Investigation of the Internal Tides in the Northwest Pacific Ocean Considering the Background Circulation and Stratification

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(Manuscript received 1 August 2019, in final form 26 July 2020)

ABSTRACT: A global ocean circulation and tide model with nonuniform resolution is used in this work to resolve the ocean circulation globally as well as mesoscale eddies and internal tides regionally. Focusing on the northwest Pacific Ocean (NWP, 0° – 35° N, 105° – 150° E), a realistic experiment is conducted to simulate internal tides considering the background circulation and stratification. To investigate the influence of a background field on the generation and propagation of internal tides, idealized cases with horizontally homogeneous stratification and zero surface fluxes are also implemented for comparison. By comparing the realistic cases with idealized ones, the astronomical tidal forcing is found to be the dominant factor influencing the internal tide conversion rate magnitude, whereas the stratification acts as a secondary factor. However, stratification deviations in different areas can lead to an error exceeding 30% in the local internal tide energy conversion rate, indicating the necessity of a realistic stratification setting for simulating the entire NWP. The background shear is found to refract propagating diurnal internal tides by changing the effective Coriolis frequencies and phase speeds, while the Doppler-shifting effect is remarkable for introducing biases to semidiurnal results. In addition, nonlinear baroclinic tide energy equations considering the background circulation and stratification are derived and diagnosed in this work. The mean flow–baroclinic tide interaction and nonlinear energy flux are the most significant nonlinear terms in the derived equations, and nonlinearity is estimated to contribute approximately 5% of the total internal tide energy in the greater Luzon Strait area.

KEYWORDS: North Pacific Ocean; Energy transport; Internal waves; Ocean dynamics; Tides; General circulation models

1. Introduction

Internal waves exist in stratified oceans; the frequencies of these internal waves are larger than the inertial frequencies and lower than the buoyancy frequencies. As a kind of internal wave, internal tides are ubiquitous phenomena generated by barotropic (BT) tides flowing over varying topographies. Internal tides feature multimodal vertical structures: low-mode internal tides can propagate over long distances and dissipate in the far field, while high-mode internal tides usually dissipate locally around their generation sites. The different propagative behaviors of low-mode and high-mode internal tides affect the distribution of global internal tide energy (Alford 2003). In addition, internal tides can lead to diapycnal mixing, which is believed to maintain global stratification as well as thermohaline circulation (Munk and Wunsch 1998). As a result, internal tides and near-inertial waves are considered to be one of the most important schemes in diapycnal mixing, as well as important components in the energy cascade of the global ocean (Alford and Gregg 2001).

The global map of internal tide energy in Niwa and Hibiya (2011) shows that the northwest Pacific Ocean (NWP, which ranges from 0° to 35° N and from 105° to 150° E) is one of the most energetic regions with regard to both diurnal and semidiurnal baroclinic (BC) tides. With double-ridge topography and strong BT tides, the Luzon Strait (LS) is the most

energetic source of internal tides and internal solitary waves throughout the entire NWP (Alford et al. 2015; Guo and Chen 2014). After being generated within the LS, internal tides distinctly propagate westward into the South China Sea (SCS) and eastward into the Philippine Sea (PS) (Jan et al. 2008). Flat and deep basins locate at the center of the SCS and PS, surrounded by shelf-slope areas, seamounts, and ridges (see Fig. 1c). Thus, strong internal tides generated from the LS show long-range propagation in the SCS and PS (Zhao 2014; Xu et al. 2016), along with interferences with locally generated internal tides (Niwa and Hibiya 2004; Kerry et al. 2013; Wang et al. 2018).

Nonstationary internal tides have been widely revealed through altimetry and mooring data in areas of interest for internal tides, such as the LS and Hawaiian Ridge (Chavanne et al. 2010; Ray and Zaron 2011; Xu et al. 2014; Pickering et al. 2015; Zaron 2017; Huang et al. 2018). Shriver et al. (2014) and Savage et al. (2017) estimated nonstationary internal tides globally from the output of the Hybrid Coordinate Ocean Model (HYCOM); their results assert that nonstationary internal tides are ubiquitous in the global ocean. To evaluate nonstationarity, variables of internal tides can be decomposed into stationary and nonstationary parts through mathematical methods. The stationary part of internal tides is periodic and phase-locked and is forced by the BT tidal potential. In contrast, according to Pickering et al. (2015), the nonstationary part can be explained by “local” and “remote” mechanisms; the remote mechanism corresponds to the interference attributable to internal tides from different generation sites (Kerry et al. 2013), while the local mechanism constitutes the effect of the background field, such as subtidal circulation and

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DOI: 10.1175/JPO-D-19-0177.1

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FIG. 1. (a) Global orthogonal curvilinear mesh grid of the MPI-OM designed for this study. Scattered points represent the grid points with an interval of 15 points. Colors indicate the model resolution, which varies from 3.5 to 135.5 km globally. (b) Global mesh zoomed in on the NWP area. (c) Model topography of the NWP area, where the colors indicate the water depth.

stratification (Zilberman et al. 2011). To consider the effect of the background field, the assumption of horizontally homogeneous stratification, which has been widely used in previous work, cannot be employed. By using a quasi-three-dimensional model considering tilted isopycnals as well as an idealized northward Kuroshio, Jan et al. (2012) evaluated the effects of the Kuroshio on the generation and propagation of internal tides in the LS. Furthermore, Dunphy and Lamb (2014) simulated the propagation of a mode-1 internal tide through an eddy in a rectangular flat domain and noted that mesoscale eddies may lead to energy conversion between different vertical modes. Ponte and Klein (2015) and Dunphy et al. (2017) simulated the propagation of low-mode internal tides through turbulent fields and concluded that mesoscale turbulent fields may lead to nonstationary internal tides as well as the refraction of the propagation paths. Via a modified linear internal wave equation considering a 2D geostrophic front, Li et al. (2019) concluded that geostrophic fronts can cause refraction/reflection of internal waves and energy transmission to high vertical modes. In addition, several studies have been conducted based on the topography of the real ocean. Zaron and Egbert (2014) analyzed assimilated model results and reported that nonstationarity mainly originates from perturbations to the phase speed of BC tides rather than generation site processes. Kelly et al. (2016) investigated the propagation of

internal tides in the greater Mid-Atlantic Bight region with a coupled-mode, shallow-water model and found that mode-1 internal tides can be refracted or reflected by the Gulf Stream, emerging as anomalous energy fluxes. Combined with a derived nonlinear vertical-mode momentum and energy equation of internal tides that considers a sheared mean flow and a horizontally nonuniform density field, Kelly and Lermusiaux (2016) analyzed the energy balance of each process with assimilated model results and concluded that the nonlinear advection of energy flux can explain most of the tide-mean flow interaction. Kerry et al. (2014a,b, 2016) systematically investigated M_2 internal tides in the PS considering the effects of subtidal circulation with a nested Regional Ocean Modeling System (ROMS), including the impacts of subtidal circulation on the generation, propagation and mixing of internal tides. Using an assimilated global circulation model, Varlamov et al. (2015) noted that the M_2 internal tides at four generation areas in the NWP are modulated considerably by low-frequency changes in the density field, including the variation of the Kuroshio, mesoscale eddy activities and seasonal variation of the thermocline. By conducting sensitivity runs of ROMS, Chang et al. (2019) found that the Kuroshio northeast of Taiwan alters the conversion rates at the I-Lan Ridge and Mien-Hua Canyon, as well as the propagation patterns nearby.

The internal tides in the NWP, especially those in the northeast SCS, have been heavily researched for decades. Many numerical investigations of internal tides have been performed under an idealized condition where the background stratification is horizontally homogeneous and the background circulation is ignored; this approach leads to a convincing assumption for most studies. However, studies of internal tides encounter limitations when the theoretical background conditions change from simple to arbitrary and the investigation focus shifts from a single oceanic phenomenon to multiple oceanic phenomena. The NWP area features multiple oceanic processes, such as internal tides, the Kuroshio and mesoscale eddies; consequently, a simple linear theory of internal tides cannot provide in-depth insights into this region. Here, we raise two scientific questions considering the real background field in internal tide studies:

- 1) With the tide energy equation under linear theory, when considering a background field that includes realistic three-dimensional stratification as well as subtidal circulation, what differences arise in the generation and propagation of internal tides compared with those under idealized conditions?
- 2) If a background field that includes subtidal circulation and realistic stratification is considered, what nonlinear effects does the background field introduce compared to linear theory? How significant are those effects quantitatively?

To answer these questions, in this work, a global ocean circulation and tide model with a curvilinear orthogonal mesh grid is employed to simulate internal tides and global circulation. The highest resolution of the mesh grid is set in the NWP to resolve internal tides as well as eddies. Sensitivity tests are carried out simultaneously with idealized settings for comparison. Furthermore, nonlinear internal tide energy equations considering background fields are derived and presented. Additionally, the nonlinear energy terms are diagnosed to quantitatively evaluate the nonlinear effect of the background field. In the second section, we mainly introduce the applied numerical model and model configuration, and we further verify the numerical model. The third section mainly describes the linear and nonlinear internal tide energy equations as well as the data processing method. In the fourth section, first, we present the results of an idealized case both to verify our model and to give a preliminary overview of the internal tides in the NWP; second, a comparison is conducted among two sets of idealized experiments and a realistic experiment to answer the first question listed above; third, with the derived nonlinear energy equations of internal tides and the result of the realistic experiment, the energetic effect of the background field is evaluated quantitatively. The final section presents our summary and conclusion.

2. Model configuration and validation

The Max Planck Institute ocean model (MPI-OM), a global ocean circulation and tide model based on the ocean primitive equations, is employed in this study (Marsland et al. 2003; Chen et al. 2005). The MPI-OM is a Z-coordinate global ocean–sea ice model with an orthogonal curvilinear C-grid and

is developed from the Hamburg Ocean Primitive Equation (HOPE) model (Wolff et al. 1997). The primitive equations of the MPI-OM based on hydrostatic and Boussinesq approximations are listed below:

$$\left\{ \begin{array}{l} \frac{d\mathbf{u}_h}{dt} + f(\hat{k} \times \mathbf{u}_h) = -\frac{1}{\rho_c} \nabla_h(p + \rho_c g \eta + \rho_c \Omega) + \mathbf{F}_h + \mathbf{F}_v \\ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \\ \frac{\partial \eta}{\partial t} = -\nabla_h \cdot \int_{-H}^{\eta} \mathbf{u}_h dz \\ 0 = -\frac{\partial p}{\partial z} - \rho g \\ \frac{d\theta}{dt} = \nabla_h \cdot (\mathbf{K}_\theta \nabla_h \theta) \\ \frac{dS}{dt} = \nabla_h \cdot (\mathbf{K}_S \nabla_h S) \end{array} \right. \quad (1)$$

Equation set (1) contains the horizontal momentum equation, continuity equation, ocean surface elevation equation, hydrostatic pressure equation, and potential temperature and salinity diffusion equations. In these equations, $\mathbf{u}_h = (u, v)$, ρ_c represents a constant reference density, and Ω represents the tidal potential. The terms $\mathbf{F}_h$ and $\mathbf{F}_v$ denote the effects of horizontal and vertical eddy viscosities, respectively; the horizontal eddy viscosity is parameterized using a scale-dependent biharmonic formulation, while the vertical eddy viscosity term is expressed through the Pacanowski and Philander (PP) scheme (Pacanowski and Philander 1981), which depends on the Richardson number and constant coefficients. An additional parameterization for the wind-induced stirring is considered to make up for the underestimation of the turbulent mixing near the surface in the PP scheme. Subgrid eddy-induced mixing is parameterized by the Gent and McWilliams scheme (Gent et al. 1995). To resolve slope convection transports in the bottom boundary layer, (i.e., dense water overflow across sills between ocean basins), a modified Legutke and Maier–Reimer scheme (Legutke and Maier–Reimer 2002) is applied in MPI-OM. The global tide in the MPI-OM is forced by the lunar and solar tidal potential by calculating the distances, ascensions and declinations of the moon and the Sun at each time step; thus, the tide module in the MPI-OM considers all tidal constituents implicitly, which is different from regional models forced by several tidal components at open boundaries. The effect of solid Earth tides on ocean tides is expressed linearly as a portion of the calculated tidal forcing and is generally set to 0.69 following Kantha (1995). Following Thomas et al. (2001), the self-attraction and loading (SAL) effect is also expressed as a proportion of the local elevation and is set to 0.085.

