions (Figs. 5 and 6). In addition, values for C_d in the rear quadrant are larger than corresponding values in the right and left forward quadrants (Fig. 6), where the lower, shorter, and younger waves are produced (Fig. 4).

Furthermore, misalignments of winds and waves are found in Haitang (Fig. 5), due to the curvature of typhoon-generated wind fields and their translational directions. The highest wind speeds occur in the right-front sector near the radius of maximum wind, although they do not completely generate the highest waves, there. The maximum directional difference between dominant waves and wind is located in the left-front direction of Haitang, especially during its fast-moving stage (Zhang et al. 2018). This misalignment further influences the effects of Stokes drift and Langmuir turbulence on the ocean mixed layer, during the passage of the typhoon (Rabe et al. 2015).

b. Sea spray-mediated turbulent fluxes

The total air–sea heat and momentum fluxes are directly modulated by sea spray in the air–sea interface. An analysis of the sea spray and interfacial turbulent fluxes for the distribution characteristics of sea spray-mediated turbulent fluxes is shown in Fig. 7, for 17 July. The evaporation of sea spray droplets enhances the air–sea latent heat exchange from the ocean to the atmosphere, while the heat exchange with the

FIG. 10. (a) Difference temperature between posttyphoon (17 Jul) and pretyphoon (15 Jul) in a vertical section along 126°E, in the CTRL run. Differences for model results on 17 Jul for (b) FULL minus CTRL, (c) PS_R minus CTRL, and (d) PS_S minus CTRL results. The vertical dash line indicates the corresponding underwater location under the location of Haitang's center.

droplets increases the sensible heat transfer from the atmosphere to the ocean (Fig. 7). Thus, owing to the complicated physical processes between the sea spray and the air–sea interface, the composite effect of sea spray droplets is to strengthen heat exchange from ocean to atmosphere (Fig. 8). In terms of momentum exchange, the sea spray-mediated momentum flux as estimated from Eq. (8), strengthens the momentum input from atmosphere to ocean (Fig. 7a). For super wind speeds ($>60 \text{ m s}^{-1}$), the sea spray-mediated net latent (sensible) fluxes provide an enhancement of as much as 1096 (-198) W m^{-2} , which is comparable to the interfacial latent fluxes 1620 (192) W m^{-2} . Considering the complicated feedback processes between sea spray and the ambient environment, the sea spray droplets ultimately enhance the air–sea heat transfer from the ocean to the atmosphere by 898 W m^{-2} , which accounts for 47% of the interfacial heat flux. Moreover, the maximum total air–sea momentum fluxes are enhanced by up to 1.2 N m^{-2} , which is an increment of about 16% compared to the interfacial momentum flux (Fig. 8). The primary location of the mediation of sea spray droplets on the turbulent fluxes is around Haitang's eyewall. The maximum regions for the sea spray-mediated turbulent fluxes are the right and left front sectors of Haitang; however, this is not completely consistent with the wind speed distribution, due to the misalignment of winds and waves. By comparison, the minimum area for sea

spray-mediated turbulent fluxes is mostly located in the rear directional sector of Haitang, affected by the influence of wave state and wind speeds in that region.

c. Influence of sea surface roughness and sea spray on the upper ocean

As Super Typhoon Haitang approached close to Taiwan on 17 July, there was a broad low temperature region with maximum sea surface temperature cooling from 30°C on 15 July to 24°C on 17 July, on the right-hand side of Haitang's track (shown in satellite observations in Fig. 9). With the traditional interfacial air–sea turbulent fluxes in the CTRL test, the sea surface cooling response is overestimated by 1°C on Haitang's right side (Fig. 9d). However, in the FULL test, the overestimated sea surface cooling is improved by considering the effects of sea spray and ocean waves; in this simulation, SST decreases to about 24°C on Haitang's right side (Fig. 9c), which is closer to the satellite observations.

