Effects of Wave–Current Interactions on Bay–Shelf Exchange

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ABSTRACT: Bay–shelf exchange is critical to coastal systems because it promotes self-purification or pollution dilution of the systems. In this study, the effects of wave–current interactions on bay–shelf exchange are explored in a micromesotidal system—Daya Bay in southern China. Waves can enlarge the shear-induced seaward transport and reduce the residual-current-induced landward transport, which benefits the bay–shelf exchange; however, tides work oppositely and slow the wave-induced bay–shelf exchange through vertical mixing and reduced shear-induced exchange. Five wave–current interactions are compared, and it is found that the depth-dependent wave radiation stress (WRS) contributes most to the bay–shelf exchange, followed by the wave dissipation as a source term in the turbulence kinetic energy equation, and the mean current advection and refraction of wave energy (CARWE). The vertical transfer of wave-generated pressure to the mean momentum equation (also known as the form drag) and the combined wave–current bottom stress (CWCBS) play minor roles in the bay–shelf exchange. The bay–shelf exchange is faster under southerly wind than under northerly wind because the bay is facing southeast; synoptic events such as storms enhance the bay–shelf exchange. The CARWE terms are dominant in both seasonal and synoptic variations of the bay–shelf exchange because they can considerably change the distribution of significant wave height. The WRS changes the bay–shelf exchange mainly through altering the flow velocity, whereas the wave dissipation on turbulence alters the vertical mixing. The form drag and the CWCBS have little impact on the bay–shelf exchange or its seasonal and synoptic variations.

KEYWORDS: Mass fluxes/transport; Ocean dynamics; Wind waves; Numerical analysis/modeling

1. Introduction

In recent decades, the rapid increase in terrigenous nutrients and other pollutants has caused severe marine environmental problems and the degeneration of coastal ecosystems. Thus, understanding the bay–shelf exchange is critical to coastal systems, which could promote the self-purification of the systems and have important implications for the transport of nutrients, larvae, pollutants, heat, and salt on inner continental shelves (Fewings et al. 2008). Various definitions of exchange time were proposed, employed, and compared to quantify the exchange capability in early works (e.g., Prandle 1984; Luff and Pohlmann 1995; Delhez et al. 2004; Lemagie and Lerczak 2014), such as flushing time, residence time, turnover time, and half-life exchange time. Due to the nature of dilution, pollutants in water will never reach zero, and thus, Luff and Pohlmann (1995) defined the half-life exchange time analogously to that of radioactive substances. A short half-life exchange time indicates a strong exchange capability of a coastal embayment.

In previous studies, several physical processes have been found to dominate the bay–shelf exchange in a coastal embayment, such as river runoffs (Okada et al. 2011; Kenov et al.

2012; Ren et al. 2014; Du and Shen 2016; Wang et al. 2016), tides (Banas and Hickey 2005; Xiao et al. 2019; J. Zhang et al. 2019), winds (Gong et al. 2009; Pfeiffer-Herbert et al. 2015; Kang et al. 2017), coastal circulations (Zu and Gan 2015; H. Zhang et al. 2019), and synoptic events (Pareja-Roman et al. 2019; Ding et al. 2019). It is well known that the wave–current interactions are of great importance to inner shelf hydrodynamics and sediment transport dynamics (Longuet-Higgins and Stewart 1964; Graber et al. 1989; Smith 2006; Wang et al. 2007; Lentz et al. 2008; Sheng and Liu 2011; Dufois et al. 2014; Gao et al. 2018). However, the effect of wave–current interactions on the bay–shelf exchange has been less mentioned. In most studies, the wave–current interactions were investigated through the wave radiation stress, the wave-induced friction, Stokes drift, and wave-driven flow (Lentz and Fewings 2012). Through the observed Eulerian wave-driven cross-shelf transport, Lentz et al. (2008) found it can be substantial over the inner shelf but may not be effective at driving the cross-shelf exchange due to opposite directions of Eulerian flow and Stokes drift; thus, the wave-driven flow may be a less effective mechanism for particle exchange. Fewings and Lentz (2011) also pointed out that the contribution of wave-driven fluctuation in velocity and temperature through Stokes drift is substantial of the cross-shelf heat flux in summer, whereas the net effect of waves on heat budget may be small.

Previously, Gao et al. (2018) examined the effects of different wave–current interactions on suspended sediment transport, including the depth-dependent wave radiation stress, vertical transfer of wave-generated pressure to the mean momentum

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equation (also known as the form drag), wave dissipation as a source term in the turbulence kinetic energy equation, and mean current advection and refraction of wave energy, which were proposed by Mellor (Mellor 2003; Mellor et al. 2008; Mellor 2015). They found these mechanisms can cause significant variations in current velocities, vertical mixing and even the bottom stress, especially during strong wave events; although the combined wave–current bottom stress has the largest effect on suspended sediment transport, the others are comparable. In this study, taking Daya Bay (DYB) in southern China as an example, the same methods will be used to examine the effect of different wave–current interactions on the bay–shelf exchange. The tidal range in DYB is less than 1 m during neap tides but larger than 2 m during spring tides, indicating a micro- to mesotidal system, which is also a sexta-diurnal tidal resonant bay (Song et al. 2016). J. Zhang et al. (2019) found that the resonant sexta-diurnal tides can enhance the exchange between the DYB and the northern South China Sea shelf. H. Zhang et al. (2019) found the large-scale coastal circulation and local winds are main drivers of the bay–shelf exchange in this region; they also mentioned the well-developed stratification in summer and the extreme weather conditions (e.g., tropical storm and typhoon) would intensify the bay–shelf exchange. However, they did not consider waves and the resonant sexta-diurnal tides in their numerical model.

The weak tidal dynamic condition in DYB provides an ideal region for this study. The half-life exchange time will be used to quantify the bay–shelf exchange with a special focus on the wave–current interactions in DYB. The forcings of tides, winds, and waves on the bay–shelf exchange will also be quantified. The paper is organized as follows. First, the model is validated on wave simulation and numerical experiments are described in section 2. The half-life exchange time is calculated and compared on the basis of model results in section 3. Mechanisms of wave–current interaction on the bay–shelf exchange are discussed in section 4, as well as the seasonal and synoptic variations of the bay–shelf exchange. Section 5 offers the conclusions of this study.

2. Study site and methods

a. Study site

Daya Bay is a semiencloded bay with an area of 560 km², which is located on the south coast of China (114°29.7′–114°49.7′E, 22°31.2′–22°50.0′N), approximately 70 km east of the Pearl River estuary (Fig. 1). DYB is composed of several subbasins, including Aotou (AT) harbor and Baishou (BS) cove in the northwest, Fanhe (FH) harbor in the northeast and Dapeng (DP) cove in the southwest. There are more than 50 islands inside the bay, such as the Mabian (MB) island, Dalajia (DLJ) island, and Huangmao (HM) island, but no major rivers discharged into the bay (J. Zhang et al. 2019). Outside the bay, there are two major islands, the Sanmen (SM) island in the southwest and the Xiaoxing (XX) island in the southeast. The water depth in DYB is 9.7 m on average, with a maximum of about 21 m.