There are mainly three reasons why we apply the global astronomical tide module. 1) The astronomical tide module calculates the position of Earth, the moon, and the Sun. That considers all tidal signals implicitly and is close to the real ocean tides. 2) The global tide module provides body force at each model grid. The in situ tidal forcing at each model grid

FIG. 2. Cotidal charts derived from the (top) MPI-OM and (bottom) TPXO8. Colors reflect the amplitude (m), while white lines denote the phase lag. Four main tidal constituents are shown in this figure.

makes our simulation more robust in physics compared to boundary-forcing regional models. 3) In a regional model, tidal movements may dissipate too much with a long traveling distance from the forced boundary. That may cause a damped tidal amplitude in the inner model domain, especially if the area of interest is as large as the NWP. However, the body-forced tide module avoids this problem.

To focus on the NWP area and to balance the computing resources and model resolution in this work, a bipolar orthogonal curvilinear model mesh grid with two mesh poles located in China and Australia is employed, as shown in Fig. 1a. The global mesh grid has 1200×740 points with a mean horizontal resolution of 6 km at the LS and 120 km in the tropical Atlantic Ocean. To satisfy the Courant–Friedrichs–Lewy (CFL) condition at the grids with the highest resolution, the model time step is set to 300 s. Forty uneven vertical layers are set from the ocean surface to the seafloor with 9 layers in the upper 100 m and 29 layers in the upper 1000 m. The model topography is interpolated from ETOPO2 (NGDC 2006); the maximum and minimum depths are set to 6500 and 31 m, respectively. The daily climatological surface forcing, including the surface air temperature, 2-m dewpoint temperature, sea level pressure, 10-m wind speed, wind stress, cloud cover, shortwave radiation, and precipitation (river runoff is ignored here), is interpolated from the database of the German Ocean Model Intercomparison Project (OMIP) (Röske 2001). Note that surface forcing is updated once per day to avoid high-frequency nontidal processes and is cycled once per year to obtain the climatological background ocean state. The initial temperature and salinity are interpolated from the Polar Science Center Hydrographic Climatology (PHC) data (Steele et al. 2001). During the model simulations, the ocean surface temperature and salinity are nudged toward the monthly PHC data. Figures 1b and 1c present the model mesh grid and topography in the NWP, which is the area of interest in this paper. Several marginal seas, such as the SCS, PS, East China Sea,

Sulu Sea, and Sulawesi Sea, along with many seamounts, ridges, and island chains, are located in this region.

The circulation and tide dynamics of the MPI-OM have been thoroughly tested and validated through the STORMTIDE project in previous studies (Müller et al. 2012; Li et al. 2015, 2017). In this work, two model configurations are used to simulate the tide and subtidal circulation dynamics.

The first model configuration is an idealized case in which all factors of the ocean surface forcing are set to zero, and the initial ocean stratification is set as horizontally homogeneous to the annual mean stratification in the LS; this approach has been widely applied in previous internal tide modeling studies. This configuration can provide a basic and overall understanding of the internal tides in the NWP area. To verify the tide module in the MPI-OM, a 13-month model run of the idealized case is performed, and the ocean surface elevation during the final 369 model days is analyzed to obtain the amplitudes and phases of the tides with the T-Tide Toolbox (Pawlowicz et al. 2002). Figure 2 shows a comparison between the cotidal charts derived from the model and from TPXO8 (Egbert and Erofeeva 2002) in the NWP area. Except for the larger amplitudes near the continental shelf area, the amplitudes and phases derived from the model and from TPXO8 agree well, indicating that the MPI-OM is adequately capable. Modulations of the internal tides at the surface elevation, such as the amplitude ripples and inflected cotidal lines in the upper four panels, are also observed, as has often been mentioned in previous work (Ray and Mitchum 1996; Jan et al. 2007). The surface elevations in the global MPI-OM exhibit larger amplitudes than the observed elevations, which is generally encountered in global ocean models (Stammer et al. 2014); this mainly originates from the underestimation of mixing in the abyssal sea, which leads to a lower sink of BT tide energy (Arbic et al. 2004). The overestimation of global BT tide energy may also lead to more conversion of BC tide energy.

The second configuration is a realistic case that considers both climatological surface forcing and astronomical tidal forcing. The realistic case is run for 10 model years to obtain a stable ocean circulation state. The result from the tenth model year is used to investigate the generation, propagation and dissipation of internal tides under realistic background conditions. The validation of the 10-yr climatological run is shown in Fig. 3. In the 10-yr run, the kinetic energy of ocean becomes stable after a 3-yr spinup and then exhibits annual cycles (Fig. 3a). The span of the model spinup is dramatically shortened due to the realistic settings of the initial temperature and salinity. In the last model year of the 10-yr climatological run, the western boundary currents, such as the Kuroshio and the Gulf Stream, as well as the Antarctic Circumpolar Current, are clearly reproduced well by our model (Figs. 3b,c). In addition, in the SCS, circulations along the western boundary as well as the northern shelf reveal seasonal variations in Figs. 3d and 3e. The seasonal variations come from opposite monsoons in winter and summer over the SCS.

The validation above demonstrates that the model utilized in this paper is reliable and that the model configuration is convincing. Based on the two configurations of our model, three numerical experiment sets are designed here (see Table 1). The standard set (STD) considers a realistic stratification and climatological surface forcing. Two control sets, including the stratification set (STRAT) and tidal set (TIDAL), both omit the surface forcing and assume an initial stratification (spatial averaged within the black box of Fig. 6) that is horizontally homogeneous. In the four cases of the TIDAL set, the initial stratification is controlled to be an annual mean stratification in the LS, while the tidal forcing is updated in real time. In the four cases of the STRAT set, the tidal forcing is controlled to that in January 2010, and the initial stratification is set to be horizontally homogeneous with seasonal values in the LS. Thus, the difference between the four cases in the STRAT set is the stratification, while the difference between the four cases in the TIDAL set is the tidal forcing. Note that each case in the two control sets is conducted separately for two model months, and only the results from the second month are used for diagnosis. Table 1 lists the tide and stratification settings of each case in the three experiment sets, as well as the case IDs for quick queries.

3. Methodology

Before presenting the model results, we first introduce the theoretical background and model postprocessing method in this section. The decomposition of variables in the derivation of BC tide energy equations is interpreted in the first part of this section. The BC tide energy equations in linear and non-linear frames are introduced in the second and third part of this section, respectively. According to the theoretical background, the model postprocessing method we use in this work is interpreted in the last part of this section.

a. Variable decomposition for theoretical background

To consider the effects of the background circulation and stratification on internal tides, we first decompose the variables into a mean state and a perturbation component as below:

$$a(x, y, z, t) = a^m(x, y, z) + a'(x, y, z, t). \quad (2)$$

In Eq. (2), the variable a can be a pressure p , density ρ , or velocity vector $\mathbf{u}$. An important assumption is that the background field is considered constant; thus, p^m , ρ^m , and $\mathbf{u}^m$ do not vary with time. We assume that tidal movement is the only source of perturbation in the system, and therefore, p' , ρ' , and $\mathbf{u}'$ are tide-induced perturbation variables.

Tidal movements are usually decomposed into BT and BC modes, which are expressed as below:

$$a^{\text{bt}} = \frac{1}{\eta + H} \int_{-H}^{\eta} a' dz, \quad a^{\text{bc}} = a' - a^{\text{bt}}. \quad (3)$$

In Eq. (3), the variable a' can be the pressure perturbation p' or horizontal tidal velocity $\mathbf{u}'_h$, while η and $-H$ represent the ocean surface and seafloor, respectively. Note that the vertical tidal velocity should be calculated from the horizontal tidal velocity by applying continuity from the seafloor to the ocean surface. The vertical BT/BC tidal velocity at the seafloor, depth z and ocean surface are expressed as an equation set (4); note that an asterisk (*) represents a superscript of either bt or bc:

$$\begin{cases} w^*(-H) = \mathbf{u}'_h \cdot \nabla_h(-H) \\ w^*(z) = w^*(-H) - \int_{-H}^z (\nabla_h \cdot \mathbf{u}'_h) dz \\ w^*(\eta) = \frac{\partial \eta}{\partial t} + \mathbf{u}'_h \cdot \nabla_h(\eta) \end{cases} \quad (4)$$

b. Linear baroclinic tide energy equations

The time-averaged, depth-integrated linear energy equations of the internal tides are given as below. The detailed derivation of this equation can be found in, for example, Gill (1982):

$$\text{Conv} - \text{Div}^{\text{bc}} = \varepsilon^{\text{bc}}. \quad (5)$$

In Eq. (5), Conv, Div^{bc} , and ε^{bc} represent the conversion of energy from BT to BC tides, the divergence of the BC tide energy flux, and the dissipation of BC tide energy, respectively. Conv and Div^{bc} are expressed as below:

$$\text{Conv} = \int_{-H}^{\eta} \rho' g w^{\text{bt}} dz, \quad (6)$$

$$\text{Div}^{\text{bc}} = \nabla_h \cdot \int_{-H}^{\eta} \mathbf{u}'_h p^{\text{bc}} dz. \quad (7)$$

In Eqs. (6) and (7), ρ' represents a density perturbation; p^{bc} and $\mathbf{u}'_h$ represent the BC components of the pressure perturbation p' and tidal velocity $\mathbf{u}'_h$, respectively; and w^{bt} denotes the vertical BT tidal velocity. The expressions of these variables are shown in Eqs. (2)–(4). Note that, according to the hydrostatic pressure equation, the energy conversion term Conv in Eq. (6) can also be expressed as Eq. (8) (see the appendix for details):

$$\text{Conv} = p^{\text{bc}}(-H)w^{\text{bt}}(-H) - p^{\text{bc}}(\eta)w^{\text{bt}}(\eta). \quad (8)$$

Equation (8) demonstrates that internal tides are generated at either the ocean surface or seafloor. Usually, the ocean surface

FIG. 3. Model validation of the 10-yr climatological run. (a) The variation of kinetic energy during the 10-yr climatological run. The blue line and red line represent the total kinetic energy of the global ocean and the SCS, respectively. (b),(c) The annual mean circulation in the Pacific Ocean and Atlantic Ocean (results from the tenth model year of the 10-yr climatological run), respectively, with colors representing the ocean surface height and vectors representing the depth-averaged velocity of the ocean in the upper 500 m. (d),(e) The seasonal variation of the circulation in the SCS (results from the tenth model year of the 10-yr climatological case), with colors representing the ocean surface height and vectors representing the depth-averaged velocity of the ocean in the upper 100 m.