To analyze the change in upper ocean temperature during Haitang's passage, we investigate a vertical sea temperature section from the sea surface to 200-m depth along 126°E in Fig. 10. This shows that the upper ocean temperature decreases significantly from the surface to 200-m depth, relative to the pretyphoon condition. A maximum cooling of about 6.5°C is located on the right-hand side of Haitang, from 0 to 20-m depth

FIG. 11. Vertical temperature profile from Argo floats (black lines): 1 (22.98°N, 123.87°E), 2 (24.66°N, 126.46°E), 3 (22.84°N, 122.3°E), and 4 (23.10°N, 122.43°E), with locations indicated in Fig. 9, compared to FULL (red dashed lines), and CTRL (blue lines) model simulations.

(Fig. 10a), which corresponds to the area of maximum wind speeds (Fig. 2). It is evident that cooling occurs in the upper ocean from the surface to the thermocline, whereby isotherms rise in response to Haitang. This phenomenon is called “cold suction” (Chen et al. 2013) produced by the cyclonic wind stress (Park et al. 2011).

The effect of the influence of sea spray-mediated turbulent fluxes (in the PS_S test) suggest that the upper ocean temperature decreases by 0.4°C at a depth of 200 m, with 0.6°C cooling on right side of Haitang (Fig. 10d). Unlike the impact of sea spray-mediated turbulent fluxes, the ocean upper temperatures in PS_R test are warmed by 1.2°C above 160-m depth on Haitang’s right side. Taking into account the competing effects of sea spray and sea surface roughness, in the FULL test, upper ocean temperatures are warmed by 0.6°C on Haitang’s right side, whereas temperatures on the left side are cooled by 0.2°C at the 200-m depth.

Observations of the temperature profiles from the Argo floats can be used to identify the subsurface ocean response near Haitang’s track, as presented in Fig. 11. Before Haitang’s passing, sea spray and sea surface roughness hardly affect the upper ocean temperature profiles, as indicated on 14 July (Argo floats 2 and 4). As Haitang passed over Argo floats 1 and 3, the upper ocean cooling processes are enhanced by inclusion of the influences of sea spray and sea surface roughness in the FULL run, giving results that are notably closer to the Argo observations, than those of the CTRL run.

At 1800 UTC 17 July, Argo float 1 is located at the left outer edges of Haitang’s track. Thus, affected by the upper turbulent mixing processes, it is evident that the ocean near-surface temperatures are mixed downward to 40-m depth with a uniform 28°C (black line in Fig. 11). From 40- to 80-m depth, the

ocean subsurface temperatures decrease to 25.5°C. The effects of sea spray and sea surface roughness modulate turbulent kinetic energy, extending downward on the left side of the typhoon (Fig. 12), strengthening the mixing between the warm surface water and the cold subsurface water, in the water column. Thus, compared to the CTRL test, the FULL test results are more consistent with the Argo vertical temperature structure, showing ocean near-surface temperatures that are cooled by 0.2°C, and ocean subsurface temperatures that are warmed by 0.2°C. Therefore, in terms of cooling, overestimates on the right side, and underestimates on the left side of Haitang, are modified by considering the impacts of sea spray and ocean wave roughness.

However, the subsurface ocean temperature cooling and mixed layer depth in the FULL simulation are not totally consistent with the observations (Argo and satellite), possibly due to the shortcomings in the nonlocal nature of convective mixing in the mixing parameterization scheme (Kantha 2012), even with the inclusion of the effects of sea spray and ocean waves. Although the constructed 10-m wind fields are able to significantly capture much of Haitang’s position and maximum surface wind field (Fig. 2), they still do not exactly simulate the actual typhoon wind fields. Thus, the ocean cooling area on Haitang’s right side is slightly wider in the simulation than the corresponding counterpart indicated in satellite observations.

5. Discussion and conclusions

High-resolution ocean and wave model systems are used, with robust parameterizations schemes for sea surface roughness and sea spray fluxes, to investigate the response of the upper ocean to Super Typhoon Haitang. The drag coefficient

FIG. 12. Vertical profile of (right) K_M ($\text{m}^2 \text{s}^{-1}$) and (left) q^2 ($\text{m}^2 \text{s}^{-1}$) at Argo float 1 located at 22.98°N , 123.87°E from FULL (red line) and CTRL (green line) simulations on 17 Jul 2005.