In recent decades, human activities have generated severe impacts on the DYB environment. A total water area of 30 km² was reclaimed, and the average water depth was increased by 38 cm from 1989 to 2014 (J. Zhang et al. 2019). The increased water depth was mainly due to the loss of tidal flat, dredging for ports, and deepening of water channels. The Deep Navigation Channel (DNC) from the bay mouth to the MB island is remarkable in Fig. 1, which is 20.2 km in length, 300 m in width, and 20 m in depth. It was completed in 2007 and is regularly dredged for shipping safety. The cross section E (15.1 km in length) at the bay mouth can be divided into three channels: the west channel between the west coast and the DLJ island, the middle channel from the DLJ island to the HM island, and the east channel between the HM island and the Hushan (HS) cape. The diurnal and semidiurnal tidal energy enters the bay from the west channel and the middle channel and leaves from the east channel (Song et al. 2016; J. Zhang et al. 2019).

b. Model setup

The numerical model employed in this study is the Mellor–Donelan–Oey (MDO) wave model (Mellor et al. 2008) coupled with the unstructured-grid, primitive-equation, Finite Volume Community Ocean Model (FVCOM; Chen et al. 2003). The MDO wave model considers the wave energy as a function of the wave propagation direction θ , the horizontal coordinates (x, y) , and time t . The wave frequency σ depends on the wave direction rather than an independent variable. It uses a specified spectrum shape to parameterize the wave–wave interaction processes and the energy distribution in the frequency space, which is based on the spectrum of Donelan et al. (1985). Therefore, as compared with some popular third-generation wave models, the model is simple but has comparable performances to those models (Mellor et al. 2008). Furthermore, the computational resources required are two orders less than the third-generation wave models (Mellor et al. 2008; Marsooli et al. 2017). The directionally dependent wave energy equation in a sigma coordinate system was given as (Mellor et al. 2008)

$$\frac{\partial E_{\theta}}{\partial t} + \underbrace{\frac{\partial}{\partial x_{\alpha}}[(\bar{c}_{g\alpha} + \bar{u}_{A\alpha})E_{\theta}] + \frac{\partial}{\partial \theta}(\bar{c}_{\theta}E_{\theta})}_{\text{Exp9}} + \int_{-1}^0 \bar{S}_{\alpha\beta} \frac{\partial U_{\alpha}}{\partial x_{\beta}} D d\varsigma = S_{\theta\text{in}} - S_{\theta\text{Sdis}} - S_{\theta\text{Bdis}}, \quad (1)$$

where E_{θ} is the spectrally integrated wave energy; x_{α} and x_{β} are the horizontal coordinates ($\alpha = x, y$ and $\beta = x, y$); $\bar{c}_g$ is the group speed; $\bar{u}_A$ is the Doppler velocity formulated in Mellor (2003, 2008); $\bar{c}_{\theta}$ is the refraction speed; $D = h + \eta$ is the total water depth, where h is the local water depth and η is the surface elevation; $S_{\alpha\beta}$ is the wave radiation stress term; U_{α} is the ocean current plus the Stokes drift; $\varsigma = (z - \eta)/D$ is the sigma variable with z being the vertical coordinate; and the three terms on the right-hand side (RHS) of Eq. (1) are the wind growth source term $S_{\theta\text{in}}$ and the dissipation terms due to wave processes at the surface $S_{\theta\text{Sdis}}$ and bottom $S_{\theta\text{Bdis}}$. The overbar represents spectral average. More details can be found in Mellor et al. (2008).

FIG. 1. The bathymetry map of DYB, situated on the south coast of China, in which BG, YMK, and DC are the wave buoys. The red line (cross section E, 15.1 km in length) indicates the entrance of the bay.

The depth-dependent wave radiation stress was given by Mellor (2015):

$$S_{\alpha\beta} = kE \left(\frac{k_{\alpha} k_{\beta}}{k^2} F_{CC} F_{CS} - \delta_{\alpha\beta} F_{SS} F_{SC} \right) + \delta_{\alpha\beta} \frac{E}{2D} \frac{\partial}{\partial \xi} (2F_{CC} F_{SS} - F_{SS}^2), \quad (2)$$

where k is the wavenumber, E is the wave energy, $\delta_{\alpha\beta}$ is a Kronecker delta function ($=1$ when $\alpha = \beta$ or 0 otherwise),

$$F_{SS} = \frac{\sinh kD(1+s)}{\sinh kD}, \quad F_{SC} = \frac{\sinh kD(1+s)}{\cosh kD}, \\ F_{CS} = \frac{\cosh kD(1+s)}{\sinh kD}, \quad \text{and} \quad F_{CC} = \frac{\cosh kD(1+s)}{\cosh kD}.$$

The FVCOM hydrodynamic model was used to study the tidal resonance (Song et al. 2016) and the impacts of human activities on tidal resonance in DYB (J. Zhang et al. 2019). The model is three-dimensional, but barotropic with constant temperature and salinity, and no river is considered in this

situation. To reproduce the resonant shallow-water tides, twelve tidal constituents are given at the open boundary, including four diurnal tides (K_1 , O_1 , P_1 , and Q_1), four semidiurnal tides (M_2 , S_2 , N_2 , and K_2), two quarter-diurnal tides (M_4 and MS_4), and two sexta-diurnal tides (M_6 and MS_6). To deal with the tidal flooding and drainage process, a critical depth of 0.01 m is adopted for the wet/dry treatment. The model has a fine horizontal resolution (about 50 m) near the coast and a coarse resolution (5–6 km) at the open boundary, with seven sigma layers in the vertical. The effect of additional vertical resolution (15 sigma layers we tested) is minor for the unstratified simulations considered here, but it costs much more computing resource. As the model was well validated on tidal simulation by Song et al. (2016), especially the resonant shallow-water tides, the same model configuration and open boundary forcings are used in this study. More details can be found in Song et al. (2016).

The full set of FVCOM equations can be found in Chen et al. (2003). Here, the momentum and continuity equations considering the wave effects are given in the sigma coordinate system:

$$\frac{\partial \eta}{\partial t} + \frac{\partial DU_\alpha}{\partial x_\alpha} + \frac{\partial \omega}{\partial s} = 0 \quad \text{and} \quad (3)$$

$$\frac{\partial(DU_\alpha)}{\partial t} + \frac{\partial(DU_\alpha U_\beta)}{\partial x_\beta} + \underbrace{\frac{\partial(DS_{\alpha\beta})}{\partial x_\beta}}_{\text{Exp8}} + \frac{\partial(\omega U_\alpha)}{\partial s} + DF_f = -gD \frac{\partial \eta}{\partial x_\alpha} - \frac{gD^2}{\rho_0} \int_{-1}^0 \left(\frac{\partial \rho}{\partial x_\alpha} - \frac{s}{D} \frac{\partial D}{\partial x_\alpha} \frac{\partial \rho}{\partial s} \right) ds + DF_h + \frac{\partial \tau_{T\alpha}}{D \partial s} + \underbrace{\frac{\partial \tau_{P\alpha}}{D \partial s}}_{\text{Exp6}}. \quad (4)$$

Note that U_α is the horizontal velocity representing current plus the Stokes drift, whereas ω is the vertical velocity in sigma coordinates; ρ is the water density; ρ_0 is the reference density; F_f is the Coriolis force vector ($-fv, fu$) and f is the Coriolis parameter; g is the acceleration due to gravity; F_h is the horizontal diffusion terms, calculated by the Smagorinsky eddy parameterization method (Smagorinsky 1963); $\tau_{T\alpha}$ is the turbulent-viscous part of the wind stress or skin friction, defined as $K_m(\partial U_\alpha / \partial s)$, where K_m is calculated using the Mellor–Yamada level-2.5 turbulence closure model (Mellor

and Yamada 1982); and $\tau_{P\alpha}$ is the wind stress induced by the form drag (Mellor et al. 2008):

$$\tau_{P\alpha} = \frac{\cosh[2kD(1+s)]}{2\pi \sinh(2kD)} \int_0^{2\pi} P_{w0} \sin \varphi \frac{\partial \hat{\eta}}{\partial x_\alpha} d\theta, \quad (5)$$

where P_{w0} is the wind pressure at the surface and $\hat{\eta} = a \cos \varphi$ is the water elevation caused by waves, with a being the amplitude and $\varphi = k(x \cos \theta + y \sin \theta) - \omega t$ being the phase.