TABLE 1. Basic information of the model cases in the standard set (STD), stratification set (STRAT), and tidal set (TIDAL). Note that the STD set is simulated for 10 model years with climatological surface forcing, and the results from January, April, July, and October in the tenth year are used. Each case in the STRAT and TIDAL sets is simulated for two model months without surface forcing, and the result from the second month is used.

		Winter (January)	Spring (April)	Summer (July)	Autumn (October)
STD set	Nonuniform stratification	Model result winter	Model result spring	Model result summer	Model result autumn
	BT tide	January 2010 tide	April 2010 tide	July 2010 tide	October 2010 tide
	Case ID	STD-Jan-Wi	STD-Apr-Sp	STD-Jul-Su	STD-Oct-Au
STRAT set	Uniform stratification	LS winter	LS spring	LS summer	LS autumn
	BT tide	January 2010 tide	January 2010 tide	January 2010 tide	January 2010 tide
	Case ID	STRAT-Wi	STRAT-Sp	STRAT-Su	STRAT-Au
TIDAL set	Uniform stratification	LS annual mean	LS annual mean	LS annual mean	LS annual mean
	BT tide	January 2010 tide	April 2010 tide	July 2010 tide	October 2010 tide
	Case ID	TIDAL-Jan	TIDAL-Apr	TIDAL-Jul	TIDAL-Oct

conversion rate is much smaller than the seafloor conversion rate (Nagai and Hibiya 2015), so the second term on the right-hand side of this equation can be neglected. In a Z -coordinate model, w actually equals zero at the seafloor, whereas the selection of another seafloor level may introduce considerable error. Thus, we choose Eq. (6) rather than (8) to calculate the energy conversion rate. Note that the linear BC tide energy equations are primarily used in this work except for section 4d.

c. Nonlinear baroclinic tide energy equations

To obtain the nonlinear BC tide energy equations considering the background field, the variable decomposition method in Eq. (2) must be applied. First, the horizontal momentum equation should be split into two momentum equations, namely, momentum equations for the mean flow and tidal flow; in addition, the tide energy equation should be derived from the tide momentum equation. Second, the tide momentum equation should be integrated with respect to depth to obtain the BT tide momentum equation and BT tide energy equation. Third, combining the tide energy equation, the BT tide energy equation and the available potential energy (APE) equation, the nonlinear BC tide energy equations can be obtained. A detailed derivation of the nonlinear BC tide energy equations is shown in the appendix.

The time-averaged, depth-integrated nonlinear BC tide energy equation is given as follows:

$$\text{Tran}^{\text{bc}} + \text{Conv} - \text{Div}^{\text{bc}} = \varepsilon^{\text{bc}}, \quad (9)$$

where Tran^{bc} , Conv , Div^{bc} , and ε^{bc} represent the transfer of energy from the mean flow to the BC tidal flow, the conversion of energy from BT to BC tidal flow, the divergence of the energy flux, and the dissipation of internal tides, respectively. The three terms on the left-hand side are given as below:

$$\text{Tran}^{\text{bc}} = - \int_{-H}^{\eta} (\rho_c \mathbf{u}_h^{\text{bc}} \mathbf{u}') \cdot \nabla \mathbf{u}_h^m dz, \quad (10)$$

$$\text{Div}^{\text{bc}} = \nabla_h \cdot \int_{-H}^{\eta} [\mathbf{u}_h (\text{ke}^{\text{bc}} + \text{ape}) + \mathbf{u}_h^{\text{bc}} p^{\text{bc}}] dz, \quad (11)$$

$$\text{Conv} = \int_{-H}^{\eta} \rho' g w^{\text{bt}} dz - \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{\text{bt}}) dz. \quad (12)$$

Compared with the linear equations, the nonlinear energy equations include several additional terms due to nonlinear effects; these nonlinear energy terms are discussed in section 4d. It should be noted that the nonlinear energy equations are based on the assumption that background currents are independent of time, so the nonlinear theory is recommended to be applied to a period longer than the time scale of ocean tides and shorter than the time scale of ocean circulations. We assume that the simulated background state is unchanged in 3 days; thus, a 3-day period is intercepted to diagnose the nonlinear energy equations in our work (see section 4d).

d. Postprocessing method

Since the MPI-OM considers all tidal constituents implicitly and the realistic background circulation is varying at relatively low frequencies, a fourth-order Butterworth filter is applied in this work to distinguish diurnal, semidiurnal, and all internal tides from all the signals in the model results. By Eqs. (6) and (7), to diagnose linear BC tide energy terms, the horizontal tidal velocity $\mathbf{u}_h'$, density perturbation ρ' , and pressure perturbation p' should first be filtered with specific bands. The filter band is set to 1.70–2.20 cpd for semidiurnal tides and 0.80–1.15 cpd for diurnal tides. A high-pass filter is used to obtain all components of internal tides; the cutoff frequency is set to 0.80 cpd. A consecutive 30-day result is used for filtering. Due to the boundary effect of filtering, the filtered results on the first and last 3 days are excluded. An average of 14 days (a spring-neap tide) is suggested to diagnose the dissipation terms to eliminate the error caused by tendency terms of the local BC tide energy. It should be noted that the filtering method is applied for both idealized and realistic sets. In addition, to diagnose the effect of the background circulation in the nonlinear Eqs. (10)–(12), we intercept a 3-day period in order to ensure that the background circulation is nearly constant.

4. Model results and discussion

a. Results of the idealized experiment

In this part, the results from an idealized case in the TIDAL set (TIDAL-Jan) are shown. As mentioned above, the initial

temperature and salinity of this case are set as horizontally homogeneous. The results of the idealized case, shown in Fig. 4, are diagnosed with the linear BC tide energy equations, Eqs. (5)–(7). Evidently, energy is converted from BT to BC tides (positive) and from BC to BT tides (negative). For example, diurnal BC tide energy is converted to diurnal BT tide energy near the Dongsha Atoll (20°N, 117°E). According to Eqs. (6) and (8), the phase difference between the vertical BT tidal velocity and density perturbation/BC pressure perturbation determines whether energy is converted from BT to BC tides or conversely (Zilberman et al. 2009).

As shown in Fig. 4a, most of the diurnal internal tides in the NWP area are generated within the LS, where two branches of diurnal internal tides propagate westward into the SCS and eastward into the PS over a long distance. Due to the flat topography within the PS Basin, internal tides can propagate over 2000 km along an arc across the entire PS. During the long-range propagation, diurnal internal tides bend toward the equator. This is caused by the refraction and is also interpreted by Zhao (2014). Westward-traveling internal tides propagate into the SCS Basin and interfere with two groups of internal tides generated from the northwest shelf-slope of the SCS (NWS) and the southwest shelf-slope of the SCS (SWS). Diurnal internal tides generated from the Sulu Islands and the Sulawesi Islands also exist in the Sulawesi Sea; part of the internal tides generated from the Sulawesi Islands propagates northeastward into the PS Basin, while the other part dissipates in the Sulawesi Sea together with the internal tides originating from the Sulu Islands. Some other sources of internal tides are located in the Ryukyu Islands, as well as some other islands, seamounts and shelf-break areas in the SCS.

Figure 4b reveals several major generation sites of semidiurnal internal tides in the NWP area. The LS is well known as an important generation site of internal tides and internal solitary waves in the northeast SCS. The region encompassing the Ryukyu Islands and the continental shelf-slope in the East China Sea generates massive semidiurnal internal tides, which is also simulated by Niwa and Hibiya (2004) to analyze the energy budget. The Sulawesi Islands and Sulu Islands also constitute important generation sites of semidiurnal internal tides. The model results show that the semidiurnal BC tide energy in both the Sulu Islands and the Sulawesi Islands is 4 times greater than the diurnal BC tide energy. The energetic semidiurnal BC tide energy in the Sulawesi Sea is also shown in Nagai and Hibiya (2015). It should be emphasized that the Ogasawara–Mariana ridge is an important generation site for semidiurnal internal tides (Hibiya et al. 2006; Niwa and Hibiya 2011; Zhao and D’Asaro 2011; Kerry et al. 2013). These generated internal tides propagate orthogonally to the strike of the Ogasawara–Mariana ridge into the PS Basin and the Pacific Ocean. Consequently, due to the interference of internal tides sourced from multiple surrounding generation sites, intricate patterns of energy fluxes emerge in the PS Basin.

A comparison of the two panels of Fig. 4 suggests that diurnal internal tides are more energetic in the SCS, while semidiurnal tides are more energetic in the PS. This can be explained by the difference in BT tidal forcing on the basis of the linear internal tide theory of Baines (1982). As shown in Fig. 2, the amplitudes

FIG. 4. Internal tide energy in the NWP, with vectors indicating linear energy fluxes and colors indicating linear energy conversion rates from BT to BC tides. Red boxes represent the main BC tide energy generation sites, including the Luzon Strait (LS), northwest shelf-slope area of the SCS (NWS), southwest shelf-slope area of the SCS (SWS), Sulu Islands (SULU), Sulawesi Islands (SULA), Ryukyu Islands (RI), Izu Ridge (IR), Ogasawara Ridge (OR), and Mariana Ridge (MR). Gray lines denote isobaths at 200, 500, 1000, and 2000 m. The results are derived from an idealized case in the TIDAL set (TIDAL-Jan).

of the diurnal BT tides are stronger in the SCS than in the PS, while the opposite is true for the semidiurnal BT tides, which is also reported in previous studies (Ye and Robinson 1983; Matsumoto et al. 2000). Note that the Izu Ridge generates both diurnal and semidiurnal internal tides, but they display different propagation characteristics. The Izu Ridge lies beyond the critical latitudes of diurnal internal tides (i.e., 30°N for the K_1 tide) yet within those of semidiurnal internal tides (i.e., 74.5°N for the M_2 tide). Hence, constrained by the dispersion features of internal tides, the diurnal internal tides generated at the Izu Ridge cannot propagate freely. Therefore, the diurnal BC tide energy fluxes are shaped like a vortex, which is quite different from the shape of freely propagating semidiurnal internal tides. These “trapped” diurnal internal tides have also been captured by other numerical models (Niwa and Hibiya 2011; Li et al. 2015), and the phenomena of trapped diurnal internal tides and propagating semidiurnal internal

FIG. 5. BC tide energy budget in the LS area, with vectors indicating linear energy fluxes and colors indicating linear energy conversion rates from BT to BC tides. Red arrows and numbers denote BC tide energy fluxes through the four boundaries of the blue box, and numbers inside the blue box denote the region-integrated BC tide energy dissipation/conversion rate. The diagnostic results of (a) diurnal, (b) semidiurnal, and (c) all tidal components with the linear energy equations. (d) The residual energy, which is the result of (c) minus the sum of (a) and (b).

tides have also been found in glider observations (Johnston and Rudnick 2015).

