Seasonality of Tides in Southeast Asian Waters[✉]

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ABSTRACT

An analysis of water level time series from 20 tide gauges in Southeast Asia finds that diurnal and semidiurnal astronomical tides exhibit strong seasonal variability of both amplitude and phase, which is not caused by known modulations of the astronomical tide-generating forces. Instead, it is found that the tidal properties are coherent with the western North Pacific monsoon index (WNPPI), indicating that monsoonal mechanisms are the likely cause. The study domain includes the Malacca Strait, Gulf of Thailand, the southern South China Sea, and Java Sea. The character of the geography and the tidal variability is different in each of these subregions. A new barotropic regional tide model is developed that incorporates the coupling between geostrophic currents, wind-driven (Ekman) currents, and tidal currents in the bottom boundary layer in order to examine the influence of these factors on tides. The dynamics thus preserve the frictional nonlinearities while neglecting advective nonlinearities and baroclinic tides, approximations that should be valid on the wide and shallow continental shelves in the study region. The model perturbation approach uses the climatological seasonal variability of wind stress and geostrophic currents, which are prescribed singly and in combination in the model, to explain the observed tidal variability. Results are most successful in the southern Gulf of Thailand and near Singapore, where it is found that the combined effect of geostrophic and Ekman currents shows increased skill in reproducing the tidal variability than individual models. Ambiguous results at other locations suggest more localized processes such as river runoff.

1. Introduction

Tides exhibit variability on multiple time scales due to changes in physical properties of the ocean and coastal morphology not attributed to astronomical variations. The tidal contribution to total water level is typically significant compared to other processes, such as wind-driven setup, so knowledge of tidal variability is important to understanding and predicting water level extremes and flooding (Devlin et al. 2017a). Previous

work examining yearly and decadal tidal variability has found a significant correlation between mean sea level (MSL) fluctuations and tidal variability at many sites in the Pacific (Devlin et al. 2014, 2017b). In the course of examining these data it was found that the seasonal variability is often larger than the interannual and decadal tidal variability. Hence, the present study specifically examines tidal seasonality, which refers to the annual variability of tidal amplitudes and phases during different months of the year not attributed to predictable variations in the astronomical tide-generating force, in order to gain better insight into the processes leading to the overall variability.

The variation in the tides as prescribed by gravitational forcing is small; for example, there is a predicted variation of 0.7% for M_2 (Hartmann and Wenzel 1995).

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Yet, there are many tide gauge records worldwide that show a seasonal modification in tidal properties (Corkan 1934). Nonastronomical tidal seasonality may be induced by many factors. The interaction of nontidal and tidal currents can alter frictional damping of the tides (Hunter 1975); this may lead to tidal seasonality through, for example, changes in river discharge (Godin 1985, 1991; Guo et al. 2015). Tidal seasonality also can be attributed to other nonlinear interactions arising directly from nontidal changes in water depth, such as occurs during meteorologically driven storm surges (Cartwright and Amin 1986). Another factor is seasonal change in stratification on the continental shelf, which can change the vertical profile of eddy viscosity and, in turn, the vertical current profile (Müller 2012). In Arctic regions and in the Southern Ocean, sea ice dynamics and underice friction can be a significant influence on the seasonality of tidal components (St-Laurent et al. 2008).

Regionally, tides have been found to have a significant seasonality in the North Sea (Pugh and Vassie 1976), attributed to nonlinear surge–tide interaction, which can modify the M_2 amplitude by up to 6% (e.g., at Cuxhaven, Germany; Leeuwenburgh et al. 1999; Huess and Andersen 2001). Additionally, in the North Sea region, the vertical profiles of M_2 currents vary seasonally, owing to changes in stratification (Howarth 1998; van Haren 2000). Foreman et al. (1995) documented seasonal variability of M_2 at Victoria, Canada, suggesting that river flow or seasonal stratification might be the reason. Annual variability of M_2 has also been documented globally using satellite altimetry (Müller et al. 2012).

Kang et al. (1995) described seasonal variations in M_2 of up to 4% of the mean tide near Korea, based on monthly harmonic analyses in February and August. This was attributed to a generalized “interacting effect,” such as meteorological factors or seasonal variations in the strengths of the Kuroshio and the Tsushima Current. More recent modeling studies have found that seasonal stratification can have a pronounced effect on tides, including current shear, frictional dissipation, and barotropic energy flux, which explain the complicated seasonal variability observed in the Yellow Sea (Kang et al. 1998, 2002).

The purpose of the present study is to examine tidal seasonality at 20 gauges in Southeast Asia, a region dominated by the influence of the Asian monsoon (Wyrski 1961; Li and Wang 2005). The monsoon causes spatially variable and seasonal changes in winds, rainfall, ocean surface circulation, stratification, and water mass transport, all of which may in turn affect the tides. As will be shown, below, time series of tidal variability show strong semiannual coherence and moderate annual coherence to the western North Pacific monsoon index (WNPMI; Wang and Fan 1999) in most of the study

domain, suggesting that monsoon-related mechanisms may be a causative factor of tidal seasonality.

The tidal seasonality observed in this region exhibits a superposition of semiannual and annual harmonics, with considerable spatial differences in magnitude and phase. Because the seasonality is generally a robust feature of the tides in this region, it seems both worthwhile and feasible to investigate its causes in detail. Our approach involves using a barotropic tide model coupled to a simplified representation of the three-dimensional dynamics, in which the (nontidal) climatological geostrophic and wind-driven flow are frictionally coupled to the tides, but the baroclinic tidal dynamics are neglected. A monthly climatology of sea level and wind stress is prescribed, using data available from a variety of independent sources, and tidal responses are computed with the model and compared with observations.

As already mentioned, the model does not include the baroclinic tide, which is known to be important to tidal dynamics in the northern South China Sea (SCS; Jan et al. 2007, 2008). The justification for this omission is simply that the sites considered here are all in shallow water where the surface expression of the internal tide is likely very small. The approach also permits higher resolution to be used, in a larger domain, than would be possible if a prognostic three-dimensional baroclinic model were used, and it avoids the need to couple the three-dimensional circulation, stratification, and water level to the surrounding seas. Present efforts to resolve the coupled tidal, wind-driven, and thermohaline circulations (Arbic et al. 2010) are not accurate enough to draw quantitative conclusions about mechanisms at the spatial scales of the present study.

The rest of this article reports on the approach outlined above. First, the observations of tidal seasonality are reported, and these are compared with the monsoon index. Then, the tide model is introduced, including the turbulence submodel, which couples the barotropic tides and nontidal motions. Results of the model and comparisons with the observations are then presented. Last, the results are discussed in terms of the evidence for particular physical mechanisms of tidal seasonality in the region, and further research questions are indicated.

2. Observations of tidal seasonality in the waters of Southeast Asia

The locations of the 20 tide gauges used in the study are shown in Figs. 1 and 2 and detailed in Table 1. The gauges are in locations of varying oceanographic and meteorological conditions, and it is useful to define four subregions as follows. The first subregion is the Malacca Strait (MS) on the west side of the Malay Peninsula, including Singapore, where eight gauges are located. The second subregion is the east side of the Malay

FIG. 1. Map of Southeast Asia, showing the depths (m) of the major ocean basins and tide gauge locations. The region in the green square is enlarged in Fig. 2.

Peninsula and the western coast of the Gulf of Thailand (GOT), with six gauges. The third subregion is the western part of the SCS, with four gauges located in Vietnam and Borneo (Malaysia). The fourth subregion is the Java Sea (JS), with two gauges. The hourly sea level records for these sites were obtained from the archive at the University of Hawai'i Sea Level Center (UHSLC; <https://uhslc.soest.hawaii.edu/>).

The Southeast Asian waters provide the main equatorial connection between the Pacific and Indian Ocean, contain complex bathymetry and sills, and exhibit varied tides. Both the diurnal and semidiurnal tides enter the SCS from the Pacific through the Luzon Strait, generating significant baroclinic energy (Jan et al. 2007, 2008). The amplitude of the semidiurnal constituents are damped as they enter the SCS, though the diurnal constituents are amplified (Zu et al. 2008; Fang et al. 1999). The incoming tidal waves travel through the deep waters of SCS, slow at the shallow Sunda Shelf, and then split into two branches at the coast of the Malay Peninsula, one flowing northward into the GOT and forming amphidromic systems, the other flowing southward into the JS. An atypical clockwise M_2 amphidromic system exists in the GOT, which is due to the near-resonant geometry and topography of the region (Yanagi and Takao 1998).