Considering the input of τ_T and τ_P , the turbulence kinetic energy (TKE) equation was given by Mellor (2013):

$$\frac{\partial q^2 D}{\partial t} + \frac{\partial q^2 U_\alpha D}{\partial x_\alpha} + \frac{\partial q^2 \omega}{\partial s} = \frac{\partial}{\partial s} \left(\frac{K_q}{D} \frac{\partial q^2}{\partial s} \right) + 2\tau_{T\alpha} \frac{\partial U_\alpha}{\partial s} + \underbrace{2\tau_{P\alpha} \frac{\partial U_\alpha}{\partial s}}_{\text{Exp6}} + 2K_h \frac{g}{\rho_0} \frac{\partial \rho}{\partial s} - \frac{2Dq^3}{B_1 l} + F_q \quad \text{and} \quad (6)$$

$$\frac{\partial q^2 l D}{\partial t} + \frac{\partial q^2 l U_\alpha D}{\partial x_\alpha} + \frac{\partial q^2 l \omega}{\partial s} = \frac{\partial}{\partial s} \left(\frac{K_q}{D} \frac{\partial q^2 l}{\partial s} \right) + E_1 l \left(\tau_{T\alpha} \frac{\partial U_\alpha}{\partial s} + \underbrace{\tau_{P\alpha} \frac{\partial U_\alpha}{\partial s}}_{\text{Exp6}} + E_3 K_h \frac{g}{\rho_0} \frac{\partial \rho}{\partial s} \right) \tilde{w} - \frac{Dq^3}{B_1} + F_l. \quad (7)$$

Here, q^2 is 2 times the turbulence energy and l is the turbulence length scale; K_q is the mixing coefficient and K_h is the vertical thermal diffusion coefficient; $\tilde{w}$ is the so-called wall proximity function; F_q and F_l are the horizontal diffusion terms for q^2 and $q^2 l$; and E_1 , E_3 , and B_1 are empirical constants. More details can be found in Mellor and Yamada (1982) and Galperin et al. (1988). Note that the buoyancy terms involving gradients of ρ in the momentum and TKE equations are actually not active for unstratified simulations, that is, the second term on the RHS of Eq. (4) and the fourth terms on the RHS of Eqs. (6) and (7).

The turbulence kinetic flux q^2 induced by surface wave dissipation was given by Mellor and Blumberg (2004) as

$$\begin{cases} \frac{\partial q^2}{\partial z} = \frac{2\alpha_{CB} u_{\tau s}^3}{K_q} \\ l = \max(\kappa z_w, \kappa z) \end{cases} \quad \text{at } z = \zeta(x, y, t), \quad (8)$$

where

$$\alpha_{CB} = 15(c_p/u^*) \exp[-(0.04c_p/u^*)^4]$$

is a parameter related to wave age c_p/u^* , c_p is the phase speed of the wave at the dominant frequency, $u^* = 30u_{\tau s}$ is the air friction velocity, $\kappa = 0.41$ is the von Kármán number, $z_w = 0.85H_s$ is the wave-related roughness, and H_s is the significant

wave height. The turbulence kinetic flux q^2 induced by bottom wave dissipation was given by Mellor (2013) as

$$\begin{cases} \frac{\partial q^2}{\partial z} = \int_{-\pi}^{\pi} S_{\theta \text{Bdis}} d\theta \\ l = 0 \end{cases} \quad \text{at } z = -h(x, y). \quad (9)$$

Furthermore, the bottom stress is enhanced by the nonlinear interaction of waves and current in the wave boundary layer. The current-induced shear stress τ_c is calculated using the quadratic friction law in this model, and the wave-induced shear stress τ_{wm} is calculated as

$$\tau_{wm} = \frac{1}{2} f_{cw} u_b^2 \quad (10)$$

where f_{cw} is a wave–current friction factor and u_b is the bottom wave orbital velocity; the combined wave–current bottom stress $\tau_{wc} = \tau_c + \tau_{wm}$ is calculated using the method proposed by Soulsby (1995).

The coupling between waves and currents includes depth-dependent wave radiation stress, Stokes drift velocities, vertical transfer of wave-generated pressure to the mean momentum equation caused by the form drag, wave dissipation as a source term in the TKE equation, mean current advection and refraction of wave energy, and combined wave–current bottom stress (Mellor 2003; Mellor et al. 2008; Mellor 2015). Details on

the two-way coupling between the MDO wave model and the FVCOM hydrodynamic model can also be found in Gao et al. (2018). It has been proved computationally efficient and well validated in theoretical experiments, and successfully applied in Lake George, Australia; the Gulf of Mexico (Wang and Shen 2010, 2012); and Jiaozhou Bay, China (Gao et al. 2018).

The $0.2^\circ \times 0.2^\circ$ sea surface wind is obtained from the National Centers for Environmental Prediction Climate Forecast System, version 2 (CFSv2), Selected Hourly Time Series products (Saha et al. 2011, 2014). The hourly wave variables (significant wave height, peak wave direction, and peak wave period) are obtained from the WaveWatch III (WW3) Global Wave Model (available at <https://data.noaa.gov/dataset/dataset/wavewatch-iii-ww3-global-wave-model2> or https://www.pacioos.hawaii.edu/metadata/ww3_global.html), which is employed as the open boundary conditions for the wave model. The model is cold started and run from 1 December 2018 to 1 January 2020.

c. Model validation

The model was validated with the measurement of tidal elevations and tidal currents by Song et al. (2016). Therefore, in this study, we only compare the model results with the measured waves at three buoy stations (Fig. 1). The wave was measured from 0000 UTC 20 July to 1530 UTC 15 August 2019 with an interval of 30 min. The spatial-averaged (in the range of Fig. 1) CFSv2 wind magnitude and wind vector of the year 2019 is plotted in Fig. 2a with a zoom-in window during the above period (Fig. 2b). It shows a strong southeasterly wind with the largest magnitude of 11.1 m s^{-1} between 30 July and 3 August, when Tropical Storm Wipha hit DYB. Correspondingly, the observed significant wave height H_s can reach 0.6, 1.2, and 3.8 m at BG, YMK, and DC stations, respectively (Fig. 2c). The model can capture main characteristics of the observed H_s . The correlation coefficients between the simulated and observed H_s are 0.79, 0.94, and 0.89 for BG, YMK, and DC stations, respectively. The determination coefficients R^2 between the simulated and observed H_s are 0.60, 0.68, and 0.74 for BG, YMK, and DC stations, and the overall R^2 can reach 0.86 (Fig. 2d). The root-mean-square (RMS) errors are 0.0648, 0.1379, and 0.2697 for the three stations, and the overall RMS error is 0.1779. Note that the imported wave energy from the open boundary is important for this coastal region model. Thus, the performance of the WW3 Global Wave Model in this region has a great impact on the model results. Because of the $0.5^\circ \times 0.5^\circ$ spatial resolution, the WW3 Model seems unable to provide elaborate information for our DYB model. Furthermore, the simplified wave processes might be another reason for some mismatches between the model results and observations. But in general, we think the model results are acceptable to study the effects of wave–current interactions on the bay–shelf exchange in this region.