Since the stratification in each idealized case is set horizontally homogeneous to the values at the LS, a detailed calculation of the BC tide energy budget in the LS is further conducted to verify our model (Fig. 5). The energy fluxes of both diurnal and semidiurnal internal tides appear as clockwise structures similar to those captured by previous work (Alford et al. 2011; Pickering et al. 2015), which can be explained by the characteristics of the double-ridge topography in the LS: the prominent topography of the west ridge mainly lies in the northern part of the LS, while that of the east ridge mainly lies in the central and southern parts of the LS. The BC tide energy fluxes are almost zonal, thus, energy fluxes through the north and south boundaries are one order smaller than those through the west and east boundaries. The local dissipation rate is calculated by the energy conversion rates and energy fluxes averaged over 14 days. From the diagnostic result, the regional dissipation efficiency of semidiurnal internal tides in the LS (66.5%) is higher than that of diurnal internal tides (47.8%). For all internal tides, a regional dissipation efficiency of 53.8% is exhibited in the LS. Figure 5c features both the diurnal and the semidiurnal internal tides, along with some other signals. Comparing Fig. 5c with Figs. 5a and 5b, we can find that the sum of diurnal and semidiurnal tides deviates from all tides, and the deviation is shown in Fig. 5d. This residual can be explained in three parts. One is the nonlinear coupling of diurnal and semidiurnal variables, as shown in Eq. (13) (the underlined

terms). Another part is the nonlinear interaction between different tidal components, leading to BT/BC tides with higher frequencies, as shown in Eq. (14) (the underlined term). The last part is the numerical error caused by the filtering. Note that in Eqs. (13) and (14), subscripts 1 and 2 stand for diurnal and semidiurnal components, respectively:

$$\left\{ \begin{array}{l} \text{Flux} = (\mathbf{u}_1^{\text{bc}} + \mathbf{u}_2^{\text{bc}})(p_1^{\text{bc}} + p_2^{\text{bc}}) = \mathbf{u}_1^{\text{bc}}p_1^{\text{bc}} + \mathbf{u}_2^{\text{bc}}p_2^{\text{bc}} \\ \quad + \mathbf{u}_1^{\text{bc}}p_2^{\text{bc}} + \mathbf{u}_2^{\text{bc}}p_1^{\text{bc}} \\ \text{Conv} = g(\rho'_1 + \rho'_2)(w_1^{\text{bt}} + w_2^{\text{bt}}) = \rho'_1gw_1^{\text{bt}} + \rho'_2gw_2^{\text{bt}} \\ \quad + \rho'_1gw_2^{\text{bt}} + \rho'_2gw_1^{\text{bt}} \end{array} \right. \quad (13)$$

$$\cos(\omega_1 t) \cos(\omega_2 t) = \frac{1}{2} \cos[(\omega_1 + \omega_2)t] + \frac{1}{2} \cos[(\omega_1 - \omega_2)t]. \quad (14)$$

By diagnosing these terms, we found that the cross terms in Eq. (13) are negligible and account for less than 1% of the total energy. That conclusion is consistent with Buijsman et al. (2014) (less than 3%). Additionally, only 1 GW of high-frequency BC tides is generated in the LS. Both factors cannot take charge of the residual. However, with a property of flat frequency response in the passband, the Butterworth filtering method may cause error in estimating the energy of filtered signals. Thus, the main part of the residual can only be explained by the numerical error caused by filtering.

FIG. 6. The five subregions in the LS. Colors indicate BC tide energy conversion rates; the results of an idealized case in the TIDAL set (TIDAL-Jan) are presented. Since the prominent topography for generating (a) diurnal and (b) semidiurnal internal tides is not consistent among the subregions, the division is slightly different for the two types of internal tides. The five subregions are named the north part of the east ridge (EN), middle part of the east ridge (EM), south part of the east ridge (ES), north part of the west ridge (WN), and middle part of the west ridge (WM). The initial stratification setting in the TIDAL and STRAT sets are regional-averaged values in the black box. Gray lines denote isobaths at 200, 500, 1000, and 2000 m.

b. Comparison between the idealized and realistic experiments: Effects of the tidal forcing and seasonal stratification on the generation of internal tides

The tidal forcing and seasonal variation of stratification are believed to be the two main factors affecting the generation of internal tides (Jan et al. 2008). In this part, we compare the two idealized experiment sets, namely, the TIDAL and STRAT sets, with the STD set to evaluate the significance of the two main factors.

We first divide the LS into five subregions to investigate the roles that these two main factors, i.e., BT tidal forcing and seasonal stratification, play in the generation of internal tides. Because the generation sites of diurnal and semidiurnal internal tides are not exactly consistent, we apply a different division method for the two types of internal tides, as shown in Fig. 6.

The horizontal integrals of the BC tide energy conversion rate over the five subregions are presented in Fig. 7. As shown in Figs. 7a and 7b, with differences in the seasonal stratification, the conversion rates in different subregions have different features: some subregions are sensitive to the seasonal variation in the stratification, while others are not. Some subregions also exhibit differences in their variability. For example, the middle part of the east ridge (EM) generates the most semidiurnal BC tide energy with winter stratification, while the north part of the west ridge (WN) generates the most semidiurnal BC tide energy with spring stratification. Figures 7c and 7d show that all five subregions exhibit the same trend in each panel, which reveals that the strength of the BT tidal forcing in the LS has a consistent effect on all five subregions. For example, among the four cases, the BT tidal forcing in the LS is stronger in July than in the other three periods, so the result in July is the most energetic for both diurnal and

semidiurnal internal tides. Closely examining Figs. 7e and 7f, we can see that the STD set features nearly the same trends as the TIDAL set. This finding reveals that the tidal forcing is the dominant factor that consistently controls the strength of the BC tide energy conversion rate, while seasonal stratification plays only a secondary role, and different subregions in the LS respond differently to the seasonal variation of the stratification. It should also be mentioned here that a general big deviation between the TIDAL and STD set (comparing Figs. 7c and 7e, Figs. 7d and 7f) emerges at the east ridge of the LS. This mainly comes from the horizontal inhomogeneity of the realistic stratification at the LS, where the strong geostrophic flow (viz., the Kuroshio) is located.

In addition to the LS, we also analyze other internal tide generation sites in the SCS and PS to estimate the deviations in the BC tide energy conversion rate from different experiment sets. Table 2 shows the results of two cases, one in the STD set and one in the STRAT set, as well as the deviation between the two cases. Compared to the idealized case (STRAT-Wi), the realistic case (STD-Jan-Wi) with 3D stratification is expected to be more robust. Table 2 further illustrates that some regions, such as the NWS and Sulu Islands (SULU) areas, are characterized by only minor deviations in the BC tide energy conversion rates than other regions, indicating that horizontally homogeneous stratification does not introduce much error in these areas. However, other regions exhibit large deviations between the realistic case and the idealized case due to large discrepancies in the stratification therein; for example, this deviation can lead to an error reaching 3.99 GW in the Sulawesi Islands (SULA). Although we previously concluded that seasonal variation in stratification does not play a primary role in controlling the generation of internal tides, the large discrepancy between the stratification in the LS and that in remote

FIG. 7. BC tide energy conversion rates (GW) in the (a),(b) STRAT set, (c),(d) TIDAL set, and (e),(f) STD set as well as the “seasonal variation” in different cases within the five subregions shown in Fig. 6. Panels (a), (c), and (e) represent diurnal internal tides, while (b), (d), and (f) represent semidiurnal ones.

regions can still lead to an error exceeding 30% in the energy conversion rate (e.g., in the SWS area), which means that a horizontally homogeneous initial stratification is not suitable for performing a simulation of the whole NWP area and that the STD set is more applicable.

c. Comparison between the idealized and realistic experiments: Effects of the background field on the propagation of internal tides

In this part, we compare the results from an idealized case (TIDAL-Apr) and a realistic case (STD-Apr-Sp) in order to investigate the effect of the background field on the

propagation of internal tides. Though the background field is changing at a low frequency, a 3-day period (from 18 April to 20 April) is selected here to satisfy the assumption that the background field is not a variable of time.

In an idealized case, without considering background circulation, the horizontal phase speed of internal tides can be given as below:

$$c_p^2 = \frac{\omega^2}{\omega^2 - f^2} c_n^2. \quad (15)$$

In Eq. (15), ω is the frequency of the internal tide and f is the inertial frequency. Parameter c_n is the Sturm–Liouville

TABLE 2. BC tide energy conversion rates (GW) at generation sites other than the LS in the SCS and PS. The table shows the results from one case in the STD set (STD-Jan-Wi) and one case in the STRAT set (STRAT-Wi). The deviation between these two cases and ratio of this deviation to the STD-Jan-Wi case are shown in the third row and the fourth row, respectively. The regions are shown in Fig. 4.

Case	Diurnal				Semidiurnal				
	NWS	SWS	SULU	SULA	RI	OR	MA	SULU	SULA
STD-Jan-Wi	4.42	1.70	3.58	8.65	25.77	12.13	12.30	10.00	19.18
STRAT-Wi	4.49	2.31	2.96	6.33	22.72	10.65	9.93	10.40	23.17
STD-STRAT	−0.06	−0.62	0.62	2.32	3.05	1.48	2.37	−0.40	−3.99
(STD-STRAT)/STD	−1.43%	−36.31%	17.38%	26.81%	11.81%	12.20%	19.30%	−3.99%	−20.80%

eigenvalue speed and depends on the water depth and stratification. The Sturm–Liouville equation is expressed as below:

$$\frac{d^2\Phi_n(z)}{dz^2} + \frac{N^2(z)}{c_n^2}\Phi_n(z) = 0. \quad (16)$$

With a rigid-lid and flat-bottom assumption, the surface and bottom boundary conditions are set to zero (Gerkema and Zimmerman 2008). Thus, one can obtain the eigenvalue speed c_n by solving Eq. (16) with the specific stratification $N^2(z)$. This method has been discussed extensively in previous studies (e.g., Li et al. 2015; Xu et al. 2016) and has been interpreted by Zhao (2014) in detail. It should be claimed here that the zero boundary condition is not accurate for free surface (Kelly 2016) and slope bottom (Lahaye and Llewellyn Smith 2020). The free surface condition is not applied in our work because the effect is small for the mode-1 internal tides. Furthermore, we apply the flat-bottom condition to the whole model domain because slope bottoms (i.e., the continental slopes) only take a small proportion of the NWP and are not the focus in this work.