To remove seasonal variability associated with the predictable astronomical tide, tidal seasonality is expressed in terms of tidal admittances, which are the complex ratios of the harmonic constants of the observed tide versus the equilibrium tide (Devlin et al. 2014). Harmonic analysis of the hourly sea level observations $Z_{\text{obs}}(t)$ is performed using

FIG. 2. Close-up map of the region in the green square in Fig. 1, showing the Strait of Singapore and nearby tide gauges.

the R_T_Tide package (Pawlowicz et al. 2002; Leffler and Jay 2009). Then, the equilibrium tide is computed from the equilibrium tidal potential V , expressed as a sea surface height, $Z_{\text{pot}}(t) = V/g$, where g is the acceleration of gravity (Cartwright and Tayler 1971; Cartwright and Edden 1973), and this is harmonically analyzed using samples at exactly the same locations and times as the observations. Monthly analyses are used to compute time series of admittance amplitude $A(t)$ and phase $\Phi(t)$:

$$A(t) = \frac{A_{\text{obs}}}{A_{\text{pot}}}, \quad \text{and} \quad \Phi(t) = \Phi_{\text{obs}}(t) - \Phi_{\text{pot}}(t), \quad (1)$$

where $(A_{\text{obs}}, \Phi_{\text{obs}})$ and $(A_{\text{pot}}, \Phi_{\text{pot}})$ denote the observed and equilibrium amplitude and phase, respectively. Admittance time series are found for two semidiurnal lunar tides (M_2 and S_2) and two diurnal lunisolar tides (K_1 and O_1), which can be stably resolved from monthly time series. The use of an admittance largely eliminates the predictable seasonal variability due to unresolved frequencies (e.g., the effects of P_1 and K_2 on K_1 and S_2) and allows the examination of factors not involving astronomy, such as seasonal climatic variability.

To highlight the seasonal variability of the monthly MSL and tidal admittance, the time series are projected onto mean, annual, and semiannual cycles by regression onto the functional form:

$$F(t) = a_0 + a_1 \sin(\omega_a t) + b_1 \cos(\omega_a t) + a_2 \sin(\omega_s t) + b_2 \cos(\omega_s t), \quad (2)$$

where a_0 is the mean, (a_1, b_1) are the coefficients of annual variability, and (a_2, b_2) are the coefficients of

TABLE 1. Tide gauges used in this study. Locations are shown in Figs. 1 and 2.

Station	Country	Latitude	Longitude	Start year	End year	Location
MS						
Ko Taphao Noi	Thailand	7.82°N	98.42°E	1985	2014	Bay
Langkawi	Malaysia	6.87°N	99.67°E	1985	2014	Island
Penang	Malaysia	5.42°N	100.35°E	1984	2014	Island
Lumut	Malaysia	4.23°N	100.62°E	1984	2014	Coastal
Kelang	Malaysia	3.05°N	101.37°E	1983	2014	Coastal
Keling	Malaysia	2.22°N	102.15°E	1984	2014	Coastal
Kukup	Malaysia	1.33°N	103.45°E	1985	2014	Island
Tanjong Pagar	Singapore	1.47°N	103.84°E	1984	2014	Harbor
GOT						
Sedili	Malaysia	1.94°N	104.12°E	1986	2014	Bay
Tioman	Malaysia	2.8°N	104.14°E	1985	2014	Island
Kuantan	Malaysia	3.98°N	104.44°E	1983	2014	Coastal
Cendering	Malaysia	5.27°N	103.18°E	1984	2014	Coastal
Getting	Malaysia	6.23°N	102.11°E	1986	2014	Coastal
Ko Lak	Thailand	11.8°N	99.82°E	1985	2014	Coastal
SCS						
Vung Tau	Vietnam	10.34°N	107.07°E	1986	2002	Estuary
Bintulu	Malaysia	3.26°N	113.06°E	1992	2014	Coastal
Kota Kinabulu	Malaysia	5.98°N	116.07°E	1987	2014	Coastal
Sandakan	Malaysia	5.82°N	118.08°E	1993	2014	Coastal
JS						
Jakarta	Indonesia	6.17°S	106.67°E	1984	2004	Coastal
Surabaya	Indonesia	7.22°S	112.73°E	1984	2004	Bay

semiannual variability, with $\omega_a = 2\pi/(1 \text{ year})$ and $\omega_s = 2\omega_a$. The amplitudes of the annual and semiannual harmonics are defined as $M_a = \sqrt{a_1^2 + b_1^2}$, and $M_s = \sqrt{a_2^2 + b_2^2}$, respectively. Tables 2 and 3 show these modulations of the M_2 and K_1 tides, respectively, and Tables S1 and S2 in the supplemental material show O_1 and

S_2 , respectively. Examples of tidal seasonality are shown in Figs. 3 and 4 at a representative station in each subregion. Figure 3 shows the variability at Lumut, Malaysia, in the MS for M_2 (Fig. 3a) and K_1 (Fig. 3c) and at Sedili, Malaysia, in the southern GOT for M_2 (Fig. 3b) and K_1 (Fig. 3d). Figure 4 shows the variability at Vung Tau, Vietnam, in the

TABLE 2. The M_2 amplitude (amp; mm) and phase-fitting (ph; °) harmonics. Mean values of tides are shown (a_0), along with annual harmonics ($M_a = \sqrt{a_1^2 + b_1^2}$) and semiannual harmonics ($M_s = \sqrt{a_2^2 + b_2^2}$), with the associated root-mean-square errors.

Station	a_0 (amp)	M_a (amp)	M_s (amp)	RMSE (amp)	a_0 (ph)	M_a (ph)	M_s (ph)	RMSE (ph)
Ko Taphao Noi	784.0	3.5	0.5	16.8	−63.6	0.2	0.3	2.0
Langkawi	802.8	3.1	3.0	10.7	−42.4	0.2	0.7	0.9
Penang	613.6	2.6	8.8	12.8	−6.5	0.5	0.7	1.0
Lumut	740.9	2.7	3.4	9.7	82.7	0.6	0.3	0.8
Kelang	1368.0	6.6	5.7	20.1	126.8	0.4	0.7	0.9
Keling	607.9	4.9	8.8	9.7	200.6	0.9	0.1	1.4
Kukup	932.5	1.2	16.0	14.3	−61.1	0.8	0.6	0.9
Tanjong Pagar	792.5	24.1	9.0	11.4	−60.4	0.2	0.8	1.5
Sedili	552.1	12.7	19.8	14.5	−92.2	4.4	0.8	1.3
Tioman	588.6	16.6	22.2	16.3	−111.1	3.4	0.9	1.2
Kuantan	522.3	19.7	21.9	15.2	−124.3	2.2	0.9	1.4
Cendering	295.6	7.8	10.3	8.9	−146.9	3.1	0.7	1.5
Getting	162.0	6.0	8.0	11.7	−131.9	3.2	0.4	6.1
Ko Lak	63.2	1.4	5.4	7.6	135.0	4.8	2.2	14.4
Vung Tau	766.4	35.4	22.6	23.5	49.4	0.5	2.2	5.6
Bintulu	189.4	9.4	3.3	8.1	47.5	2.0	1.8	2.4
Kota Kinabulu	233.8	3.7	0.8	4.3	−40.5	0.6	0.4	1.3
Sandakan	415.1	0.8	4.3	12.7	−38.4	1.3	1.3	5.8
Jakarta	62.4	3.5	1.5	8.9	−3.2	27.7	8.3	28.5
Surabaya	344.2	7.0	8.3	17.1	−19.3	1.6	0.5	6.5

TABLE 3. As in Table 2, but for K_1 .