d. Model experiments

To calculate the half-life exchange time, the tracer-tracking module provided by FVCOM is used in this study. The tracer-tracking equation is given as

$$\frac{\partial(DC)}{\partial t} + \frac{\partial(DU_\alpha C)}{\partial x_\alpha} + \frac{\partial(\omega C)}{\partial \varsigma} = \frac{1}{D} \frac{\partial}{\partial \varsigma} \left(K_h \frac{\partial C}{\partial \varsigma} \right) + DF_C, \quad (11)$$

where C is the tracer concentration and F_C is the horizontal diffusivity calculated using the Smagorinsky eddy parameterization (Smagorinsky 1963). The tracer is initialized with $C = 1$ inside the bay (north of the cross section E in Fig. 1) and $C = 0$ elsewhere. The tracer is released at 0000 UTC 1 January 2019. Then the half-life exchange time is calculated as the period from the tracer release to when the inside-bay averaged tracer concentration reaches 0.5. Numerical experiments are designed to compare the half-life exchange time under different driving forces and wave–current interaction processes (Table 1).

Exp1 is the control run, in which winds, tides, waves, and all wave–current interaction processes are included. Exp2 (No Waves) only considers winds and tides; and Exp3 (Tides Only) is driven by tides only. In Exp4 (No Tides), only winds and waves are included in the model. From Exp5 to Exp9, an individual wave–current interaction process is removed from Exp1 to investigate their contribution to the bay–shelf exchange. In Exp5, the combined wave–current bottom stress (CWCBS) $\tau_{wc} = \tau_c + \tau_{wm}$ is replaced by the current bottom stress τ_c only. In Exp6, the form drag τ_{Pa} is not considered, that is, the last term on the RHS of Eq. (4) and the third term on the RHS of Eqs. (6) and (7). In Exp7, the source term equal to wave dissipation is removed from the TKE equation, and thus Eqs. (8) and (9) are changed to

$$\begin{cases} \frac{\partial q^2}{\partial z} = 0 \\ l = \kappa z \end{cases} \quad \text{at } z = \zeta(x, y, t) \quad \text{and} \quad (12)$$

$$\begin{cases} \frac{\partial q^2}{\partial z} = 0 \\ l = 0 \end{cases} \quad \text{at } z = -h(x, y). \quad (13)$$

In Exp8, the depth-dependent wave radiation stress (WRS) is removed from the momentum equation, that is, the third term on the left-hand side (LHS) of Eq. (4). In Exp9, the current advection and refraction of wave energy (CARWE) are excluded in Eq. (1), that is, the second and third terms on the LHS of Eq. (1).

All experiments are summarized in Table 1. Note that the Stokes drift is not discussed in this study. Indeed, it is included in the momentum equations when the wave–current interaction terms are present, and Mellor (2005) claimed that it would be difficult and unnecessary to separate the Stokes drift from the ocean current. In the MDO wave model, it is only separated once to calculate the turbulence momentum mixing term (Mellor et al. 2008), which has little impact on the wave–current interactions.

e. Decomposition of cross-sectional fluxes

To study the dispersion of salt in estuaries, Dyer (1974) proposed a flux decomposition method to show the mean salt flux through a cross section. Here it is used to examine the tracer concentration flux through the cross section E (Fig. 1), which can be calculated as

FIG. 2. (a) The wind speed (blue) and vector (black) throughout 2019; (b) detailed wind vector from 20 Jul to 15 Aug [the period within the red-outlined box in (a)]. The simulated significant wave height is also compared with the observations [for the same time interval as in (b)] (c) over time and (d) as a scatterplot.

$$F = \overline{A \langle UC \rangle} = \underbrace{\overline{A} \times \overline{\langle U \rangle} \times \overline{\langle C \rangle}}_{T1} + \underbrace{\overline{\langle C \rangle} \times \overline{A_t U_z}}_{T2} + \underbrace{\overline{\langle U \rangle} \times \overline{A_t C_z}}_{T3} + \underbrace{\overline{A} \times \overline{U_z C_z}}_{T4} + \underbrace{\overline{A_t U_z C_z}}_{T5} + \underbrace{\overline{A} \times \overline{\langle U_{z,d} C_{z,d} \rangle}}_{T6} + \underbrace{\overline{A_t \langle U_{z,d} C_{z,d} \rangle}}_{T7} \quad (14)$$

where $U = U(z, t)$ is the velocity at depth z and time t obtained from the momentum equation, which is perpendicular to the cross section; $C = C(z, t)$ is the tracer concentration; and $A = A(t)$ is the cross-sectional area. Note that the unit of flux F is meters cubed per second in this study because of the unit tracer concentration. The overbar was defined as tidal average in Dyer (1974), whereas it denotes the temporal mean in this study as the 1-year length simulation data is used; the angle brackets denote the depth mean. At any depth let $U = \langle U \rangle + U_z$ and $C = \langle C \rangle + C_z$, where U_z and C_z are the deviation of the observation from the depth mean. Because of tidal fluctuations, $\langle U \rangle = \overline{\langle U \rangle} + U_{z,d}$ and $\langle C \rangle = \overline{\langle C \rangle} + C_{z,d}$, where $U_{z,d}$ and $C_{z,d}$ are the deviation of the depth mean over the temporal mean. The tidal fluctuation of the cross-sectional area can be written as $A = \overline{A} + A_t$, where A_t is the deviation of the cross-sectional area from the

temporal mean, induced by the variation of tidal height. According to Dyer (1974), T1 represents the mean flow; T2 is a compensation flow for the landward transport on the partially progressive tidal wave; T3 is due to the correlation of tidal period variation of tidal height and salinity; T4 is the correlation of tidal period variations of salinity and current; T5 is the third-order correlation of tidal period variations in salinity, velocity, and tidal height; T6 is the mean shear effect; and T7 is the covariance of the shear effect and tidal height. T1 and T2 are associated with nontidal drift, T3–T5 can be regarded as the transport from the tidal pumping effect, and T6 and T7 indicate the vertical-gradient-induced shear effect. To be more accurate, the cross section E is divided into 151 subsections with the same width of about 100 m for each. The flux of each subsection is first decomposed and then summarized for terms T1–T7.

TABLE 1. Experiment settings and the half-life exchange time and residual tracer concentration in each experiment.