When considering the existence of the background circulation $\mathbf{u}^m(x, y, z)$, the horizontal and vertical geostrophic shear can affect the propagation of internal tides. The effect can be explained to change the Coriolis frequency f and buoyancy frequency N^2 to the effective Coriolis frequency f_{eff} and effective buoyancy frequency N_{eff}^2 , which is shown in Eqs. (17) and (18) (Kunze 1985):

$$f_{\text{eff}} \approx f + \frac{1}{2} \left(\frac{\partial v^m}{\partial x} - \frac{\partial u^m}{\partial y} \right), \quad (17)$$

$$N_{\text{eff}}^2 = N^2 + 2M_x^2 \frac{k_x k_z}{k_h^2} + 2M_y^2 \frac{k_y k_z}{k_h^2}. \quad (18)$$

In Eq. (18), M_x^2 and M_y^2 can be expressed as the horizontal gradient of background density ρ^m or the vertical shear of background circulation (u^m, v^m) due to the thermal wind relationship:

$$\begin{cases} M_x^2 = \frac{g}{\rho_0} \frac{\partial \rho^m}{\partial x} = -f \frac{\partial v^m}{\partial z} \\ M_y^2 = \frac{g}{\rho_0} \frac{\partial \rho^m}{\partial y} = +f \frac{\partial u^m}{\partial z} \end{cases}. \quad (19)$$

Combined with the dispersion relation of internal waves under the hydrostatic approximation, the final form for calculating N_{eff}^2 in this work is shown in Eq. (20). In our work, the unit vector ($k_x/k_h, k_y/k_h$) is prescribed via the horizontal energy flux at each grid point in the corresponding idealized case of the TIDAL set:

$$N_{\text{eff}}^2 = N^2 - 2f \sqrt{\frac{N^2}{\omega^2 - f^2}} \frac{\partial v^m}{\partial z} \frac{k_x}{k_h} + 2f \sqrt{\frac{N^2}{\omega^2 - f^2}} \frac{\partial u^m}{\partial z} \frac{k_y}{k_h}. \quad (20)$$

By replacing f with f_{eff} in Eq. (15) and N^2 with N_{eff}^2 in Eq. (16), one can obtain the equations below to calculate the phase speeds of internal tides considering the background circulation:

$$c_p^2 = \frac{\omega^2}{\omega^2 - f_{\text{eff}}^2} c_n^2, \quad (21)$$

$$\frac{d^2\Phi_n(z)}{dz^2} + \frac{N_{\text{eff}}^2(z)}{c_n^2} \Phi_n(z) = 0. \quad (22)$$

It should also be noted here that the model result is represented in the Eulerian frame, so the frequency of internal tides is changed to a Eulerian Doppler-shifted frequency $\omega' = \omega + \mathbf{k} \cdot \mathbf{u}^m$, in which $\mathbf{k}$ represents wavenumber and $\mathbf{u}^m$ denotes background circulation. The estimation of the Doppler-shifted frequencies in this work is simplified as $\omega' = \omega + \mathbf{k}_h \cdot \mathbf{u}_h^m$, by using horizontal wavenumber $\mathbf{k}_h$ and horizontal background circulation $\mathbf{u}_h^m$. The calculation of the horizontal wavenumber $\mathbf{k}_h$ can be divided into two components. The magnitude of the horizontal wavenumber is calculated as $|\mathbf{k}_h| = \sqrt{(\omega^2 - f^2)/c_n^2}$, where c_n is the eigenvalue speed from Eq. (16), and the direction of the horizontal wavenumber is prescribed as the calculated energy flux vector at each grid point. The horizontal background circulations $\mathbf{u}_h^m$ are depth-averaged values over the upper 500 m. Note here that we use the “old” horizontal wavenumber $\mathbf{k}_h$ from the corresponding idealized case (TIDAL-Apr), assuming that the horizontal wavenumber is not significantly changed from the idealized case to the realistic case because the horizontal wavenumber $|\mathbf{k}_h|$ itself is small and less important compared to the strength of the background flow $\mathbf{u}_h^m$ and the angle between $\mathbf{u}_h^m$ and $\mathbf{k}_h$. Additionally, in a Eulerian frame, the filtered results in the realistic case include inaccuracies caused by the Doppler-shifting effect (see below Fig. 10).

The calculated eigenvalue speeds in both the idealized case and realistic case, along with the background circulation in the realistic case, are shown in Fig. 8. According to Eqs. (16) and (22), for each grid point, the difference between $N^2(z)$ and $N_{\text{eff}}^2(z)$ is the only factor that affects the eigenvalue speed. Thus, the deviation between the eigenvalue speeds in these two cases is shown in Figs. 8b and 8d to demonstrate the influence of 1) stratification difference and 2) isopycnal tilt [see Eqs. (18) and (19)]. Examining Figs. 8b and 8d, we can find negative and positive biases of eigenvalue speeds located west and east of the Kuroshio path, indicating a significant difference in stratification between the SCS and the PS. Although the stratification conditions in the PS and LS differ greatly, the deviation in the eigenvalue speed in the PS Basin is less than 0.1 m s^{-1} . By estimating with Eq. (15), the difference of the phase speed caused by changing from $N^2(z)$ to $N_{\text{eff}}^2(z)$ is $\sim 0.1 \text{ m s}^{-1}$.

After calculating the eigenvalue speeds c_n , the horizontal phase speeds of internal tides can be calculated via Eqs. (15) and (21). Note that if the horizontal phase speed satisfies $c_p^2 < 0$, then complex solutions may be obtained. A complex solution of the horizontal phase speed indicates that the internal tide cannot propagate horizontally. In Eq. (15), the condition for a propagating internal tide is $\omega > f$. However, when considering the background circulation, the condition is changed to $\omega > f_{\text{eff}}$. The calculation of the effective Coriolis frequencies is shown in Eq. (17), demonstrating that the effective Coriolis frequency consists of the Coriolis frequency and background vorticity. If f_{eff} is increased to a value close to

FIG. 8. (a) Distribution of the mode-1 internal tide eigenvalue speed in the idealized case (TIDAL-Apr). (b),(d) Differences in the mode-1 internal tide eigenvalue speed between the realistic case (STD-Apr-Sp) and idealized case (TIDAL-Apr) for K_1 and M_2 tides. (c) The mean background circulation between 18 and 20 April in the realistic case (STD-Apr-Sp). Vectors in (c) indicate the depth-averaged background circulation of the ocean in the upper 500 m, while colors represent the depth of the 1027 kg m^{-3} isopycnal.

or even larger than the tidal frequency ω , then internal tides are refracted or blocked in this area.

In the NWP, the propagation of diurnal internal tides is sensitive to the effective Coriolis frequencies since the latitudes are close to the critical latitudes of diurnal internal tides (i.e., 30°N for the K_1 tide). Figure 9 compares the result of an idealized case in the TIDAL set (TIDAL-Apr) with the corresponding realistic case in the STD set (STD-Apr-Sp). Comparing Fig. 9e with Fig. 8c, we can see that background shear affects f_{eff} directly; in some locales, the frequency can approach the critical value (frequency ratio $f_{\text{eff}}/\omega_{K_1}$ close to 1). In Fig. 8c, a warm eddy is located west of the LS, while four eddies are located east of the LS. Warm eddies reduce f_{eff} , while cold eddies raise f_{eff} . Figure 9d demonstrates that the horizontal phase speeds of internal tides in some areas are dramatically changed due to the changing of the effective Coriolis frequencies. Zhao (2014) concluded that the refraction of internal tides obeys Snell's law, indicating that propagating internal tides tend to bend to locations with lower horizontal phase speeds. Thus, by comparing Figs. 9a and 9b, we can find that the eastward-propagating internal tides are split into two branches by the cold eddy whose center is located at 21.5°N , 122.5°E . Subsequently, the two branches of these internal tides propagate through two warm eddies, where the horizontal phase speeds are somewhat low. The westward-propagating internal tides come across with the warm background eddy whose center is located at 20°N , 119°E . With smaller horizontal phase speeds caused by the warm eddy, the westward-propagating internal tides are expected to be more

convergent. However, the convergence is not obvious because of the hyperbolic formula [see Eq. (21)]. Horizontal phase speeds are more affected when f_{eff} approaches ω , and vice versa.

The propagation of semidiurnal internal tides is not as sensitive to the eddies as diurnal ones because the critical latitudes are far away (i.e., 74.5°N for the M_2 tide). Thus, f_{eff} cannot be raised close to a frequency as high as ω_{M_2} in this region (Fig. 10e). Figure 10d also proves that the changing of M_2 phase speeds due to the vorticity of background current is not as prominent as the K_1 phase speeds shown in Fig. 9d. However, changing of propagation path can still be noticed by comparing Figs. 10a and 10b. By observing the Doppler-shifted frequency (Fig. 10f), we can find that frequencies in some areas are significantly changed beyond or below the selected filtering band (encircled by black dashed line) due to the Doppler-shifting effect. The changing of frequencies in the Eulerian frame leads to the “blocking” of propagating internal tide at a fixed filtering band. Comparing Figs. 9f and 10f, we can find that the Doppler-shifting effect ($\mathbf{k}_h \cdot \mathbf{u}_h^m$) is more prominent for semidiurnal internal tides because semidiurnal internal tides have higher wavenumbers (smaller wavelengths) than diurnal ones.

Thus, in our result, the vorticity of background circulation mainly alters the propagating diurnal internal tides by changing the Coriolis frequencies into the effective Coriolis frequencies, while for semidiurnal internal tides, the Doppler-shifting effect plays an important role. In the Eulerian frame, which is used in most of the ocean models and in situ measurements,

FIG. 9. Energy fluxes of diurnal internal tides in the (a) idealized case (TIDAL-Apr) and (b) realistic case (STD-Apr-Sp), with colors indicating the vector magnitude. (c),(d) The calculated horizontal phase speeds via Eqs. (15) and (21), respectively. Note that (d) exhibits differences with the phase speeds shown in (c). (e) The distribution of the effective Coriolis frequencies f_{eff} divided by K_1 frequency ω_{K_1} in the realistic case. (f) The distribution of the Doppler-shifted K_1 frequency $\omega'_{K_1} = \omega_{K_1} + \mathbf{k}_h \cdot \mathbf{u}_h^m$ in the realistic case, and the black dashed lines encircle areas where frequencies are shifted out of the diurnal filtering band. Note that all the values are calculated in a 3-day period (from 18 to 20 April).

the background circulation shifts wave frequencies. This causes wave signals to be missed from the filtered result and leads to the “different propagation paths” between the idealized cases and realistic cases. The vertical shear of geostrophic flow (or the tilt of isopycnals) changes the buoyancy frequencies into the effective buoyancy frequencies, but the effect is not significant.

Note here that the vertical shear of geostrophic flows also alters the minimum frequency $\omega_{\min}$ (Whitt and Thomas 2013), which is not included in our work. Considering the baroclinicity of the background circulation, the lower bounds of frequencies decrease from the effective Coriolis frequencies f_{eff} to the minimum frequencies $\omega_{\min}$, extending the propagating range of trapped subinertial internal waves [i.e., $(0.95 \pm 0.05)f$]. In our case, $\omega_{K_1} = 1.46f$ at the LS (20°N) and is too large for

near-inertial frequencies, indicating the negligible effect. However, we speculate that, for internal tides close to the critical latitudes, the baroclinicity is influential. For example, the trapped diurnal internal tides at the Izu Ridge where $\omega_{K_1} = 0.92f$ (see Fig. 4a) can be significantly affected by the vertical shear of geostrophic flows.

d. The effects of nonlinear energy terms

The nonlinear energy equations of internal tides considering the background field are presented in Eqs. (9)–(12), as presented in section 3c and derived in the appendix. To further assess the significance of nonlinear effects, we compare the linear equations with the nonlinear equations and diagnose the differences using the model results from a realistic case (STD-Jan-Wi). Note that in this part, we compare the nonlinear BC

FIG. 10. As in Fig. 9, but for M_2 tides. Note that the frequency ratio in (e) is $f_{\text{eff}}/\omega_{M_2}$, while in (f), Doppler-shifted M_2 frequency $\omega'_{M_2} = \omega_{M_2} + \mathbf{k}_h \cdot \mathbf{u}_h^m$, and the black dashed lines encircle areas where frequencies are shifted out of the semidiurnal filtering band.

tide energy budget with the linear budget in consideration of all the tidal components; thus, the tidal movements a' derived from the variable decomposition method of Eq. (2) represent all tides and are extracted by applying a high-pass filter as mentioned before. Additionally, the results in this part are extracted from 13 to 15 January to guarantee that the background field is nearly unchanged.