Station	a_0 (amp)	M_a (amp)	M_s (amp)	RMSE (amp)	a_0 (ph)	M_a (ph)	M_s (ph)	RMSE (ph)
Ko Taphao Noi	136.6	9.6	4.0	6.4	−36.8	4.3	2.0	3.0
Langkawi	176.2	12.2	5.6	7.6	−28.3	5.4	1.7	2.8
Penang	208.6	19.4	5.6	10.0	−22.6	7.2	1.1	3.0
Lumut	228.0	17.3	12.8	13.3	−12.2	8.4	2.5	3.6
Kelang	202.1	24.7	15.5	16.9	6.5	11.8	4.3	5.6
Keling	103.5	33.8	14.0	16.8	−224.5	21.7	9.8	14.0
Kukup	273.2	19.8	11.5	13.6	−213.6	9.3	3.9	2.9
Tanjong Pagar	319.9	7.6	16.4	13.3	95.8	5.7	3.0	3.2
Sedili	374	10.0	12.9	13.0	30.1	2.8	0.9	2.2
Tioman	488.3	9.9	14.5	12.3	7.2	1.1	1.4	2.0
Kuantan	561.7	12.8	19.9	12.5	−3.8	0.5	1.4	1.9
Cendering	522.6	12.7	21.8	12.2	−14.3	0.4	1.5	2.0
Getting	262.5	15.5	19.2	14.4	−23.6	1.6	2.5	4.0
Ko Lak	530.1	9.6	24.1	21.0	−199.0	0.9	0.8	3.2
Vung Tau	615.5	20.0	18.0	17.2	−44.5	1.6	1.5	3.8
Bintulu	434.7	7.5	8.7	19.5	−47.7	1.9	1.5	1.7
Kota Kinabulu	371.7	2.1	10.4	7.2	−50.2	1.0	1.8	1.2
Sandakan	368.1	6.3	7.0	9.3	−37.3	2.7	0.4	1.6
Jakarta	268.1	20.8	16.2	11.6	−38.4	3.8	1.4	4.7
Surabaya	485.7	14.2	22.0	14.6	138.7	1.9	0.6	3.3

SCS for M_2 (Fig. 4a) and K_1 (Fig. 4c) and at Surabaya, Indonesia, in the JS for M_2 (Fig. 4b) and K_1 (Fig. 4d). The amplitude anomaly is displayed in dimensional units (mm) by multiplying the admittance by the mean amplitude of the equilibrium tide and subtracting the mean.

Seasonal variability in the southern part of the GOT is generally larger than the other regions of the study domain for semidiurnal amplitudes (Figs. 3b,d). The M_2 amplitudes exhibit nearly equal semiannual and annual magnitudes and phase variability of 5° or more (not shown). The K_1 amplitudes exhibit a more varied mix of annual and semiannual magnitudes, but the phase variability is generally mild or nonexistent. In the MS the variability of the diurnal tides is larger than that of the semidiurnal tides. The K_1 variability generally shows a pulselike pattern of mixed annual and semiannual behavior, with very steep transitions occurring around August/September and around February/March at most gauges (Figs. 3a,c). The O_1 amplitudes vary semiannually by as much as 25 mm (which is nearly 50% of the mean tide), and O_1 phases also have large semiannual variations, up to 40° . In the SCS the seasonal variability is typically small (10 mm or less), except at Vung Tau, where seasonal variability is large (200 mm or more) and annual and semiannual harmonics are comparable. In the JS the M_2 variability is weak but still identifiable. In contrast, the K_1 seasonality is large at both JS gauges. Phase variations are generally small and insignificant in the JS.

The connection of the observed tidal amplitudes to the monsoon (represented by the WNPMI) is highlighted in Fig. 5. Figure 5a shows a 10-yr (1990–2000) time series of M_2 amplitude at Sedili (blue line) along with the WNPMI (green line). The monsoon index has a clear annual cycle.

The tidal amplitudes, as was seen in Figs. 3 and 4, have more pronounced semiannual behavior, with positive but unequal peaks during the maximum positive and negative phases of the monsoon index. This yields a correlation between the tidal amplitudes and the monsoon at both annual and semiannual frequencies, with the semiannual correlations typically stronger than the annual. This is demonstrated by the coherence of M_2 admittance with the WNPMI, shown in Fig. 5c. The coherence values at all stations are mapped in Fig. 5b (semiannual) and Fig. 5d (annual). The semiannual coherence of M_2 with the WNPMI is largest in the southern GOT and SCS and moderate in the MS and the JS. The annual coherence M_2 with the WNPMI is weak or moderate at all sites. Coherence values for all stations and analyzed tides are listed in Tables S3 and S4 for the annual and semiannual coherence of tidal amplitudes and Tables S5 and S6 for tidal phases.

The above highlights the main characteristics of tidal seasonality at the sites studied. The variability is correlated with a large-scale meteorological process, the monsoon, but there are a range of seasonal mechanisms associated with the monsoon that might cause the variability. The large-scale correlations of the variability throughout the region indicate that it is not likely related to the baroclinic internal tide, which is spatially coherent over very small distances. Also, the magnitudes of the tidal changes, which are on the scale of centimeters of sea surface height, cannot easily be supported by internal tides, at least not within the shallow GOT, JS, and MS. For example, in the 60-m mean depth of the GOT, because of the relatively small surface-to-bottom temperature difference, a 1-cm steric height anomaly

FIG. 3. Examples of observed tidal variability at representative tide gauges. Seasonality of (a) M_2 and (c) K_1 admittance amplitude at Lumut in the MS. Seasonality of (b) M_2 and (d) K_1 at Sedili, in the GOT. Dots show the admittance anomaly that is obtained from each monthly admittance minus the annual mean admittance. The red line shows the annual and semiannual fit to the seasonality [Eq. (2)]. Dimensional units (mm) are obtained by multiplying the nondimensional admittance anomaly by the annual mean of the equilibrium tidal amplitude.

would correspond to internal isotherm displacements equal to or exceeding the full water column depth (Yanagi et al. 2001). Instead, we seek other mechanisms for the tidal seasonality.

3. A model for barotropic tides coupled to nontidal stratification and currents

It is hypothesized that frictional interactions between the barotropic tides and nontidal processes are responsible for the observed tidal seasonality. In this section a numerical model is developed to test this hypothesis. The model consists of an approximation to the horizontal momentum equations describing the nontidal ocean currents driven by wind stress and barotropic pressure gradients that are coupled to barotropic tides via the bottom stress. The seasonal cycles of wind stress and barotropic pressure gradient, as estimated from observation-based climatologies, are then applied both individually and in combination with the model to determine their influence on tides. The nonlinearity of the turbulent stresses makes the

dynamics nontrivial, and it necessitates an iterative scheme to couple the tidal and subtidal flow fields.

More precisely, the horizontal momentum balance of the model is given by

$$\frac{\partial \mathbf{u}}{\partial t} + f \hat{\mathbf{k}} \times \mathbf{u} = -g \nabla \left(\eta + \int_z^0 \rho g / \rho_0 ds + \tilde{V}/g \right) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}}{\partial z} \right), \quad (3)$$

where $\mathbf{u} = (u, v)$ is the horizontal current vector; a function of latitude θ , longitude ϕ , depth z , and time t ; $f = 2\Omega \sin\theta$ is the Coriolis parameter, where Ω is the rotation rate of Earth; η is the elevation of the water surface, relative to the undisturbed depth H ; ρ is the seawater density and ρ_0 is its mean value; and ν is the vertical component of the turbulent viscosity. The $\hat{\mathbf{k}}$ indicates the unit vector in the vertical direction. The $\tilde{V}$ is a modified form of the equilibrium astronomical tidal potential, which equals V plus a correction for tidal self-attraction and solid-Earth loading, following Egbert and Erofeeva (2002). Boundary conditions

FIG. 4. As in Fig. 3, but for Vung Tau, in the SCS, and Surabaya, in the JS.

on $\mathbf{u}$ are no slip at the bottom, $\mathbf{u} = 0$ on $z = -H$, and prescribed stress at the ocean surface given by $\nu \partial \mathbf{u} / \partial z = \boldsymbol{\tau}_w$, where $\boldsymbol{\tau}_w$ is the wind stress vector divided by ρ_0 . Thus, the hydrostatic and Boussinesq approximations have been made, and the advective nonlinearities are neglected; the remaining nonlinearity in the model is in the specification of ν , which shall be described below.