Expt	Tides	Winds	Waves	Wave–current interactions	$T_{1/2}$ (days)	Residual tracer
Exp1 (Control Run)	Y	Y	Y	Y	99	0.094
Exp2 (No Waves)	Y	Y	N	N	165	0.249
Exp3 (Tides Only)	Y	N	N	N	396	0.534
Exp4 (No Tides)	N	Y	Y	Y	21	0.063
Exp5 (No CWCBS)	Y	Y	Y	Y, but combined wave–current bottom stress	99	0.092
Exp6 (No Form drag)	Y	Y	Y	Y, but form drag	99	0.092
Exp7 (No WDinTKE)	Y	Y	Y	Y, but wave dissipation in turbulence kinetic energy equation	106	0.136
Exp8 (No WRS)	Y	Y	Y	Y, but depth-dependent wave radiation stress	126	0.157
Exp9 (No CARWE)	Y	Y	Y	Y, but current advection and refraction of wave energy	106	0.148

3. Results

a. Significant wave height

Figure 3a shows the distribution of 1-year-averaged H_s in Exp1 (Control Run). The isoline parallels the northeast–southwest direction with a greater H_s in the southeast. At the bay mouth, the H_s is about 0.4 m in the west and 0.6 m in the east. The middle channel is the main wave propagation path, through which waves can reach the top of the bay with an H_s of 0.2 m. The subbasins are well sheltered due to their entrances perpendicular to the wave propagation. Excluding the CARWE terms in Exp9 significantly changes the H_s distribution (Fig. 3b). The other wave–current interaction processes change the H_s below 0.02 m (not shown). The CARWE enhances the H_s in the eastern part of the bay, but reduces it in the western part, which might be determined by the current.

As a result of the monsoon in this region, waves vary seasonally. In January 2019, the averaged northerly wind is 3.8 m s^{-1} with a maximum of 6.4 m s^{-1} ; and in July 2019, the averaged southerly wind is 3.1 m s^{-1} with a maximum of 7.0 m s^{-1} (Fig. 2a). The northerly wind blows the surface wave offshore in winter; thus, H_s is relatively weaker than that in summer because of the limited fetch (Figs. 3c,d). The H_s is generally below 0.2 m inside DYB in winter but can reach 0.4 m at the top of the bay in summer.

Before Tropical Storm Wipha (from 25 to 29 July), the H_s has a similar pattern as the monthly-averaged H_s in July (Fig. 3e), but the magnitude is less than that because of the weak southerly wind (Fig. 2b). During the storm (from 30 July to 3 August), the H_s is largely increased to 3.0 m on the shelf, 2.2 m at the bay mouth, and 0.7 m at the top of the bay (Fig. 3f). Furthermore, the storm can also generate an H_s of 0.3 m in those subbasins.

b. Residual current

The residual current is calculated as 1-year-averaged and vertically averaged current for the first four experiments (Fig. 4). Note that in experiments including the wave–current interactions the current represents the ocean current plus Stokes drift; otherwise, the ocean current is solo. In Exp3 (Tides Only), the tide-driven flow is larger around islands and in shallow-water areas than that on the shelf (Fig. 4a), with a maximum of 0.243 m s^{-1} and an average of 0.010 m s^{-1} inside the bay. The difference between Exp2 (No Waves) and Exp3

illustrates that the wind-driven flow travels westward on the shelf (0.177 m s^{-1} in maximum and 0.010 m s^{-1} in average) with smaller speeds inside the bay (0.032 m s^{-1} in maximum and 0.003 m s^{-1} in average), which travels basically along the coasts (Fig. 4b). It forms two clockwise gyres, one at the northeastern part of the bay and the other on the west of the DNC at the bay mouth. The difference between Exp1 (Control Run) and Exp 2 indicates wave-driven flow, which reverses the wind-driven flow outside the bay (Fig. 4c). An anticlockwise gyre is induced on the east of the DNC near the bay mouth. It also generates strong offshore currents, one along the east coast from the FH harbor and another from the DP cove but separated by the DLJ island. The wave-driven flow can reach 0.376 m s^{-1} in maximum with an average of 0.008 m s^{-1} inside the bay and 0.478 m s^{-1} in maximum with an average of 0.017 m s^{-1} on the shelf. The BS cove and the FH harbor are little affected by waves, which is consistent to the H_s shown in Fig. 3a. Figure 4d shows that the residual current in Exp4 (No Tides) combines the wind- and wave-driven flow, reaching 0.354 m s^{-1} in maximum and 0.014 m s^{-1} in average inside the bay, and 0.485 m s^{-1} in maximum and 0.028 m s^{-1} in average on the shelf.

c. Comparison between different forcings

The spatial-averaged tracer concentration in the bay is calculated as

$$\hat{C}(t) = \frac{1}{S} \int_0^S \int_{-h}^{\eta} C(n, s, t) ds dS,$$

where C is the depth-dependent tracer concentration at each triangle element n and S is the total water area of the bay (north of the cross section E in Fig. 1). The half-life exchange time $T_{1/2}$ is given as

$$\hat{C}(T_{1/2}) = 0.5.$$

Figure 5a shows the daily-averaged $\hat{C}$ between the first four experiments. Driving by tides only (Exp3), the tracer concentration inside the bay reduces to about 0.534 after 1 year of simulation (Table 1). In Exp2 (No Waves), winds enhance the bay–shelf exchange compared to that in Exp3. The tracer concentration in DYB lowers to 0.249 at the end of simulation and the $T_{1/2}$ reduces from 396 to 165 days. When the wave–current interactions are active (Exp1), the tracer concentration further lowers to less than 0.094 and $T_{1/2}$ shortens to 99 days.

FIG. 3. The 1-year-averaged significant wave height (m) in (a) Exp1 (Control Run) and (b) Exp1 - Exp9 (No CARWE), the monthly-averaged significant wave height in (c) January and (d) July in Exp1, and the 5-day-averaged significant wave height (e) before Tropical Storm Wipha (from 25 to 29 Jul) and (f) during the storm (from 30 Jul to 3 Aug) in Exp1. Note that the color scale differs from panel to panel.

FIG. 4. The vertically averaged residual current (m s^{-1}) in (a) Exp3 (Tide Only), (b) Exp2 (No Waves) – Exp3, (c) Exp1 (Control Run) – Exp2, and (d) Exp4 (No Tides). The arrows show the direction, and the color shows the common logarithm of the velocity magnitude or the absolute difference.

During Tropical Storm Wipha, there is a fast drop of the tracer concentration in Exp2, and a more dramatic drop in Exp1. It indicates that during the extreme weather conditions, the half-life exchange time is significantly shortened by the intensified bay–shelf exchange, in which waves contribute more than winds. Removing tides from Exp4, the remaining tracer concentration is 0.063 at the end of the 1-year simulation, and it only needs 21 days to reach $T_{1/2}$. It indicates the tidal current hinders the bay–shelf exchange induced by winds/waves. H. Zhang et al. (2019) also mention that “[t]ides increase the residence time with evident seasonal differences,” but without an explanation.

d. Comparison between different wave–current interactions

The calculated $T_{1/2}$ and the residual tracer concentration after 1 year of simulation in Exp5–Exp9 is also listed in Table 1.

Figure 5b shows the decay of daily-averaged tracer concentration in these experiments. In Exp5 (No CWCBS) and Exp6 (No Form drag), the bay–shelf exchange changes little and the decay of $\hat{C}$ overlaps that in the control run (thus not shown in Fig. 5b). If not considering the wave dissipation in TKE equation, $T_{1/2}$ is extended to 106 days and the residual tracer concentration is increased to 0.136 in Exp7 (No WDinTKE). The result of Exp8 (No WRS) shows that $T_{1/2}$ is extended to 126 days and the residual tracer concentration is increased to 0.157. Exp9 (No CARWE) extends $T_{1/2}$ to 106 days and the left tracer concentration to 0.148. In the view of the half-life exchange time, the wave radiation stress (Exp8) is the most important wave–current interaction, followed by the CARWE (Exp9) and the wave dissipation on turbulence (Exp7). The CWCBS (Exp5) and the form drag (Exp6) have little effects on the bay–shelf exchange.