In the idealized cases, we assume that the background stratification is horizontally homogeneous and that only vertical movements can produce APE. However, when a three-dimensional background stratification is considered, horizontal movements can also introduce APE by diapycnal movements of water masses, and the difference in the BC tide energy conversion rate can be given as below:

$$\delta C = \int_{-H}^{\eta} \rho' g w^{\text{bt}} dz - \int_{-H}^{\eta} \rho' g \frac{-\nabla \rho^m}{|\nabla \rho^m|} \cdot \mathbf{u}^{\text{bt}} dz. \quad (23)$$

When decomposing variables into mean states and tidal components, the nonlinear interaction between the background

shear and internal tides leads to the energy transfer between the two systems, which is expressed as below:

$$I_{\text{m-bc}} = -\rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}') \cdot (\nabla \mathbf{u}_h^m) dz. \quad (24)$$

The formula in the integral sign can be expanded as follows:

$$\begin{aligned} (\mathbf{u}_h^{\text{bc}} \mathbf{u}') \cdot (\nabla \mathbf{u}_h^m) &= \mathbf{u}_h^{\text{bc}} \cdot (\mathbf{u}' \cdot \nabla \mathbf{u}_h^m) \\ &= \mathbf{u}_h^{\text{bc}} \cdot \left(u' \frac{\partial \mathbf{u}_h^m}{\partial x} + v' \frac{\partial \mathbf{u}_h^m}{\partial y} + w' \frac{\partial \mathbf{u}_h^m}{\partial z} \right) \\ &= u^{\text{bc}} \left(u' \frac{\partial u^m}{\partial x} + v' \frac{\partial u^m}{\partial y} + w' \frac{\partial u^m}{\partial z} \right) \\ &\quad + v^{\text{bc}} \left(u' \frac{\partial v^m}{\partial x} + v' \frac{\partial v^m}{\partial y} + w' \frac{\partial v^m}{\partial z} \right). \end{aligned} \quad (25)$$

Analogously, when decomposing tidal variables into BT and BC components, the nonlinear interaction between BT and BC tides leads to a similar transfer of energy between the two systems, which is expressed as below:

FIG. 11. Diagnostic results of the four nonlinear terms in a realistic case (STD-Jan-Wi): (a) δC in Eq. (23), (b) I_{bt-bc} in Eq. (26), (c) I_{m-bc} in Eq. (24), and (d) $\mathbf{F}_{non}^{bc}$ in Eq. (28). The color bars are logarithmic from approximately -0.3 to 0.3 . Note that all the values are calculated in a three-day period (from 13 to 15 January).

$$I_{bt-bc} = -\rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{bc} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{bt}) dz. \quad (26)$$

The formula in the integral sign can be expanded as follows:

$$\begin{aligned} (\mathbf{u}_h^{bc} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{bt}) &= \mathbf{u}_h^{bc} \cdot (\mathbf{u}_h \cdot \nabla_h \mathbf{u}_h^{bt}) = \mathbf{u}_h^{bc} \cdot \left(u \frac{\partial \mathbf{u}_h^{bt}}{\partial x} + v \frac{\partial \mathbf{u}_h^{bt}}{\partial y} \right) \\ &= u^{bc} \left(u \frac{\partial u^{bt}}{\partial x} + v \frac{\partial u^{bt}}{\partial y} \right) + v^{bc} \left(u \frac{\partial v^{bt}}{\partial x} + v \frac{\partial v^{bt}}{\partial y} \right). \end{aligned} \quad (27)$$

Considering the advective effect of the background circulation on local internal tide energy, the nonlinear energy flux term is expressed as follows:

$$\mathbf{F}_{non}^{bc} = \int_{-H}^{\eta} [\mathbf{u}_h (k e^{bc} + a p e)] dz. \quad (28)$$

Equation (23) expresses the impact on the energy budget when considering inclined isopycnals, while Eqs. (24)–(28) reveal the residual terms of the nonlinear energy equations in comparison with the linear equations. Note that Eqs. (24) and (26) present similar expressions. The strength of the nonlinear interaction is related to the shear of other flows (mean flow or BT tide).

Since the expressions have already been listed in Eqs. (23)–(28), a diagnosis of the model result (using STD-Jan-Wi as an example) is performed below to quantitatively evaluate the nonlinear energetic effect of a realistic background.

Figure 11 shows the diagnosed result, where Figs. 11a–d successively present the values of δC in Eq. (23), I_{bt-bc} in Eq. (26), I_{m-bc} in Eq. (24) and $\mathbf{F}_{non}^{bc}$ in Eq. (28). Clearly, the magnitudes of the nonlinear terms are one to two orders smaller than those of the linear terms (comparing to Figs. 4 and 5). The discrepancy in the conversion rate shown in Fig. 11a occurs only along the Kuroshio path with strong BT tides, such as the LS and the Tokara Strait. The nonlinear interaction between BT and BC tides reflected in Fig. 11b usually occurs at the generation sites of internal tides, while the nonlinear energy flux term shown in Fig. 11d is significant at places where the internal tides and background flow are both strong. However, comparing Fig. 11c with the other three panels suggests that the nonlinear interaction between the mean flow and internal tide is relatively strong and is more widely distributed because the two influencing factors, internal tides and background shear, are both ubiquitous throughout the whole domain. To further assess the local significance of these terms, the greater LS is chosen to analyze these energy terms quantitatively.

The diagnosed result in the greater LS area is shown in Fig. 12. Evidently, I_{m-bc} is more significant than both δC and I_{bt-bc} . The regional integrals of positive/negative values are shown at the bottoms of Figs. 12a–c. For example, in Fig. 12c, in the greater LS area, internal tides receive 5.67 GW of energy from background shear while losing 4.69 GW at the same time. Comparing Fig. 12c with Figs. 12a and 12b, it is obvious that the term I_{m-bc} is about one order of magnitude larger than the

FIG. 12. As in Fig. 11, but for the greater LS area (outer range). Note that the color bars are linear in this figure. The numbers shown in (a)–(c) present the regionally integrated positive/negative values in this domain, while the numbers shown in (d) are the line-integrated positive/negative zonal energy fluxes along the blue line.

other two terms. However, in comparison with Fig. 5c, the term I_{m-bc} is clearly one order of magnitude smaller than the linear energy conversion rate. The vectors in Fig. 12d reflect the advective effect of background flow on BC tide energy, which can form nonlinear energy fluxes reaching approximately 15 KW m^{-1} in the LS. The “loop” structure of the Kuroshio transports 0.93 GW of BC tide energy westward through the southern half of the blue line while simultaneously transporting 0.92 GW eastward through the northern half.

Summarizing all four cases in the STD set, the terms caused by inclined isopycnals and the nonlinear BT–BC tide interaction can be omitted, whereas the other two terms, namely, the term comprising the nonlinear interaction between the mean flow and BC tide and the nonlinear energy flux term, cannot be neglected. By summing all the nonlinear terms and linear terms, it is estimated that nonlinearity can contribute 5% of the total BC tide energy budget in the LS. The resolution in the LS is approximately 6 km in our model configuration, which is still coarse for nonlinear phenomena. Thus, we speculate that the nonlinearity could become more prominent with an increasingly finer resolution.

5. Summary and conclusions

With abundant BC tide energy, the internal tides in the NWP area have attracted considerable attention. Although the assumption of horizontally homogeneous stratification has been widely used for the background state in previous work, this approach may cause some interesting phenomena in areas with multiscale ocean processes to be missed. With a global circulation and tide model with nonuniform resolution and curvilinear orthogonal mesh grid, we simulate the global circulation and BT–BC tides simultaneously. The global mesh grid is focused on the NWP area to resolve the internal tides,

the Kuroshio and other eddy processes therein. Considering the lunisolar tidal potential, the global MPI-OM contains all the tidal components implicitly, which is different from regional models forced by specific tidal components at the open boundaries.

First, by assuming that the background stratification is horizontally homogeneous, idealized cases are analyzed to diagnose the generation and propagation of internal tides and facilitate a comparison with previous results. The model results indicate that diurnal internal tides are more energetic in the SCS, while semidiurnal tides are more energetic in the PS, which is consistent with previous numerical findings (Xu et al. 2016) and altimetric results (Zhao 2014). Except for the LS, which is the most important source region of internal waves, most generation sites of diurnal internal tides are located in the shelf-slope area and seamounts of the SCS, while semidiurnal internal tides are generated mainly at ridges around the PS Basin. Moreover, a detailed energy budget for the LS is calculated and discussed.

Second, by comparing realistic cases with idealized cases, we discuss the effects of realistic settings of the background state on the generation of internal tides. Tidal forcing and stratification have always been considered to be the two main influential factors on the generation of internal tides. By comparing the TIDAL set and the STRAT set with the STD set, the BC tide energy conversion rate is determined to be affected mainly by the strength of the tidal forcing, while the seasonal variation of stratification has only a secondary effect. Although the seasonal variations of stratification are not large enough to prominently affect the BC tide energy conversion rate, the deviations in different regions can lead to an error exceeding 30%, which suggests that a realistic setting of the background stratification is necessary to simulate a large region such as the NWP.

Third, two main factors are found to influence the propagation of internal tides. Eddies and background shear can refract propagating internal tides by changing the effective Coriolis frequencies as well as phase speeds. Because inertial frequencies are close to diurnal frequencies in the NWP, diurnal internal tides are more sensitive to background vorticities. The Doppler-shifting effect is more remarkable for semidiurnal internal tides because they have higher wavenumbers than diurnal ones. In the Eulerian frame, the intrinsic wave frequencies are changed to Doppler-shifted frequencies and are shifted beyond or below the filtering band, which causes the difference in semidiurnal internal tides between the idealized cases and realistic cases. It should be noted here that both the effective Coriolis frequencies and the Doppler-shifting effect are qualitatively estimated using simplified 2D (depth-averaged over the upper 500 m) background velocities, which is not recommended for theoretical arithmetic.

Finally, the energetic effects of the realistic settings are discussed. By deriving the nonlinear energy equations of internal tides, we list four terms to evaluate the energetic effect of the background field. As the results of our diagnosis show, the advective effect of the mean flow and the transfer of energy between the mean flow and internal tides are the two most important nonlinear terms. In addition, the regional integration results in the greater LS area demonstrate that nonlinear energy terms contribute approximately 5% of the total value. Moreover, the rate of nonlinearity is expected to rise with an increasingly higher resolution.

Challenges still remain for future work. First, the amplitudes of the global BT tides are large compared with the amplitude in the TPX08 data because of the underestimation of dissipation in the abyssal sea, which may lead to the overestimation of BC tide generation. Thus, a term considering internal wave drag should be added to the tide module to estimate the BT/BC tide energy budget more accurately. Second, since most of the ocean models are “observing” in the Eulerian frame, the Doppler-shifting effect caused by background circulation appeals for a better data filtering method in the model postprocessing. Methods such as fixed bandpass filtering or harmonic analysis may introduce biases to the postprocessing results. Third, the seasonal variation in the Kuroshio intrusion into the SCS is not the same as that reported in previous studies; we assume that the OMIP surface forcing may not be the best choice in simulating the NWP with a high resolution. Thus, more surface forcing databases will be tested to provide a better subtidal circulation result for the NWP area. Last but not least, the surface forcing in our model is climatological; we plan to simulate real-time cases in the future to compare the model results with altimeter data while also conducting case studies on the interactions between internal tides and specific mesoscale/submesoscale activities.