The total current is assumed to be the sum of a barotropic tidal component $\mathbf{u}_1$, which is time variable, and a subtidal (steady) component $\mathbf{u}_2$, which is diagnosed from a prescribed pressure gradient and wind stress. Substituting $\mathbf{u} = \mathbf{u}_1 + \mathbf{u}_2$ into Eq. (3), it is straightforward to obtain the following:

$$\begin{aligned} \frac{\partial \mathbf{u}_1}{\partial t} + f \hat{\mathbf{k}} \times \mathbf{u}_1 = & -g \nabla \left(\eta_1 + \int_z^0 \rho_1 g / \rho_0 ds + \tilde{V} / g \right) \\ & + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_1}{\partial z} \right) + f \hat{\mathbf{k}} \times \mathbf{u}_2 \\ & - g \nabla \left(\eta_2 + \int_z^0 \rho_2 g / \rho_0 ds \right) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_2}{\partial z} \right). \end{aligned} \quad (4)$$

Note that the turbulent viscosity is a function of both the tidal and subtidal fields: $\nu = \nu(\rho_1, \rho_2, u_1, u_2)$. Next,

neglect the baroclinic tide (i.e., assume the gradient of the ρ_1 term is negligible compared to the gradient of the η_1 term):

$$\begin{aligned} \frac{\partial \mathbf{u}_1}{\partial t} + f \hat{\mathbf{k}} \times \mathbf{u}_1 = & -g \nabla (\eta_1 + \tilde{V} / g) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_1}{\partial z} \right) \\ & + f \hat{\mathbf{k}} \times \mathbf{u}_2 - g \nabla \left(\eta_2 + \int_z^0 \rho_2 g / \rho_0 ds \right) \\ & + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_2}{\partial z} \right). \end{aligned} \quad (5)$$

The turbulent viscosity is given by a Prandtl mixing-length model, $\nu = \ell^2 |\partial \mathbf{u} / \partial z|$, where the mixing length ℓ is a function of the shear and stratification according to a modified law-of-the-wall relation. The nominal law-of-the-wall mixing length is given by

$$\ell^* = -\kappa z (1 + z/H), \quad (6)$$

where the von Kármán constant is $\kappa = 0.41$, which describes the structure of the turbulence in unstratified levels near the ocean bottom and the surface (Hinze 1975); note that ℓ^* is positive for $z \in [-H, 0]$. Stratification limits the maximum mixing length to the

FIG. 5. Connections of tidal amplitudes to the monsoon index (WNPMI). (a) A 10-yr (1990–2000) time series of M_2 amplitude at Sedili (blue) along with the WNPMI (green). The (b) semiannual and (d) annual correlations of M_2 amplitudes with the WNPMI at all stations are mapped. The color of the dots indicates the correlation value found at the annual frequency band; light blue is a correlation of 0–0.25, dark blue is 0.25–0.50, green is 0.5–0.75, and red is greater than 0.75. (c) The correlation of M_2 admittance with the WNPMI is shown, giving the calculated correlations with pronounced peaks at the annual and semiannual frequency bands.

Ozmidov scale, $\ell_o = (\varepsilon N^{-3})^{1/2}$, which is expressed in terms of the dissipation rate, $\varepsilon = \nu(\partial|\mathbf{u}|/\partial z)^2$, and the buoyancy frequency N . The nominal mixing length is then altered to give a value corresponding to the modified law of the wall (Perlin et al. 2005):

$$\ell^{**} = \frac{\ell^*}{1 + \ell^*/\ell_o}, \quad (7)$$

which smoothly interpolates between $\ell^{**} = \ell^*$ and $\ell^{**} = \ell_o$ in the limits $\ell^* \rightarrow 0$ and $\ell^* \rightarrow \infty$, respectively. Finally, outside the boundary layer, the above mixing length is further modified by the stratification and shear according to

$$\ell = \frac{\ell^{**}}{(1 + \alpha \text{Ri})^2}, \quad (8)$$

where the gradient Richardson number is given by $\text{Ri} = N^2(\partial|\mathbf{u}|/\partial z)^{-2}$ and α is a constant. As summarized recently by Forryan et al. (2013), a variety of functions

have been proposed to model the Richardson-number dependence of shear-driven stratified mixing. The numeric constant used in Eq. (8), $\alpha = 4.7$, is borrowed from a model of stratified turbulence used in numerical weather prediction (Louis 1979); it is essentially identical to the value used in ocean models (Pacanowski and Philander 1981; Yu and Schopf 1997).

To proceed, note that the dynamics of $\mathbf{u}_1$ and $\mathbf{u}_2$ can be separated:

$$\frac{\partial \mathbf{u}_1}{\partial t} + f \hat{\mathbf{k}} \times \mathbf{u}_1 = -g \nabla(\eta_1 + \tilde{\mathbf{V}}/g) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_1}{\partial z} \right), \quad \text{and} \quad (9)$$

$$f \hat{\mathbf{k}} \times \mathbf{u}_2 = -g \nabla \left(\eta_2 + \int_z^0 \rho_2 g / \rho_0 ds \right) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}_2}{\partial z} \right), \quad (10)$$

which shows how, at this level of approximation, the tidal and subtidal dynamics are coupled only through ν .

FIG. 6. Seasonal fields of geostrophic velocity $\mathbf{u}_g$ variability (m s^{-1}), assimilated from AVISO. Four months are shown: (a) January, (b) April, (c) July, and (d) October.

Next, vertically integrate Eq. (9) from $z = -H$ to $z = \eta$, assuming zero stress at the water surface and neglect small nonlinearities proportional to the product $\mathbf{u}_1\eta_1$ to obtain (Egbert et al. 1994; Egbert and Erofeeva 2002)

$$\frac{\partial \mathbf{U}_1}{\partial t} + f\hat{\mathbf{k}} \times \mathbf{U}_1 = -gH\nabla(\eta_1 + \tilde{V}/g) - \tau_b. \quad (11)$$

Note that $\tau_b = \nu \partial \mathbf{u}_1 / \partial z$ at $z = -H$. Equation (10) is solved with no-slip boundary conditions at the bottom, $\mathbf{u}_2 = 0$, and prescribed wind stress at the ocean surface, $\tau_w = \nu \partial \mathbf{u}_2 / \partial z$.

At this point, the strong assumption is made that the turbulent viscosity ν can be determined independently of the tidal current. In practice, the subtidal momentum in Eq. (10) is solved iteratively by prescribing the forcing fields (η_2 , ρ_2 , τ_w) and an initial guess of ν . Subsequent iterates update ℓ and ν using the $\mathbf{u}_2$ field calculated, keeping η_2 , ρ_2 , and τ_w fixed.

The stress of the subtidal flow adds to the bottom stress experienced by the tidal flow. Let the tidal flow be represented as the sum of tides at a finite and discrete set of frequencies $\{\omega_i\}$ for $i = 1, \dots, N$:

$$\mathbf{u}_1 = \sum_{i=1}^N \mathbf{u}^{(i)}. \quad (12)$$

As analyzed in Snyder et al. (1979), when the velocity field consists of a nontidal flow plus a sum of tidal flows at distinct frequencies, the linearized quadratic bottom stress experienced by a particular tidal component can be represented as

$$\tau_b^{(i)} = C_d(r_0 + r_1 u_f + r_2 u_f^2 + \dots) \mathbf{u}^{(i)}, \quad (13)$$

where C_d is a constant drag coefficient term, and u_f is a scalar friction velocity, given by

$$u_f = \left[|\mathbf{u}_2|^2 + \sum_{i=1}^N |\mathbf{u}^{(i)}|^2 \right]^{1/2}. \quad (14)$$

For any particular set of tidal frequencies and amplitudes, the coefficients r_j in Eq. (13) can be found from an asymptotic expansion, but since there is considerable uncertainty in the numeric value of C_d , the r_j coefficients are best thought of as adjustable parameters, where the main couplings among the tides are described by the first few terms in the expansion (Snyder et al. 1979).

FIG. 7. Seasonal fields of Ekman velocity $\mathbf{u}_e$ variability (m s^{-1}), calculated using data assimilated from ECMWF. Four months are shown: (a) January, (b) April, (c) July, and (d) October.

This simplification of the bottom drag is used to model the tidal flows in the present case. For each tidal frequency ω_i the tide model is given by Eq. (11) combined with the stress model of Eq. (13):

$$\frac{\partial \mathbf{U}^{(i)}}{\partial t} + f \hat{\mathbf{k}} \times \mathbf{U}^{(i)} = -gH \nabla [\eta^{(i)} + \tilde{V}^{(i)}/g] - C_d u_f \mathbf{U}^{(i)}/H, \quad (15)$$

and the continuity equation, given by

$$\frac{\partial \eta^{(i)}}{\partial t} + \nabla \cdot \mathbf{U}^{(i)} = 0. \quad (16)$$

Note that each tidal frequency is coupled through the friction velocity u_f , so, in practice, the solution of Eqs. (15) and (16) is iterative; an initial guess for u_f is specified ($u_f = 1 \text{ m s}^{-1}$), then the $\mathbf{U}^{(i)}$ fields are computed, and these are used to compute u_f with Eq. (14), and the iteration continues. Typically three to seven iterations are needed to obtain convergence within a few percent, and seven iterations are used in all experiments here. The numerical implementation is built as a modification to

the tidal solver in the Oregon State University Tidal Inversion Software (OTIS; Egbert et al. 1994; Egbert and Erofeeva 2002).