FIG. 5. The residual tracer concentration inside the bay in each experiment [Exp1 is repeated in (b) for ease of comparison].

4. Discussion

a. Counteraction between tides and winds/waves

The temporal-averaged current is calculated along cross section E (Fig. 6). In Exp3 (Tides Only), the residual current shows a sandwich pattern from the west coast to the DNC (Fig. 6a); the alternate seaward and landward flow in the eastern of the cross section indicates a recirculation around the HM island (Fig. 4a). Comparison between Exp2 (No Waves) and Exp3 can find that winds drive the surface flow landward and the compensation flow seaward below the surface (cf. Figs. 6a,b), indicating that in average the southerly wind dominates the year. Comparison between Exp1 (Control Run) and Exp2 indicates waves enhance the landward flow on the surface and flanks of each channel (cf. Figs. 6b,c). Excluding tides in Exp4, the residual current shows a two-layer structure, i.e., the landward flow in the upper layer and seaward flow in the lower layer (Fig. 6d). The landward flow is large in the shallow water and the seaward flow is greatest in the middle channel. Comparing Figs. 6a and 6d, it can be found that winds/waves generate opposite current to tides in most of the cross section. As the model is unstratified, the tracer concentration is nearly vertically uniform. The distribution of tracer flux at the bay mouth is mainly determined by the residual current (thus not shown). Tides basically generate barotropic flows; and winds and waves induce two-layer flows in the wide west channel and middle channel. So do the fluxes.

The decomposed cross-sectional flux for section E is listed in Table 2. The results show that the residual current induces a landward flux, but the other terms induce seaward fluxes. The shear effect dominates the seaward fluxes. The residual

tidal-current induced transport is comparable to the shear-induced transport only in Exp3 (Tides Only). It indicates tides can generate a relative stronger landward residual current but a weaker shear-induced seaward current. Figures 6e–h shows the 1-year-averaged vertical shear of current S_h along cross section E, which is calculated as

$$S_h = \sqrt{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2}.$$

The vertical shear in velocity is induced by bottom friction and bathymetry gradients in Exp3 (Fig. 6e). However, winds can enhance the vertical shear, especially on the surface (Fig. 6f). The shear-induced flux (T6 + T7) in Exp2 (No Waves) is also enlarged. Waves further enhance the vertical shear by about one order (Fig. 6g) and the shear-induced flux in Exp1 (Control Run) is more than 2 times that in Exp3 (Table 2). Waves can increase the pumping effect (T3 + T4 + T5) by one-third relative to that in Exp3. Meanwhile, the residual-current-induced flux (T1) is reduced by about one-half in both Exp1 and Exp2, which indicates a seaward flow induced by winds and waves (Figs. 4b,c). For Exp4 (No Tides), the shear-induced flux is about 3 times that in Exp3. A comparison of Figs. 6g and 6h shows the vertical shear in Exp4 is less than that in Exp1. However, why is T6 + T7 larger in Exp4 than in Exp1? This is due to a larger vertical gradient of tracer concentration ($\partial C/\partial z$) in Exp4. The RMS of $\partial C/\partial z$ for Exp1–Exp4 are 1.8×10^{-4} , 1.9×10^{-4} , 1.9×10^{-3} , and 3.0×10^{-3} (m^{-1}), respectively. It indicates a more inhomogeneous distribution of $\partial C/\partial z$ in Exp4 than in Exp1 because of the removal of tidal mixing, which benefits for the shear-induced transport. From Table 2, we can

FIG. 6. (left) The residual current (m s^{-1}) in (a) Exp3 (Tide Only), (b) Exp2 (No Waves), (c) Exp1 (Control Run), and (d) Exp4 (No Tides). The thin black line indicates zero, and the thick black line indicates the seabed. Positive indicates seaward. (right) the corresponding common logarithm of the current shear S_h (s^{-1}) for (e) Exp3, (f) Exp2, (g) Exp1, and (h) Exp4.

also find that without tidal forcings in Exp4 the T2 term is almost zero and the pumping effect is reduced to half, while the wind- and wave-driven flows turn landward. This can also be found in Fig. 4d, where the northeastward current between the SM island and mainland may affect the transport. Furthermore, the northward current between the HS cape and HM island is enhanced in this experiment compared to Figs. 4a–c.

Comparison between Exp1 and Exp4 (Table 2) shows that tides can enhance the landward residual-current-induced flux (T1), but largely reduce the shear effect (T6 + T7). However, comparison between Exp1 and Exp3 shows winds/waves can largely reduce T1 but enhance T6 + T7. Thus, in DYB, tides work opposite to winds/waves on the bay–shelf exchange. Winds/waves can enhance the bay–shelf exchange induced by tides, but tides can slow the bay–shelf exchange induced by winds/waves.

b. Wave–current interactions on bay–shelf exchange

1) WAVE DISSIPATION ON TURBULENCE

The wave dissipation is a source term in the TKE equation but a sink term in the wave energy equation [Eq. (1)], which is removed from the TKE equation in Exp7. Figure 7a shows the K_m along the cross section E in the control run (Exp1), in which the K_m is relatively large in the middle layer and small in the subsurface and bottom layers. The bathymetry gradient enlarges the K_m . Figure 7b shows the difference of K_m between Exp1 and Exp7 (No WDinTKE), indicating the wave dissipation enhances the vertical mixing only on the surface but reduces it elsewhere. This is consistent with the result of Gao et al. (2018).

Including all wave–current interactions (Exp1–Exp2), the residual current shows a three-layer structure from the west coast to the DNC, but is basically vertical-uniform around the

TABLE 2. Decomposition of cross-sectional flux in each experiment.

Expt	Cross-sectional flux ($\text{m}^3 \text{s}^{-1}$)				
	T0	T1	T2	T3 + T4 + T5	T6 + T7
Exp1 (Control Run)	91.73	−23.97	2.95	11.53	101.22
Exp2 (No Waves)	72.37	−25.39	2.21	9.14	86.40
Exp3 (Tides Only)	7.67	−53.85	2.78	9.05	49.69
Exp4 (No Tides)	126.34	−19.81	0.02	4.70	141.44
Exp5 (No CWCBS)	92.26	−18.86	2.94	11.72	96.46
Exp6 (No Form drag)	92.16	−16.13	3.07	9.59	95.64
Exp7 (No WDinTKE)	85.08	−21.51	2.44	15.69	88.47
Exp8 (No WRS)	72.19	−14.30	5.48	6.71	74.31
Exp9 (No CARWE)	78.47	−0.45	6.32	11.69	60.91

FIG. 7. The vertical eddy viscosity K_m ($\times 10^{-3} \text{ m}^2 \text{ s}^{-1}$) in (a) Exp1 and the difference of (b) Exp1 – Exp7 (No WDinTKE), (c) Exp1 – Exp8 (No WRS), (d) Exp1 – Exp9 (No CARWE), (e) Exp1 – Exp5 (No CWCBS), and (f) Exp1 – Exp6 (No Form drag). The thick black line indicates the seabed. Note that the color scale differs in (a).

HM island and in the east channel (Fig. 8a). The wave dissipation on turbulence enhances the seaward flow in the middle of the cross section but reduces the seaward flow in the east channel (Fig. 8b). As the wave dissipation enhances the vertical shear of K_m (Fig. 7b), when neglecting it in Exp7, both the flow anomaly and the tracer flux anomaly in Figs. 8b and 8f (Exp1–Exp7) are more depth uniform than that in Figs. 8a and 8e (Exp1–Exp2).