Acknowledgments. The authors appreciate the comments from the editors and anonymous reviewers and the help from Dr. Johann Jungclauss, Dr. Helmuth Haak, Dr. Jin-Song von Storch and Dr. Zhuhua Li from the Max Planck Institute for Meteorology regarding the MPI-OM. The authors are also grateful to Dr. Taira Nagai from the University of Tokyo for his fruitful discussion on the derivation of the energy equations. This work is supported by the National Key Research and

Development Plan, Grant 2016YFC1401300 (“Oceanic Instruments Standardization Sea Trials (OISST)”), the National Science Foundation of China (NSFC), “Study on the influence of background current and stratification on the generation and propagation of internal tide in the Luzon Strait”, and the Taishan Scholar Program. We also thank the National Supercomputer Center in Jinan and the Pilot National Laboratory for Marine Science and Technology in Qingdao for the provision of computing resources.

APPENDIX

Derivation of the Baroclinic Tide Energy Equations Considering the Background Field

a. Primitive equations

To evaluate the effect of the background field on internal tides, the primitive equations with a Boussinesq approximation and a hydrostatic approximation in Cartesian coordinates are listed below. The Coriolis force is omitted because it does not have an energetic effect on the energy equation, which is expressed as $\mathbf{u} \cdot (\boldsymbol{\omega} \times \mathbf{u}) = 0$.

$$\frac{\partial \mathbf{u}_h}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u}_h = -\frac{1}{\rho_c} \nabla_h p - \nabla_h \Omega + \mathbf{V}, \quad (\text{A1})$$

$$\nabla \cdot \mathbf{u} = 0, \quad (\text{A2})$$

$$0 = -\frac{\partial p}{\partial z} - \rho g, \quad (\text{A3})$$

$$\frac{d\rho}{dt} = D_\rho. \quad (\text{A4})$$

Equations (A1)–(A4) are (in order) the horizontal momentum equation, continuity equation, hydrostatic pressure equation and density transport equation. In these equations, the subscript h means a horizontal component, and bold font denotes a vector, such as $\mathbf{u} = (u, v, w)$ and $\mathbf{u}_h = (u, v)$. The term ρ_c is the constant density in the Boussinesq approximation; ρ represents the density, and Ω represents the tidal potential. The terms $\mathbf{V}$ and D_ρ indicate the eddy viscosity and diffusivity, respectively, without specific expressions; therefore, in this work, dissipation is diagnosed by the rest of the terms in these equations.

In the subsequent derivation, we decompose the whole system into two parts, a background (mean) part and a tidal (perturbation) part. Because the background frequency is much lower than the tidal frequencies, an important assumption is that the background state is independent of time.

b. Available potential energy equations

We decompose the density as $\rho(x, y, z, t) = \rho^m(x, y, z) + \rho'(x, y, z, t)$, where ρ^m denotes the three-dimensional background stratification and ρ' denotes the density perturbation. The density transport equation [Eq. (A4)] can be written as below:

$$\frac{d\rho'}{dt} + \mathbf{u} \cdot \nabla \rho^m = D_\rho. \quad (\text{A5})$$

According to the hydrostatic pressure equation [Eq. (A3)], the pressure can also be decomposed as $p = p^m + p'$, where p^m

and p' denote the background and perturbation components of pressure, respectively:

$$0 = -\frac{\partial p^m}{\partial z} - \rho^m g, \quad (\text{A6})$$

$$0 = -\frac{\partial p'}{\partial z} - \rho' g. \quad (\text{A7})$$

The buoyancy frequency (N^2) and available potential energy (ape) are usually defined as below, referring to Kang and Fringer (2012) as an example.

$$N^2 = -\frac{g}{\rho^m} \frac{\partial \rho^m}{\partial z}, \quad (\text{A8})$$

$$\text{ape} = \frac{1}{2} \rho^m N^2 \eta'^2 = \frac{g^2 \rho'^2}{2 \rho^m N^2}. \quad (\text{A9})$$

To modify the above expressions in the form of a three-dimensional stratification, we introduce a generalized buoyancy frequency (S^2) and a corresponding available potential energy (ape_1), which are written as below:

$$S^2 = \frac{g}{\rho^m} |\nabla \rho^m|, \quad (\text{A10})$$

$$\text{ape}_1 = \frac{1}{2} \rho^m S^2 \eta'^2 = \frac{g^2 \rho'^2}{2 \rho^m S^2}. \quad (\text{A11})$$

Note that η' represents the isopycnal displacement in Eqs. (A9) and (A11).

Then, the APE equation can be derived from Eq. (A5):

$$\frac{d(\text{ape}_1)}{dt} = \rho' g \frac{-\nabla \rho^m}{|\nabla \rho^m|} \cdot \mathbf{u} + D_\rho. \quad (\text{A12})$$

In addition, if we consider a uniform background stratification $\rho^m(z)$, the APE equation turns into Eq. (A13):

$$\frac{d(\text{ape})}{dt} = \rho' g w + D_\rho. \quad (\text{A13})$$

We can conclude from Eqs. (A13) and (A12) that despite diffusivity, only diapycnal movement can generate APE. Compared with a uniform background stratification, under a three-dimensional background stratification, a horizontal movement (u, v) can also generate APE, which is diagnosed in section 4d. For the conciseness of the subsequent derivation, hereafter, we use Eqs. (A9) and (A13) briefly.

c. Energy equations of background flow and tidal flow

We decompose the velocity into background and tidal components; then, Eq. (A1) can be separated into two momentum equations. The superscript m and prime symbol represent the background and tidal states, respectively:

$$\frac{\partial \mathbf{u}_h^m}{\partial t} + \mathbf{u}^m \cdot \nabla \mathbf{u}_h^m + \nabla \cdot (\overline{\mathbf{u}_h' \mathbf{u}'}) = -\frac{1}{\rho_c} \nabla_h p^m + \mathbf{V}^m, \quad (\text{A14})$$

$$\frac{\partial \mathbf{u}_h'}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u}_h' + \mathbf{u}' \cdot \nabla \mathbf{u}_h^m - \nabla \cdot (\overline{\mathbf{u}_h' \mathbf{u}'}) = -\frac{1}{\rho_c} \nabla_h p' - \nabla_h \Omega + \mathbf{V}'. \quad (\text{A15})$$

If we consider tidal flow as a perturbation of the mean flow, then the tensor terms $\mathbf{u}_h' \mathbf{u}'$ can be regarded as the Reynolds stress. With Eqs. (A1), (A13), (A14), and (A15), we can obtain the following energy equations, referring to the derivation of the turbulent kinetic energy (TKE) equations.

Total energy equation:

$$\begin{aligned} \text{Tend} + \text{Div} + \text{Gra} + \varepsilon &= 0, \\ \left\{ \begin{array}{l} \text{tendency term: } \text{Tend} = \frac{\partial}{\partial t} (\text{ke} + \text{ape}) \\ \text{energy flux divergence: } \text{Div} = \nabla \cdot [\mathbf{u}(\text{ke} + \text{ape}) + (\mathbf{u}p) + (\rho_c \mathbf{u}_h' \mathbf{u}')] \\ \text{gravity work: } \text{Gra} = \rho^m g w \end{array} \right. \end{aligned} \quad (\text{A16})$$

Energy equation of background flow:

$$\begin{aligned} \text{Tend}^m + \text{Div}^m - \text{Tran}^m + \text{Gra}^m + \varepsilon^m &= 0, \\ \left\{ \begin{array}{l} \text{tendency term: } \text{Tend}^m = \frac{\partial}{\partial t} (\text{ke}^m) \\ \text{energy flux divergence: } \text{Div}^m = \nabla \cdot [(\mathbf{u}^m \text{ke}^m) + (\mathbf{u}^m p^m) + \mathbf{u}_h^m \cdot (\rho_c \overline{\mathbf{u}_h' \mathbf{u}'})] \\ \text{mean flow-tide interaction: } \text{Tran}^m = (\rho_c \overline{\mathbf{u}_h' \mathbf{u}'}) \cdot \nabla \mathbf{u}_h^m \\ \text{gravity work: } \text{Gra}^m = \rho^m g w^m \end{array} \right. \end{aligned} \quad (\text{A17})$$

Energy equation of tidal flow:

$$\text{Tend}' + \text{Div}' + \text{Tran}' - \rho_c \mathbf{u}'_h \cdot [\nabla \cdot (\overline{\mathbf{u}'_h \mathbf{u}'})] + \text{Gra}' + \varepsilon' = 0, \quad (\text{A18})$$

$$\left\{ \begin{array}{l} \text{tendency term: } \text{Tend}' = \frac{\partial}{\partial t} (\text{ke}' + \text{ape}) \\ \text{energy flux divergence: } \text{Div}' = \nabla \cdot [\mathbf{u}(\text{ke}' + \text{ape}) + (\mathbf{u}'p') + (\mathbf{u}'\rho_c\Omega)] \\ \text{mean flow-tide interaction: } \text{Tran}' = (\rho_c \mathbf{u}'_h \mathbf{u}') \cdot \nabla \mathbf{u}_h^m \\ \text{gravity work: } \text{Gra}' = \rho^m g w' \end{array} \right.$$

In the above equations, kinetic energy is expressed by $\text{ke}^* = \rho_c (\mathbf{u}_h^* \cdot \mathbf{u}_h^*)/2$, and ε^* denotes the dissipation term (including viscosity and diffusivity). Note that due to the periodicity of the tidal system, for a time-averaged tide energy equation, the fourth and fifth terms in Eq. (A18) vanish. Therefore, since tide energy equations are usually averaged over time, we omit $\nabla \cdot (\overline{\mathbf{u}'_h \mathbf{u}'})$ from the tide momentum equation [Eq. (A15)] in advance for the conciseness of the subsequent derivation.

d. Energy equations of barotropic and baroclinic tides

By simplifying the tide momentum equation [Eq. (A15)] and the tide energy equation [Eq. (A18)], the following can be obtained:

$$\frac{\partial \mathbf{u}'_h}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u}'_h + \mathbf{u}' \cdot \nabla \mathbf{u}_h^m = -\frac{1}{\rho_c} \nabla_h p' - \nabla_h \Omega + \mathbf{V}', \quad (\text{A19})$$

$$\begin{aligned} & \frac{\partial}{\partial t} (\text{ke}' + \text{ape}) + \nabla \cdot [\mathbf{u}(\text{ke}' + \text{ape})] + \nabla \cdot (\mathbf{u}'p') \\ & + \nabla \cdot (\mathbf{u}'\rho_c\Omega) + (\rho_c \mathbf{u}'_h \mathbf{u}') \cdot \nabla \mathbf{u}_h^m + \varepsilon' = 0. \end{aligned} \quad (\text{A20})$$

We decompose the tidal velocity $\mathbf{u}'_h$, whole velocity $\mathbf{u}_h$, and pressure perturbation p' into barotropic (BT) and baroclinic (BC) parts. Angle brackets denote a depth average, given as $\langle a \rangle = [1/(\eta + H)] \int_{-H}^{\eta} a \, dz$:

$$\mathbf{u}_h^{\text{bt}} = \langle \mathbf{u}'_h \rangle, \quad \mathbf{u}_h^{\text{w-bt}} = \langle \mathbf{u}_h \rangle, \quad p^{\text{bt}} = \langle p' \rangle, \quad (\text{A21})$$

$$\mathbf{u}_h^{\text{bc}} = \mathbf{u}'_h - \mathbf{u}_h^{\text{bt}}, \quad \mathbf{u}_h^{\text{w-bc}} = \mathbf{u}_h - \mathbf{u}_h^{\text{w-bt}}, \quad p^{\text{bc}} = p' - p^{\text{bt}}. \quad (\text{A22})$$