The model described above was implemented in a domain bounded by 15°S and 15°N, 90° and 120°E, using a resolution of 1/30°, with sea-floor topography H obtained from the Digital Bathymetric Data Base (DBDB2) (D.-S. Ko 2010, personal communication). Tidal elevations at open boundaries were taken from the global TPXO data-assimilating tide model (Egbert and Erofeeva 2002), which are very accurate in the deep ocean (Stammer et al. 2014). To isolate the influence of domain size and boundaries on the tidal solutions, the tide model was calibrated against the observed mean tides for the case $\mathbf{u}_2 = 0$. Optimal values of $C_d = 2 \times 10^{-3}$, $r_0 = 0$, and $r_1 = 0.5$ were found, which correspond to a minimum root-mean-square error (rmse) of 0.08 m. Experiments were performed to identify r_i for $i > 2$, but optimizing these higher-order terms led to very little reduction in the rmse, so $r_i = 0$ is used for $i > 2$. The northern boundary of the model domain passes through the deep middle basin of the SCS where the TPXO

FIG. 8. Time series of all relevant seasonal variability in physical properties at four locations in the model domain: Vung Tau in the SCS; Jakarta in the JS; Sedili in the GOT; and Lumut in the MS. (top) Ekman velocity u_e variability, and (bottom) geostrophic velocity u_g . Note that different vertical scales are used.

boundary conditions are very accurate. Experiments with larger computational domains found that the solutions were insensitive to the precise locations of the open boundaries, so long as they were located in deep water. Examples of the unperturbed calibration solutions are not shown here but are provided in the supplemental material as Figs. S1–S4 for M_2 , S_2 , K_1 , and O_1 , respectively.

The numerical model for the subtidal flow, Eq. (10), is solved on the same horizontal grid as the tide model, but it uses a high-resolution (1 m) grid in the vertical. The high-resolution vertical grid is necessary to resolve the bottom boundary layer and variability of ν throughout the water column.

4. Data sources for the subtidal currents

As already mentioned, the subtidal currents diagnosed from Eq. (10) are forced by the barotropic pressure gradient $-g\nabla\eta_2$, the surface wind stress τ_w , and

the baroclinic pressure gradient $(-\nabla\int\rho_2g/\rho_o dz)$, though only the first two of these are given focus in this study. The data used to compute each of these quantities is described in this section.

The barotropic pressure gradient is computed from maps of sea surface height provided by Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVISO; <http://www.aviso.altimetry.fr/duacs/>). These maps, which combine data from multiple satellite altimeters, provide estimates of the absolute dynamic topography (i.e., sea level relative to the marine geoid) at weekly time resolution and $1/4^\circ$ spatial resolution (Dibaroure et al. 2011; DUACS/AVISO team 2014). This spatial resolution is much coarser than the model grid, so the maps are extrapolated to the coastline, as necessary, using a relaxation technique to solve $\nabla^2\eta_2 = 0$ on these nodes. These data have been averaged during the 1992–2015 period to form a climatology of monthly mean sea level η_2 . Figure 6 shows the current fields

FIG. 9. Modeled adjustments of the M_2 tidal amplitudes due to the seasonal changes in geostrophic velocity $\mathbf{u}_g$ (mm) in (a) January, (b) April, (c) July, and (d) October. We display the difference field as red showing positive departures from the annual average at that location, blue indicating negative departures from the annual average, and white regions showing insignificant or near-zero change from the average; see the color-bar scale to indicate the magnitude of significance.

obtained from the climatology to illustrate the main features of the barotropic seasonal circulation. These currents are essentially in geostrophic balance except for very near the equator where $f \rightarrow 0$ and bottom friction enters the dynamical balance. The seemingly disconnected region at 5° N and 5° S is due to the processing done by AVISO, in which geostrophic current velocities are computed in a 5° band across the equator, using second-derivative methods based on Lagerloef et al. (1999), in which altimetry-based geostrophic currents were calibrated against in situ drifter observations, done to ensure continuity with classical geostrophy.

The wind-driven Ekman component of $\mathbf{u}_2$ is computed from weekly wind stress data provided by ERA-Interim (Dee et al. 2011), which is averaged in monthly windows, and the speed of the surface Ekman current is shown in Fig. 7. It can be seen that Ekman currents are of significant magnitude during monsoon periods (January and July) and relatively unimportant at other times.

Note that although the above forcing components are specified separately, the subtidal currents are coupled through the vertical turbulent viscosity. Away from the equator, though, the viscous term is a small part of the term balance, and the currents essentially just add linearly. Close to the equator, within about 1° , the currents are strongly influenced by friction, but here the validity of the model is doubtful since the nonlinear terms involving relative vorticity and the horizontal eddy fluxes are likely to be significant. Fortunately, the equatorial strip involves such a small fraction of the domain that its influence on the tides is negligible. Hence, the components of $\mathbf{u}_2$ can essentially be considered as decoupled in the present application.

Although the barotropic and baroclinic pressure gradients have been referred to separately, they are not unambiguously separable since the observed sea surface height η_2 is the combined result of the baroclinic and barotropic pressure at the ocean surface. Henceforth,

FIG. 10. As in Fig. 9, but for modeled adjustments of the K_1 tidal amplitudes due to the seasonal changes in geostrophic velocity $\mathbf{u}_g$.

the barotropic and baroclinic components of the subtidal current shall be lumped together and referred to as the geostrophic current $\mathbf{u}_g$ since it is nearly in geostrophic balance. This should be justifiable in our shallow water domain, which likely has a very small baroclinic pressure gradient over most of the study area. The subtidal current driven by the wind stress shall be referred to as the Ekman current $\mathbf{u}_e$.

To highlight the differences in the basic patterns of seasonal variability within each subregion, the monthly values of u_g and u_e are shown at a representative subset of tide gauge locations in Fig. 8. Values of the fields are shown for stations within each subregion: Vung Tau in the SCS (first column), Jakarta in the JS (second column), Sedili in the GOT (third column), and Lumut in the MS (last column). The variability of $\mathbf{u}_g$ is annual in the JS and GOT; both regions have extended minima during November to March, with a maximum during August in the JS and in May in the GOT. However, the magnitude of $\mathbf{u}_g$ in the JS is about twice that of the GOT. In the SCS and within the MS, geostrophic velocities are

of the same magnitude and mainly semiannual, with a greater maximum in December/January and a lesser maximum in June/July. Both the SCS and the MS have weak minima in spring and fall months. The variability of $\mathbf{u}_e$ is mixed but mainly semiannual in the SCS, the JS, and the GOT, with larger maxima in December/January and lesser maxima in June or July. Only within the MS is this pattern different, being of small magnitude and mainly annual, with a maximum from June to August and lower values seen at most other times. During April and October, when the monsoon winds have subsided, $\mathbf{u}_e$ is negligible at most locations.

5. Can tidal seasonality be explained by frictional coupling of tides and subtidal currents?

The influence of seasonal variability of the subtidal currents on the tides has been examined by solving Eqs. (10), (15), and (16), which include the mixing length model [Eqs. (6)–(8)] and the bottom stress model [Eqs. (13) and (14)], given the monthly climatologies of

FIG. 11. As in Fig. 9, but for modeled adjustments of the M_2 tidal amplitudes due to the seasonal changes in Ekman velocity $\mathbf{u}_e$.

subtidal forcings described in section 4. The influences of the geostrophic and Ekman currents $\mathbf{u}_g$ and $\mathbf{u}_e$ are examined separately and in combination. For all model results, the yearly averaged model tide is subtracted from the monthly tide to produce a monthly anomaly. Twelve monthly results are produced for each of the four modeled constituents and for each of the mechanisms investigated ($\mathbf{u}_g$ and $\mathbf{u}_e$). In the figures, four months are shown (January, April, July, and October) to have a convenient representation of the seasonal monsoon and transition periods.