Decomposition of the flux in the cross section E (Table 2) illustrates that without the wave dissipation in the TKE equation in Exp7, the residual-current-induced landward transport (T1) is reduced by 10.3% and the shear-induced transport (T6 + T7) is reduced by 12.6%. Although the pumping-effect-induced transport (T3 + T4 + T5) is increased by 36.1%, the total cross-sectional flux is reduced toward the sea. Thus, the bay–shelf exchange is slowed.

2) WAVE RADIATION STRESS

Wave radiation stress (WRS) is important to the wave–current interactions (Longuet-Higgins and Stewart 1964), which is modeled as the divergence of a stress tensor (Lane et al. 2007). Both the half-life exchange time (Table 1) and the cross-sectional flux (Table 2) indicate the WRS plays the most important role in the bay–shelf exchange among the proposed wave–current interactions. Previous work by Gao et al. (2018) also found the WRS may alter the current velocity significantly, even though the three-dimensional WRS proposed by Mellor (2003, 2015) missed the divergence of vertical flux associated with nonhydrostatic pressure perturbations (Ardhuin et al. 2017). In this study, the residual current between Exp1 and Exp8 (No WRS) also shows the most significant difference (Fig. 8c), which is almost equal to the difference between Exp1 and Exp2 (Fig. 8a). The vertical mixing is mostly little affected by the WRS but changed a lot in some spots (Fig. 7c). A comparison of the difference of the current (Fig. 8c) and the flux (Fig. 8g) between Exp1 and Exp8 indicates that the contribution of WRS

on the bay–shelf exchange is mostly due to its effect on the current rather than on the vertical mixing.

Decomposition of the cross-sectional flux (Table 2) also shows that in Exp8 the residual-current-induced landward transport (T1) is reduced by 40.3%, but the shear-induced transport (T6 + T7) is reduced by 26.6% compared to that in Exp1. The former is affected by the current and the latter is affected by both the current and the vertical mixing. The other two terms are changed relatively less than these two. Also, the bay–shelf exchange is slowed due to the decrease of the net seaward transport.

3) CURRENT ADVECTION AND REFRACTION OF WAVE ENERGY

The second and third terms on the left-hand side of Eq. (1) represent the CARWE. Both terms are the feedback from currents to waves; thus, in this study we combine them together for discussion. The CARWE can enhance the vertical mixing on the surface layer, but the K_m is reduced near the west coast and the middle to bottom layers in the three channels (Fig. 7d). The CARWE enhances the seaward flow in the central part of the three channels, but the landward flow elsewhere (Fig. 8d). Basically, the CARWE does not change the direction of the residual flow except near the bottom of the middle section. Comparison of the mean flux between Exp1 and Exp9 (no CARWE) illustrates that the CARWE can drive strong seaward flux on the surface, which extends to the bottom in the west channel and the middle channel (Fig. 8h). Compared to Fig. 8d, the seaward transport in the east channel is reduced, as the CARWE generates a landward flow in this channel.

Decomposition of the cross-sectional flux (Table 2) shows that, excluding the CARWE terms, the landward residual-current-induced transport (T1) almost vanishes and the shear-induced transport (T6 + T7) is reduced by 39.8%. As compared with Exp7 (No WDinTKE) and Exp8 (No WRS), in Exp9 the CARWE alters T1 and T6 + T7 more than the other two experiments. However, the reduced landward transport is

FIG. 8. (left) The residual current (m s^{-1}) and (right) tracer flux ($\text{m}^3 \text{s}^{-1}$) in (a),(e) Exp1 – Exp2 (No Waves); (b),(f) Exp1 – Exp7 (No WDinTKE); (c),(g) Exp1 – Exp8 (No WRS); and (d),(h) Exp1 – Exp9 (No CARWE). The thin black line indicates zero, and the thick black line indicates the seabed. Positive indicates seaward. Note that the color scale differs between (a)–(d) and (e)–(h).

balanced by the reduced seaward transport; thus, the net cross-sectional flux is between that of Exp7 and Exp8 and the bay–shelf exchange is correspondingly slowed.

4) COMBINED WAVE–CURRENT BOTTOM STRESS AND FORM DRAG

The bottom stress induced by the wave orbital velocities and its interaction with the current-induced bottom stress were addressed to be significant to the sediment suspension and transport (e.g., Tang and Grimshaw 1996; Wiberg and Sherwood 2008; Song and Wang 2013; Gao et al. 2018). However, its impact on the bay–shelf exchange is insignificant. The residual current and tracer transport induced by the CWCBS is one or two order less than the other wave–current interactions (not shown). It also has little impact on the vertical mixing except the west coast (Fig. 7e).

The wind stress working on the sea surface can be divided into a turbulent viscosity (skin friction) and a horizontal pressure gradient (form drag) due to the variation of wave-induced water elevation (Mellor et al. 2008). The form drag competes with the turbulence-supported stress in the momentum equations and the vertical turbulent closure equations (Mellor 2005). Thus, the form drag enhances the vertical mixing throughout the water column (Fig. 7f). Similar as the CWCBS, the form drag also generates negligible residual current (not shown). But different from the CWCBS, the form drag has a larger impact on the tracer transport due to its effect on the vertical mixing, which drives the seaward transport in the lower layer but landward flux in the upper layer. However, the net transport is almost balanced by the two-layer transport. Therefore, the form drag has little effect on the bay–shelf exchange.

The decomposed cross-sectional flux also confirms that the CWCBS and the form drag play minor roles in the bay–shelf exchange (Table 2). In both Exp5 (No CWCBS) and Exp6 (No Form drag), all terms vary less than the other wave–current interactions. In Exp6, due to the removal of form drag, the terms alter more than those in Exp5, but they together change little on the net transport, which is nearly the same as that in Exp5.

c. Seasonal variation of wave–current interactions

Figure 9a shows the difference between the monthly-averaged K_m in January and July in Exp1–Exp2. It illustrates that the surface K_m is dramatically increased in summer, when the southerly wind can generate a larger H_s in DYB (cf. Figs. 3c,d) as the bay is facing southeast. Thus, the upper-layer mixing is enhanced in summer. The wave dissipation on turbulence leads to the reduction of K_m on the surface and the increase of K_m beneath the surface during winter (Fig. 9b). The WRS reduces the K_m throughout the water column (Fig. 9c). The CARWE leads to almost the same seasonal variation in the cross section (Fig. 9d) as that shown in Fig. 9a. The form drag increases the K_m in the middle layer, but the magnitude is much less than that in the other experiments (not shown). The CWCBS also has little effect on the seasonal variation of the vertical mixing (not shown). Note that in these experiments, the model runs with constant temperature and salinity, and thus the seasonal variation of K_m is only caused by waves in different seasons.

Compared to summer, the landward current in winter is greater in the central west channel and central middle channel; the seaward current is greater in shallow waters and the around island circulation is enhanced (Fig. 9e). The wave dissipation

FIG. 9. The seasonal difference (winter – summer) of (left) K_m ($\times 10^{-3} \text{ m}^2 \text{ s}^{-1}$) and (right) residual current (m s^{-1} ; positive indicates seaward) in (a),(e) Exp1 (Control Run) – Exp2 (No Waves); (b),(f) Exp1 – Exp7 (No WDinTKE); (c),(g) Exp1 – Exp8 (No WRS); and (d),(f) Exp1 – Exp9 (No CARWE). The thin black line indicates zero, and the thick black line indicates the seabed. Note that the color scale differs between (a)–(d) and (e)–(h).

on turbulence leads to the increase of landward current in the west channel and middle channel (Fig. 9f). Figures 9c and 9d show almost the same patterns, which indicates the WRS and the CARWE have the same effect on the seasonal difference of residual current. Both the CWCBS and the form drag have little impact on the seasonal variation (not shown), one order less than other wave–current interactions.