C_d is derived from sea surface dynamics roughness and exhibits variability depending on the characteristics of both the sea surface wind and the ocean wave state. The drag coefficient C_d reaches a minimum in the right quadrant of the typhoon track, where the “trapped fetch” effect generates unimodal wave spectra with long, smooth, old waves. By contrast, low, short, young waves are generated in the rear quadrants, according to the limited fetch and duration conditions of wind forcing in the rear quadrants of Haitang, where the values of C_d are larger than counterparts in the right and left quadrants. Once wind speeds increase to values of $30\text{--}35 \text{ m s}^{-1}$, values for C_d become maximal, as wave ages decrease to 8–10.

The air–sea heat exchange from the ocean to the atmosphere is ultimately enhanced by the evaporation and heat transfer of sea spray droplets. With the inclusion of sea spray to mediate the stress, the momentum transfer from the atmosphere to the ocean is strengthened. Sea spray–mediated turbulent fluxes primarily are located on the front sectors of Haitang, while areas of minimal effect occur in the rear region. Because of the misalignment of wind speeds and ocean waves, the distribution characteristics of sea spray–mediated heat fluxes are not completely consistent with the wind speeds. For super wind speeds ($>60 \text{ m s}^{-1}$), sea spray–mediated net heat and momentum fluxes are 898 W m^{-2} and 1.2 N m^{-2} , accounting for 47% and 16% of the interfacial heat and momentum fluxes, respectively.

With the traditional interfacial air–sea turbulent fluxes, the sea surface cooling response to Haitang is overestimated

by 1°C on the right side of Haitang. Compared to the traditional interfacial air–sea turbulent fluxes, upper ocean temperatures on the right side (left side) are warmed (cooled) by 0.6°C (0.2°C) from the sea surface to 200-m depth, taking into account the effects of both sea spray and sea surface roughness. Sea surface cooling on Haitang’s right side increases to 24°C , consistent with the AVHRR and AMSRE satellite observations. In the latter case, the simulated upper ocean temperatures are closer to Argo observations. Moreover, the overestimates (underestimates) in cooling on the right-hand (left-hand) side of Haitang, are diminished by the sea spray–mediated turbulent fluxes and ocean wave roughness, compared with the traditional interfacial air–sea turbulent fluxes.

The proposed ocean surface roughness and sea spray parameterizations may help improve surface flux parameterizations and the upper turbulent mixing processes. This study attempts to analyze the effects of sea spray and sea surface roughness on the upper ocean, as generated by a super typhoon, based on the reconstructed wind forcing fields. The results from CTRL show that air–sea turbulent fluxes calculated from the original COARE version 2.6 bulk model, induce not only *overestimates* in ocean surface cooling due to super wind conditions (CTRL in Fig. 9), but also *underestimates* in ocean subsurface cooling (Argo in Fig. 11). To compensate for these deficiencies, to some degree, air–sea turbulent fluxes and drag coefficient are modified by introducing the parameterizations for sea spray and the sea surface roughness. Although there is still considerable uncertainty in the magnitude of the

air–sea fluxes and the corresponding drag coefficients, these estimates provide some new evidence for the effects of ocean waves and sea spray droplets on the upper ocean temperature from a qualitative view.

In addition, the COARE bulk model is used to calculate the air–sea interfacial turbulent fluxes. In this approach, the feedback processes between SST and air–sea turbulent fluxes within the ocean and wave coupled models are considered. It is well known that tropical cyclones are special atmosphere–ocean coupled systems, where the atmosphere and ocean interact through relatively complicated physical processes. Atmosphere and ocean interactions are very important for the generation of sea spray generation, for feedbacks and simulations of impacts in the air–sea interface. The effect of sea spray droplets on the turbulent atmospheric marine near-surface boundary layer is similar to that on the ocean temperature stratification. Once the wave and ocean models are dynamically coupled with the atmosphere model, the wind, sea level pressure and propagation track of Typhoon Haitang can be modulated by the effects of sea spray and sea surface roughness in simulations (Zhang and Perrie 2008). Therefore, these results provide motivation for the application of these parameterizations in atmosphere–ocean–wave coupled models (Aijaz et al. 2017) to investigate the feedback processes between sea spray and the air–sea interface in future studies.

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