Figure 9 shows the M_2 amplitude anomaly caused by the geostrophic velocity $\mathbf{u}_g$. In the eastern SCS, there are slightly lowered amplitudes in January, April, and October, but near the Mekong in the western SCS, there are increased tidal amplitudes at these times and lowered amplitudes in July. In the GOT, maximum positive amplitude departures are observed in April and maximum negative departures are in July. The behavior of tidal amplitudes in the MS are mixed, with a steep and reversing gradient within a narrow geographical area.

The maximum changes are seen near the tip of the Malay Peninsula. The northern JS near the southern coast of Borneo shows strong positive fluctuations in April and negative fluctuations in October; unfortunately, there are no tide gauges in this area to validate the model here.

Figure 10 shows the K_1 amplitude anomaly caused by $\mathbf{u}_g$. Changes in the diurnal tide are weakly semiannual in the majority of the GOT, with greater changes in the northern and southern parts of the Gulf. In the northern MS, amplitude changes are positive in January and negative at other times; in the southern MS, the pattern is reversed. The northwestern and eastern JS show broad areas of strong seasonal change, with strong maxima in April and October. Plots for S_2 and O_1 are provided in the supplemental material (Figs. S5 and S6).

Figure 11 shows the response of M_2 to the Ekman velocity $\mathbf{u}_e$. Ekman forcing is only relevant during times of persistent winds (i.e., during the northeast monsoon in winter and the southwest monsoon in summer); therefore, any modifications to the tidal amplitudes are

FIG. 12. As in Fig. 9, but for modeled adjustments of the K_1 tidal amplitudes due to the seasonal changes in Ekman velocity $\mathbf{u}_e$.

only apparent during January and July. The spatial pattern is nearly identical in both windy seasons, with negative adjustments in the GOT and in the northwest and eastern JS and positive adjustments in the western SCS and central JS. Figure 12 shows the K_1 changes caused by $\mathbf{u}_e$. Again, significant adjustments are found only in January and July, and the entire shallow water region undergoes negative amplitude adjustments from annual averages, with the exception of the tip of the Malay Peninsula, which has a slight positive adjustment. Plots for S_2 and O_1 are provided in the supplemental material (Figs. S7 and S8).

We now compare the tide model perturbations (averaged over a window of 3-by-3 grid cells nearest to the geographical location of each tide gauge) to the observed variability at each individual gauge. This is performed first for the $\mathbf{u}_g$ and $\mathbf{u}_e$ processes separately and then for their combination ($\mathbf{u}_g + \mathbf{u}_e$). We first show plots of the model responses at representative stations in each subregion as compared to the observed data and the fitted line as provided by Eq. (2), an expansion of that shown in Figs. 3 and 4. For ease of viewing, the three

individual model results are plotted in one figure, and the four combinations of model responses are plotted in a separate figure. Figure 13 shows the M_2 and K_1 amplitude variability at Lumut and Sedili in the MS and GOT, respectively, for $\mathbf{u}_g$ and $\mathbf{u}_e$, both individually and for the combination of both processes. The observed M_2 seasonality at Lumut (Fig. 13a) is quite small, as are the model responses, but the combined model recreates some of the observations. The K_1 seasonality is more significant (Fig. 13c), with a steep transition in the latter half of the year, and the combined model comes closest to recreating the observations. At Sedili, the M_2 and K_1 seasonalities (Figs. 13b,d) are partially reconstructed by each of the individual models, but for both cases, a better fit is provided by the combined model.

Figure 14 shows the results of the individual and combined processes at Vung Tau in the SCS and at Surabaya in the JS. The M_2 seasonality at Vung Tau (Fig. 14a) is moderate recreated by each of the individual models as well as the combined model, though none of them find a perfect fit; in particular, the peak in the second half of the year is not captured adequately,

FIG. 13. Model responses compared to observed tidal variability (mm) at regional representative tide gauges. The seasonality of the (a) M_2 and (c) K_1 admittance amplitude in the MS at Lumut. The (b) M_2 and (d) K_1 admittance amplitude in the Gulf of Thailand at Sedili. In all plots, the raw observed data is shown as black scatter points, binned by the day of year and detrended from the mean tidal amplitude. The fitting algorithm of Eq. (2) is shown as a solid red line. The Ekman velocity $\mathbf{u}_e$ response (blue), the geostrophic velocity $\mathbf{u}_g$ response (light blue), and the combined model of $\mathbf{u}_e + \mathbf{u}_g$ (green) are also shown.

with the model responses being delayed from the observed variability. The K_1 model responses are mainly small and fail to capture the strong peak occurring later in the year. Both of these mismatches are likely due to river effects of the Mekong River, whose dynamics are not sufficiently resolved in our model. Finally, in Figs. 14b,d, we show the variability at Surabaya, which is less successful in the model, particularly for K_1 , where the strong observed semiannual behavior is not recreated well.

Next, to determine the skill of the models in recreating the observed tidal variability, linear regression statistics are calculated between the complex amplitudes of detrended observed variability and the detrended model responses. There are 12 calculations of amplitude and phase for each model at each location, one for each month. The in-phase and quadrature parts of $Z(t)$ are separated and combined into a single dataset of 24 values for each observed constituent $Z_{\text{obs}}(t)$ and for each model response $Z_{\text{mod}}(t)$. A robust linear regression is then applied between the observed variability and each

of the three model results, with the resultant quantities denoted tidal reconstruction factors (TRFs), in dimensionless units, given by

$$\text{TRF} = \frac{Z_{\text{obs}}(t)}{Z_{\text{mod}}(t)}.$$

TRF values and the associated error bounds express how well each model recreates the observed variability. A value of 1.0 would indicate a perfect correspondence, values greater than 1.0 indicating that the model overestimates the observed variability, and values from 0 to 1.0 indicating that the model underestimates the observed variability. Negative TRF values indicate that the model response is inverted from that of the observed variability, with a value of -1.0 indicating perfect inversion. Comparisons are made in each of the seven model combinations for each tidal constituent at each gauge. Tables 4 and 5 report the M_2 and K_1 TRF values and errors (95% confidence intervals). Positive TRF values greater than

FIG. 14. As in Fig. 13, but for (a),(c) Vung Tau in the SCS and (b),(d) Surabaya in the JS.

0.1 and having a signal-to-noise ratio (SNR) greater than 2.0 are in bold text in the tables, and negative (inverted) TRF are italicized. TRFs for S_2 and O_1 are reported in Tables S7 and S8 in the supplemental material. In addition to the tables, maps are provided that show the calculated TRF for each model run at each tide gauge; Fig. 15 shows the M_2 TRFs, and Fig. 16 shows K_1 TRFs.

The TRF values for M_2 (Table 4; Fig. 15) in the MS show that three of eight stations were fit by the combination of Ekman and geostrophic velocities, though overestimated by a factor of 1.25 to 1.93. The Ekman contribution is very small in the strait, hence the insignificant results for the Ekman forcing alone. The geostrophic model result for Singapore is notable, with a TRF of +1.33; it recreates quite well the strong annual cycle in the M_2 amplitudes observed there. In the GOT, all models fit most gauges to some extent, but the combined model was generally a better fit. In the northern GOT at Ko Lak, Thailand, the Ekman forcing alone was the best fit, with a TRF value of 0.65. In the central GOT, responses are significant but somewhat weaker. In the SCS, only Ekman forcing gave a good fit of M_2 seasonality at Vung Tau and Bintulu, Malaysia. There is a good recreation at Surabaya for the combined

velocity models. Very large model responses are observed at Vung Tau, and the Ekman response yielded a TRF value of 0.75 with a small relative error.

The K_1 TRFs (Table 5; Fig. 16) in the northern MS are weak for the geostrophic model response. As was seen for M_2 , the Ekman response is mainly negligible here. Moderately accurate positive reconstructions are found at Sedili and Tioman, Malaysia, for the Ekman model response and at Genting, Malaysia, for the combined velocity model outputs. In the SCS, Vung Tau's large K_1 tidal variability is not well explained by any model output combination (which, as mentioned above, might be modified by the Mekong River), and elsewhere in the SCS and the JS, model responses are negligible with the exception of Bintulu, which has moderate negative responses in most models. Finally, in the JS, weak positive reconstructions are seen in the combined velocity results.