By integrating Eq. (A19) with respect to depth, one can obtain the BT tide momentum equation [Eq. (A23)]. Note that the Leibniz integral rule is used here. Parameter D denotes the whole water depth and is equal to $(\eta + H)$:

$$\begin{aligned} & \frac{\partial}{\partial t} (D \mathbf{u}_h^{\text{bt}}) + \nabla_h \cdot (D \mathbf{u}_h^{\text{bt}} \mathbf{u}_h^{\text{w-bt}}) + \nabla_h \cdot (D \langle \mathbf{u}_h^{\text{bc}} \mathbf{u}_h^{\text{w-bc}} \rangle) + \int_{-H}^{\eta} \nabla \cdot (\mathbf{u}' \mathbf{u}_h^m) \, dz \\ & = -\frac{1}{\rho_c} [\nabla_h (D p^{\text{bt}}) - p'(\eta) \nabla_h \eta + p'(-H) \nabla_h (-H)] - [\nabla_h (D \Omega) - \Omega \nabla_h \eta + \Omega \nabla_h (-H)] + \mathbf{V}^{\text{bt}}. \end{aligned} \quad (\text{A23})$$

The depth-integrated tide energy equation can be derived from Eq. (A20):

$$\begin{aligned} & \frac{\partial}{\partial t} \int_{-H}^{\eta} (\text{ke}' + \text{ape}) \, dz + \nabla_h \cdot \int_{-H}^{\eta} [\mathbf{u}_h (\text{ke}' + \text{ape})] \, dz + \int_{-H}^{\eta} (\rho_c \mathbf{u}'_h \mathbf{u}') \cdot \nabla \mathbf{u}_h^m \, dz + \nabla_h \cdot \int_{-H}^{\eta} (\mathbf{u}'_h p') \, dz + \nabla_h \cdot \int_{-H}^{\eta} (\mathbf{u}'_h \rho_c \Omega) \, dz \\ & + \rho_c \Omega \frac{\partial \eta}{\partial t} + \varepsilon' = 0. \end{aligned} \quad (\text{A24})$$

The depth-integrated BT tide energy equation can be derived from Eq. (A23):

$$\begin{aligned} & \frac{\partial}{\partial t} (D \cdot \text{ke}^{\text{bt}}) + \nabla_h \cdot (D \mathbf{u}_h^{\text{w-bt}} \cdot \text{ke}^{\text{bt}}) + \nabla_h \cdot (D \mathbf{u}_h^{\text{bt}} \cdot \rho_c \langle \mathbf{u}_h^{\text{bc}} \mathbf{u}_h^{\text{w-bc}} \rangle) + \nabla_h \cdot (D \mathbf{u}_h^{\text{bt}} \cdot p^{\text{bt}}) + \nabla_h \cdot (D \mathbf{u}_h^{\text{bt}} \cdot \rho_c \Omega) - p^{\text{bc}}(\eta) w^{\text{bt}}(\eta) \\ & + p^{\text{bc}}(-H) w^{\text{bt}}(-H) + \rho_c \Omega \frac{\partial \eta}{\partial t} - \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{\text{bt}}) \, dz + \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bt}} \mathbf{u}') \cdot \nabla \mathbf{u}_h^m \, dz + \varepsilon^{\text{bt}} = 0. \end{aligned} \quad (\text{A25})$$

With Eqs. (A24) and (A25), the depth-integrated BC tide energy equation can be obtained:

$$\begin{aligned} & \frac{\partial}{\partial t} \int_{-H}^{\eta} (\text{ke}^{\text{bc}} + \text{ape}) dz + \nabla_h \cdot \int_{-H}^{\eta} [\mathbf{u}_h (\text{ke}^{\text{bc}} + \text{ape})] dz \\ & + \nabla_h \cdot \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} p^{\text{bc}}) dz + p^{\text{bc}}(\eta) w^{\text{bt}}(\eta) - p^{\text{bc}}(-H) w^{\text{bt}}(-H) \\ & + \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{\text{bt}}) dz + \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}') \cdot \nabla \mathbf{u}_h^m dz + \varepsilon^{\text{bc}} = 0. \end{aligned} \quad (\text{A26})$$

By simplifying Eqs. (A24), (A25), and (A26), time-averaged and depth-integrated forms of the tide energy equations can be expressed as follows.

Time-averaged, depth-integrated tide energy equation:

$$\begin{cases} \text{Tran}' - \text{Div}' = \varepsilon', \\ \left\{ \begin{array}{l} \text{mean flow-tide interaction: Tran}' = -\rho_c \int_{-H}^{\eta} (\mathbf{u}_h' \mathbf{u}') \cdot \nabla \mathbf{u}_h^m dz \\ \text{energy flux divergence: Div}' = \nabla_h \cdot \int_{-H}^{\eta} [\mathbf{u}_h (\text{ke}' + \text{ape}) + \mathbf{u}_h' p' + \mathbf{u}_h' \rho_c \Omega] dz \end{array} \right. \end{cases} \quad (\text{A27})$$

Time-averaged, depth-integrated BT tide energy equation:

$$\begin{cases} \text{Tran}^{\text{bt}} - \text{Div}^{\text{bt}} - \text{Conv} = \varepsilon^{\text{bt}}, \\ \left\{ \begin{array}{l} \text{mean flow-BT tide interaction: Tran}^{\text{bt}} = -\rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bt}} \mathbf{u}') \cdot \nabla \mathbf{u}_h^m dz \\ \text{energy flux divergence: Div}^{\text{bt}} = \nabla_h \cdot [D \mathbf{u}_h^{\text{w-bt}} \cdot \text{ke}^{\text{bt}} + D \mathbf{u}_h^{\text{bt}} \cdot \rho_c \langle \mathbf{u}_h^{\text{bc}} \mathbf{u}_h^{\text{w-bc}} \rangle + D \mathbf{u}_h^{\text{bt}} \cdot (p^{\text{bt}} + \rho_c \Omega)] \\ \text{BT-BC tide conversion: Conv} = \int_{-H}^{\eta} \rho' g w^{\text{bt}} dz - \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{\text{bt}}) dz \end{array} \right. \end{cases} \quad (\text{A28})$$

Time-averaged, depth-integrated BC tide energy equation:

$$\begin{cases} \text{Tran}^{\text{bc}} - \text{Div}^{\text{bc}} + \text{Conv} = \varepsilon^{\text{bc}}, \\ \left\{ \begin{array}{l} \text{mean flow-BC tide interaction: Tran}^{\text{bc}} = -\rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}') \cdot \nabla \mathbf{u}_h^m dz \\ \text{energy flux divergence: Div}^{\text{bc}} = \nabla_h \cdot \int_{-H}^{\eta} [\mathbf{u}_h (\text{ke}^{\text{bc}} + \text{ape}) + \mathbf{u}_h^{\text{bc}} p^{\text{bc}}] dz \\ \text{BT-BC tide conversion: Conv} = \int_{-H}^{\eta} \rho' g w^{\text{bt}} dz - \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{\text{bc}} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{\text{bt}}) dz \end{array} \right. \end{cases} \quad (\text{A29})$$

As a complement, by using the hydrostatic pressure equation, the energy conversion term can be written in two

forms, which have already been used in the previous derivation:

$$\begin{aligned}
\int_{-H}^{\eta} \rho' g w^{bt} dz &= \int_{-H}^{\eta} \left(-w^{bt} \frac{\partial p'}{\partial z} \right) dz = \int_{-H}^{\eta} \left(-\frac{\partial p' w^{bt}}{\partial z} + p' \frac{\partial w^{bt}}{\partial z} \right) dz \\
&= \int_{-H}^{\eta} \left(-\frac{\partial p' w^{bt}}{\partial z} + p^{bt} \frac{\partial w^{bt}}{\partial z} + p^{bc} \frac{\partial w^{bt}}{\partial z} \right) dz = \int_{-H}^{\eta} \left(-\frac{\partial p' w^{bt}}{\partial z} + p^{bt} \frac{\partial w^{bt}}{\partial z} \right) dz \\
&= -p'(\eta) w^{bt}(\eta) + p'(-H) w^{bt}(-H) + p^{bt} w^{bt}(\eta) - p^{bt} w^{bt}(-H) \\
&= -p^{bc}(\eta) w^{bt}(\eta) + p^{bc}(-H) w^{bt}(-H) \approx p^{bc}(-H) w^{bt}(-H).
\end{aligned} \tag{A30}$$

e. Discussion of the derivation

The derivation refers mainly to the work in Nagai and Hibiya (2015) and Kang and Fringer (2012). However, the above derivation has some differences from the derivations of previous work.

First, by considering a three-dimensional realistic stratification, we know that horizontal movements can also produce APE along with internal waves. The consideration of background flow leads to an advective effect as well as nonlinear interactions in our derivation. More quantitative diagnostic results can be found in section 4d.

Second, $\rho_c g \eta^2/2$ in the tendency term of the BT tide energy equation vanishes since the surface pressure term ($\rho_c g \eta$) is considered in the pressure perturbation p' to satisfy the ocean surface condition such that $p' = 0$. If we separate the surface pressure from the pressure perturbation term, the ocean surface condition would satisfy $z = 0$ other than $z = \eta$, which is not suitable for integrating with respect to depth from $z = -H$ to $z = \eta$.

Third, when we subtract the BT tide energy equation [Eq. (A25)] from the total tide energy equation [Eq. (A24)], the “cross term” of kinetic energy always exists but is usually neglected after integrating over depth and averaging over time. However, this cross term does not have a physical meaning and is not zero after temporal and spatial averaging. Therefore, a modification is made to Eq. (A31) when deriving the BT tide energy equation [Eq. (A25)] to make it physically reasonable:

$$\begin{aligned}
\rho_c \mathbf{u}_h^{bt} \cdot [\nabla_h \cdot (D(\mathbf{u}_h^{bc} \mathbf{u}_h^{w-bc}))] &= \nabla_h \cdot (D \mathbf{u}_h^{bt} \cdot \rho_c \langle \mathbf{u}_h^{bc} \mathbf{u}_h^{w-bc} \rangle) \\
&\quad - \rho_c \int_{-H}^{\eta} (\mathbf{u}_h^{bc} \mathbf{u}_h) \cdot (\nabla_h \mathbf{u}_h^{bt}) dz.
\end{aligned} \tag{A31}$$

Moreover, Eq. (A32) shows that when deducting the BT tide energy equation [Eq. (A25)] from the total tide energy equation [Eq. (A24)], the cross term of kinetic energy in the advection term counteracts part of the BT tide energy equation. Therefore, the omitted cross term of kinetic energy is actually part of the BT tide energy flux:

$$\begin{aligned}
\nabla_h \cdot \int_{-H}^{\eta} \mathbf{u}_h (k e' - k e^{bt}) dz &= \nabla_h \cdot \int_{-H}^{\eta} \rho_c \mathbf{u}_h (\mathbf{u}_h^{bt} \cdot \mathbf{u}_h^{bc}) dz \\
&= \nabla_h \cdot (D \mathbf{u}_h^{bt} \cdot \rho_c \langle \mathbf{u}_h^{bc} \mathbf{u}_h^{w-bc} \rangle).
\end{aligned} \tag{A32}$$

Note that the radiation stress tensor terms, such as the second term on the right-hand side of Eq. (A31) and the Tran* terms in Eqs. (A27) through (A29), have similar expressions, indicating

the nonlinear effects of different systems. These nonlinear effects lead to the transfer of energy between different flows, which is also mentioned by Chavanne et al. (2010).

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