To quantify the contribution, the RMS is calculated between the seasonal variation in Exp1 and the seasonal variation induced by each wave–current interaction (Table 3). It indicates the CARWE contributes most to the seasonal variation of the residual flow, followed by the WRS. It also contributes most to the seasonal variation of vertical mixing, but followed by the wave dissipation on turbulence.

d. Synoptic variation of wave–current interactions

As shown in Fig. 5, Tropical Storm Wipha dramatically enhances the bay–shelf exchange in DYB. To quantify the performance of each wave–current interaction, the K_m and residual flow before the storm (from 25 to 29 July) and during

the storm (from 30 July to 3 August) are calculated and compared in Fig. 10. The tropical storm significantly enhances the mixing throughout the water column (not shown), but its induced waves can only enhance the mixing on the surface (Fig. 10a), in which the wave dissipation on turbulence (Fig. 10b) and the CARWE (Fig. 10d) contributes most. The WRS can only enhance the mixing around the HM island and near the coasts, and slightly reduce the mixing elsewhere (Fig. 10c). The form drag can enhance the mixing in most of the cross section and only reduce it in the upper layer of the east channel (Fig. 10e). Compared to the other processes, the CWCBS has little effect on the vertical mixing (not shown). Obviously, the summation of the above terms is unequal to the total wave effect on the vertical mixing, which indicates they may work together nonlinearly.

The difference between Exp1 and Exp2 shows the storm-induced wave can increase the seaward flow in the central west channel, the landward flow near the west coast and the eastern cross section (Fig. 10f). The wave dissipation on turbulence (Fig. 10g), the WRS (Fig. 10h) and the CARWE (Fig. 10i) can

TABLE 3. The root-mean-square of seasonal/synoptic variations on residual flow and vertical mixing between Exp1 and the other experiments.

Expt	Seasonal		Synoptic	
	Residual flow	Vertical mixing	Residual flow	Vertical mixing
Exp5 (No CWCBS)	0.0230	0.001 58	0.1250	0.0124
Exp6 (No Form drag)	0.0223	0.001 60	0.1311	0.0120
Exp7 (No WDinTKE)	0.0206	0.000 96	0.0853	0.0102
Exp8 (No WRS)	0.0110	0.001 42	0.0362	0.0123
Exp9 (No CARWE)	0.0088	0.000 33	0.0278	0.0100

FIG. 10. The synoptic difference (during the typhoon – before the typhoon) of (left) K_m ($\times 10^{-3} \text{ m}^2 \text{ s}^{-1}$) and (right) residual current (m s^{-1} ; positive indicates seaward) in (a),(f) Exp1 – Exp2 (No Waves); (b),(g) Exp1 – Exp7 (No WDinTKE); (c),(h) Exp1 – Exp8 (No WRS); (d),(i) Exp1 – Exp9 (No CARWE); and (e),(j) Exp1 – Exp6 (No Form drag). The thin black line indicates zero, and the thick black line indicates the seabed. Note that the color scale differs between (a)–(e) and (f)–(j).

drive a similar flow pattern with that shown in Fig. 10f. It indicates that during the storm the three processes may have a similar effect on the current, but their interactions are also nonlinear. The residual flows generated by the form drag (Fig. 10j) and the CWCBS (not shown) during the storm have much smaller magnitudes than those by the other processes.

To quantify their contribution, the RMS is also calculated between the synoptic variation in Exp1 and that induced by each wave–current interaction (Table 3). It indicates the CARWE also contributes most to both the synoptic variation of the residual flow and the vertical mixing with the former followed by the WRS and the latter followed by the wave dissipation on turbulence.

5. Conclusions

In this paper, the effects of wave–current interactions on the bay–shelf exchange are studied in a micro- to mesotidal system. Based on the tracer-tracking experiments, the half-life exchange time is calculated and the performance of five wave–current interactions is evaluated. Comparison between the experiments driving by different forcings (tides, winds and waves) shows a counteraction on the bay–shelf exchange between tides and winds/waves. Winds/waves can enlarge the shear-induced seaward transport and reduce the

residual-current-induced landward transport, which benefits the bay–shelf exchange or the pollution dilution. Oppositely, tides can increase the residual-current-induced landward transport and reduce the shear-induced seaward transport. This is because the tidal mixing reduces the vertical gradient of pollutant concentration, which is disadvantage to the shear-induced seaward transport and further the bay–shelf exchange.

Comparison between the experiments driving by different wave–current interaction processes shows the WRS dominates the bay–shelf exchange due to its change on the residual current. It is followed by the wave dissipation on turbulence and the CARWE, which alter both the residual current and the vertical mixing. The form drag enhances the vertical mixing throughout the water column but has little impact on the residual current. It forms a two-layer transport, which is balanced between the surface landward transport and the bottom seaward transport. Thus, it has little contribution to the bay–shelf exchange. The CWCBS has little effect on both the residual current and the vertical mixing. Therefore, the importance of wave–current interactions to the bay–shelf exchange ranks from the most to the least as the WRS, the wave dissipation on turbulence, the CARWE, the form drag, and the CWCBS.

The southerly wind can generate a larger H_s in DYB than that by the northerly wind. Thus, the bay–shelf exchange is

generally faster in summer than that in winter. The seasonal variation of wave–current interactions is dominated by the CARWE terms as they can significantly change the distribution of the H_s , which is sensitive to the wind direction. This may also be the reason for the dominance of the CARWE terms in the synoptic variation of wave–current interactions. The WRS contributes second to both the seasonal and synoptic variations on the flow velocity, while the wave dissipation on turbulence contributes second to both variations on the vertical mixing. The form drag and the CWCBS still have little impact on the variations. Although the CWCBS plays the most important role in suspended sediment transport (Gao et al. 2018), it plays an insignificant role in the bay–shelf exchange. It indicates that waves are important to several ocean processes, but the dominant wave–current interaction mechanism may be different.

In this study, we also examined the wave–current interactions on the horizontal eddy viscosity. However, it only varies more in coasts with a great depth gradient. The horizontal eddy viscosity usually varies consistently (not in magnitude but in direction) through the entire cross section. The wave dissipation on turbulence, the WRS and the CARWE can increase the horizontal eddy viscosity, while the form drag and the CWCBS can decrease it. The southerly wind can enhance the horizontal mixing. In general, the change on horizontal eddy viscosity is not as notable as the residual current and the vertical eddy viscosity.

The work by H. Zhang et al. (2019) found that the large-scale coastal circulation and local winds are the main drivers of the upper-layer residence time in DYB. However, we find waves have more important effects on the bay–shelf exchange than winds in this study. The effect of stratification on wave-driven bay–shelf exchange is still unclear (e.g., Lentz and Fewings 2012). Comparison between the large-scale coastal circulation, stratification, and waves will be explored in future, because it needs a well-validated baroclinic model.

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Data availability statement. The data used in this study can be accessed online (<https://doi.org/10.6084/m9.figshare.12916043.v2>).

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