In summary, the M_2 models can recreate the observed tidal variability at 14 of 20 tide gauges (Fig. 15), with TRF values running from 0.25 through 2.66 (Table 4). The K_1 models recreate the observed variability at 11 gauges (Fig. 16) with TRFs ranging from 0.11 to 0.73 (Table 5). For S_2 , nine gauges are adequately recreated, with TRF values running from 0.24 to 1.90. Finally, for O_1 ,

TABLE 4. M_2 model responses. Values shown indicate the slope relation between the complex observed tidal seasonality and the complex model responses, denoted TRFs, in dimensionless units. TRFs are shown the individual and combined model response, along with the associated errors of the slope determinations. Statistically significant trends are defined as those where the SNR is greater than 2.0. Bold numbers indicate statistically significant model reconstructions, and italicized numbers indicate significant but inverted model reconstructions.

Station	u_e	u_g	$u_e + u_g$
Ko Taphao Noi	-0.30 ± 0.13	2.34 ± 0.87	1.93 ± 0.81
Langkawi	0.10 ± 0.05	-0.10 ± 0.44	0.01 ± 0.40
Penang	-0.01 ± 0.01	0.28 ± 0.37	0.35 ± 0.36
Lumut	-0.06 ± 0.03	1.50 ± 0.34	1.42 ± 0.32
Kelang	-0.08 ± 0.04	-1.57 ± 0.29	-1.72 ± 0.28
Keling	0.01 ± 0.02	0.57 ± 0.34	0.54 ± 0.33
Kukup	-0.11 ± 0.07	-1.18 ± 1.07	-1.13 ± 1.01
Singapore	-0.03 ± 0.04	1.33 ± 0.16	1.28 ± 0.17
Sedili	0.12 ± 0.03	0.49 ± 0.06	0.56 ± 0.07
Tioman	0.26 ± 0.00	0.40 ± 0.05	0.62 ± 0.08
Kuantan	0.26 ± 0.06	0.20 ± 0.07	0.38 ± 0.07
Cendering	0.36 ± 0.08	0.02 ± 0.09	0.36 ± 0.06
Getting	0.28 ± 0.09	0.12 ± 0.10	0.39 ± 0.15
Ko Lak	0.65 ± 0.14	1.22 ± 0.33	1.75 ± 0.36
Vung Tau	0.75 ± 0.09	0.29 ± 0.27	0.63 ± 0.19
Bintulu	0.25 ± 0.07	0.28 ± 0.33	0.53 ± 0.34
Kota Kinabalu	-0.09 ± 0.03	0.09 ± 0.06	0.01 ± 0.08
Sandakan	0.01 ± 0.01	0.01 ± 0.01	0.01 ± 0.01
Jakarta	-0.44 ± 0.17	-0.48 ± 0.23	-1.18 ± 0.58
Surabaya	-0.46 ± 0.11	1.26 ± 0.57	1.09 ± 0.35

10 gauges were recreated, with TRFs of 0.14 to 0.49. Overall, the best explanations for the observed seasonal variability of diurnal and semidiurnal tides are found by the combination of Ekman and geostrophic currents (11 for M_2 , 8 for S_2 , 8 for K_1 , and 9 for O_1). Tidal variability is better explained in the GOT than in the other subregions, with all gauges in the GOT being significantly recreated for M_2 models and half of GOT gauges for K_1 . The Ekman velocity model responses were limited to specific locations and only had a noticeable effect during winter and summer monsoon, when winds were strong. In the narrow MS, the effect of Ekman forcing is virtually nonexistent for all constituents.

6. Discussion

The results in section 5 have indicated partial success in explaining the tidal seasonality in terms of frictional interactions of the tides and subtidal flows. But the model describing these interactions contains some significant limitations, both in the dynamics and in the forcing, which should be considered.

One key issue is uncertainty in the ocean bottom topography H , a field that occurs in both the dynamics and the mixing length model. During the course of model implementation estimates of H from ETOPO1

TABLE 5. As in Table 4, but for K_1 .

Station	u_e	u_g	$u_e + u_g$
Ko Taphao Noi	0.01 ± 0.01	0.09 ± 0.02	0.01 ± 0.02
Langkawi	0.01 ± 0.01	0.12 ± 0.03	0.15 ± 0.04
Penang	0.06 ± 0.02	0.37 ± 0.12	0.46 ± 0.14
Lumut	0.04 ± 0.01	0.45 ± 0.13	0.46 ± 0.10
Kelang	-0.02 ± 0.02	-0.16 ± 0.03	-0.22 ± 0.05
Keling	0.02 ± 0.03	0.18 ± 0.18	0.28 ± 0.26
Kukup	-0.17 ± 0.06	-0.01 ± 0.06	-0.22 ± 0.09
Singapore	0.12 ± 0.02	0.05 ± 0.03	0.11 ± 0.02
Sedili	0.33 ± 0.04	0.22 ± 0.07	0.28 ± 0.07
Tioman	0.18 ± 0.04	-0.17 ± 0.07	0.12 ± 0.16
Kuantan	0.02 ± 0.02	0.03 ± 0.05	0.25 ± 0.06
Cendering	0.02 ± 0.03	0.04 ± 0.04	0.03 ± 0.05
Getting	0.06 ± 0.02	0.11 ± 0.01	0.15 ± 0.02
Ko Lak	0.03 ± 0.04	0.05 ± 0.03	0.01 ± 0.06
Vung Tau	-0.11 ± 0.03	0.16 ± 0.04	0.01 ± 0.10
Bintulu	-0.11 ± 0.02	0.04 ± 0.04	-0.20 ± 0.05
Kota Kinabalu	0.03 ± 0.01	0.01 ± 0.01	0.03 ± 0.01
Sandakan	0.01 ± 0.01	0.01 ± 0.01	0.01 ± 0.01
Jakarta	0.01 ± 0.02	0.10 ± 0.03	0.11 ± 0.04
Surabaya	0.06 ± 0.01	0.09 ± 0.02	0.15 ± 0.02

(Amante and Eakins 2009) and the General Bathymetric Chart of the Oceans (GEBCO; Tani et al. 2011) were compared with DBDB2, and significant differences in topography were noted in the MS and JS. While these different topographies could lead to different tidal solutions, it was hoped that the anomaly admittances reported in section 5 would be less sensitive than the absolute admittances.

Another factor to consider is the mismatch in resolution between the dynamical model and the forcing climatologies. All sources of climatic data used have a much coarser resolution than the model bathymetry ($1/4^\circ$ or $1/8^\circ$ vs $1/30^\circ$). Each of these climatological fields were interpolated to the native model resolution, but these fields may not be representative near the coasts. This is particularly true in the MS, which is very shallow and narrow, in some places only $1/2^\circ$ wide.

Another reason for model mismatch with observed behavior might be due to unmodeled processes. Because the tide model is barotropic, the internal tide and baroclinic river effects are not represented. Vung Tau is near the mouth of the Mekong River, and it is hypothesized that seasonal discharge explains tidal seasonality there, which was largely unexplained by our models. The representation of bottom friction with the quasi-linear bottom stress, the use of a spatially constant drag coefficient C_d , and the mixing length model are also sources of error. Examination of these processes would require a finer-resolution, fully three-dimensional model and is beyond the scope of this study. Finally, storminess is another factor not considered that may have an effect on tidal variability. Storminess may drive fluctuations of water levels on a daily scale or modulate the bottom stress via surface

FIG. 15. TRFs of M_2 model responses compared to observed seasonality. Shades of red indicate positive TRFs of ranges of 0.1–0.25, 0.25–0.50, 0.50–1.00, and greater than 1.00, with darker reds indicating the larger magnitude TRF values. Shades of blue show inverted (negative) TRF responses, with the same ranges as the positive TRFs and darker blues indicating larger magnitude TRFs. (top) Individual model responses: (a) u_e and (b) u_g . (c) The combined model response of $u_e + u_g$. (bottom right) The legend shows the correspondence of colored dots to TRFs.

waves, contributing to tidal seasonality (Bromirski et al. 2011). As to the question of why the semidiurnal model responded better than diurnal models, a definitive answer is not yet known, but it may be indirectly due to the resonant dynamics that amplify the diurnal constituents and dampen the semidiurnal constituents in the SCS and surrounding waters (Yanagi and Takao 1998) or could be related to nonlinear interaction between constituents, such as a resonant triad interaction between M_2 , K_1 , and O_1 , which was observed in the Solomon Sea by Devlin et al. (2014), since all of these constituents are of near-equal magnitude in our study domain. Part of the answer could also lie in how different tidal frequencies respond to river runoff, which is significant in this region during monsoon season and is not resolved by our model.

7. Conclusions

Diurnal and semidiurnal tidal amplitudes and phases exhibit a high degree of seasonality in the seas of Southeast Asia. The magnitude of M_2 tidal variability

is about twice that of S_2 , and K_1 is about twice that of O_1 . The majority of the observed tidal variability is coherent with the western North Pacific monsoon index (WNPMI) in the annual frequency band, with strong correlations in the majority of the MS and all of the GOT, moderate correlations in the SCS, and weak correlations in the JS.

A model for the frictional interaction of subtidal currents with the tides was developed that utilized monthly climatologies of wind stress and water surface elevations. The model indicated that the wind-driven Ekman transport is important only during monsoon seasons in shallow regions and is negligible in narrow passages such as the MS. Based on the regressions of observed versus modeled tidal variability at the 20 sites, it was found that 14 were at least partially explained for M_2 , 9 for S_2 , 11 for K_1 , and 10 for O_1 . The best comparisons in general were produced by the combination of geostrophic and Ekman currents.

The model–data correlations were ubiquitously larger in shallow water than in deep ocean locations. The best

FIG. 16. As in Fig. 15, but for K_1 .

model reconstructions for all constituents were in the GOT, with the semidiurnal models performing better than diurnal models. Tidal variability in the SCS and JS is generally small, and the model agreement was better for semidiurnal than diurnal seasonality. The exception to this was Vung Tau, which exhibits large tidal seasonality unexplained the model, thought to be due to Mekong River discharge not captured in the model. Results in the MS are mixed and less reliable, with some constituents at some gauges being well explained and others not, particularly the K_1 amplitudes and phases, owing to possibly unreliable data in this shallow narrow region.

Overall, this study showed that even though tidal seasonality is spatially and temporally complex, a simplified model of frictional coupling can explain much of the variability. The hypothesis of monsoonal dynamics being the causative factor is supported, as areas where the tidal variability was strongly correlated to the WNPMI were regions where the models had the most success in reconstructing the observed variability. Further work to improve this study includes the utilization of a higher-resolution, fully three-dimensional baroclinic model that could include riverine and internal tide

mechanisms, in addition to future studies applying similar methods to other monsoon-affected regions, such as India.

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REFERENCES

- Amante, C., and B. W. Eakins, 2009: ETOPO1 1 arc-minute global relief model: Procedures, data sources and analysis. NOAA Tech. Memo. NESDIS NGDC-24, 25 pp., <https://www.ngdc.noaa.gov/mgg/global/relief/ETOPO1/docs/ETOPO1.pdf>.

- Arbic, B. K., A. J. Wallcraft, and E. J. Metzger, 2010: Concurrent simulation of the eddying general circulation and tides in a global ocean model. *Ocean Modell.*, **32**, 175–187, <https://doi.org/10.1016/j.ocemod.2010.01.007>.
- Bromirski, P. D., A. J. Miller, R. E. Flick, and G. Auad, 2011: Dynamical suppression of sea level rise along the Pacific coast of North America: Indications for imminent acceleration. *J. Geophys. Res.*, **116**, C07005, <https://doi.org/10.1029/2010JC006759>.
- Cartwright, D. E., and R. J. Tayler, 1971: New computations of the tide-generating potential. *Geophys. J. Roy. Astron. Soc.*, **23**, 45–73, <https://doi.org/10.1111/j.1365-246X.1971.tb01803.x>.
- , and A. C. Edden, 1973: Corrected tables of tidal harmonics. *Geophys. J. Roy. Astron. Soc.*, **33**, 253–264, <https://doi.org/10.1111/j.1365-246X.1973.tb03420.x>.
- , and M. Amin, 1986: The variances of tidal harmonics. *Dtsch. Hydrogr. Z.*, **39**, 235–253, <https://doi.org/10.1007/BF02226467>.
- Corkan, R. H., 1934: An annual perturbation in the range of tide. *Philos. Trans. Roy. Soc. London*, **144A**, 537–559, <https://doi.org/10.1098/rspa.1934.0067>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Devlin, A. T., D. A. Jay, S. A. Talke, and E. D. Zaron, 2014: Can tidal perturbations associated with sea level variations in the western Pacific Ocean be used to understand future effects of tidal evolution? *Ocean Dyn.*, **64**, 1093–1120, <https://doi.org/10.1007/s10236-014-0741-6>.
- , —, —, J. Pan, and H. Lin, 2017a: Coupling of sea level and tidal range changes, with implications for future water levels. *Sci. Rep.*, **7**, 17021, <https://doi.org/10.1038/s41598-017-17056-z>.
- , —, E. Zaron, S. A. Talke, J. Pan, and H. Lin, 2017b: Tidal variability related to sea level variability in the Pacific Ocean. *J. Geophys. Res. Oceans*, **122**, 8445–8463, <https://doi.org/10.1002/2017JC013165>.
- Dibarboure, G., C. M. Pujol, F. Briol, P.-Y. LeTraon, G. Larnicol, N. Picot, F. Mertz, and M. Ablain, 2011: Jason-2 in DUACS: First tandem results and impact on processing and products. *Mar. Geod.*, **34**, 214–241, <https://doi.org/10.1080/01490419.2011.584826>.
- DUACS/AVISO team, 2014: A new version of SSALTO/DUACS products available in April 2014. AVISO/CNES Rep., 38 pp., <https://www.aviso.altimetry.fr/fileadmin/documents/data/duacs/Duacs2014.pdf>.
- Egbert, G. D., and S. Y. Erofeeva, 2002: Efficient inverse modeling of barotropic ocean tides. *J. Atmos. Oceanic Technol.*, **19**, 183–204, [https://doi.org/10.1175/1520-0426\(2002\)019<0183:EIMOB>2.0.CO;2](https://doi.org/10.1175/1520-0426(2002)019<0183:EIMOB>2.0.CO;2).
- , A. F. Bennett, and M. Foreman, 1994: TOPEX/Poseidon tides estimated using a global inverse model. *J. Geophys. Res.*, **99**, 24 821–24 852, <https://doi.org/10.1029/94JC01894>.
- Fang, G., Y.-K. Kwok, K. Yu, and Y. Zhu, 1999: Numerical simulation of principal tidal constituents in the South China Sea, Gulf of Tonkin and Gulf of Thailand. *Cont. Shelf Res.*, **19**, 845–869, [https://doi.org/10.1016/S0278-4343\(99\)00002-3](https://doi.org/10.1016/S0278-4343(99)00002-3).
- Foreman, M. G., R. A. Walters, R. F. Henry, C. P. Keller, and A. G. Dolling, 1995: A tidal model for eastern Juan de Fuca Strait and the southern Strait of Georgia. *J. Geophys. Res.*, **100**, 721–740, <https://doi.org/10.1029/94JC02721>.
- Forryan, A., A. P. Martin, M. A. Srokosz, E. E. Popova, S. C. Painter, and A. H. H. Renner, 2013: A new observationally motivated Richardson number based mixing parametrization for oceanic mesoscale flow. *J. Geophys. Res. Oceans*, **118**, 1405–1419, <https://doi.org/10.1002/jgrc.20108>.
- Godin, G., 1985: Modification of river tides by the discharge. *J. Waterw. Port Coastal Ocean Eng.*, **111**, 257–274, [https://doi.org/10.1061/\(ASCE\)0733-950X\(1985\)111:2\(257\)](https://doi.org/10.1061/(ASCE)0733-950X(1985)111:2(257)).
- , 1991: Frictional effects in river tides. *Tidal Hydrodynamics*, B. B. Parker, Ed., John Wiley and Sons, 379–402.
- Guo, L., M. van der Wegen, D. A. Jay, P. Matte, Z. B. Wang, D. Roelvink, and Q. He, 2015: River-tide dynamics: Exploration of nonstationary and nonlinear tidal behavior in the Yangtze River estuary. *J. Geophys. Res. Oceans*, **120**, 3499–3521, <https://doi.org/10.1002/2014JC010491>.
- Hartmann, T., and H.-G. Wenzel, 1995: The HW95 tidal potential catalogue. *Geophys. Res. Lett.*, **22**, 3553–3556, <https://doi.org/10.1029/95GL03324>.
- Hinze, J. O., 1975: *Turbulence*. 2nd ed. McGraw-Hill Publishing, 790 pp.
- Howarth, M. J., 1998: The effect of stratification on tidal current profiles. *Cont. Shelf Res.*, **18**, 1235–1254, [https://doi.org/10.1016/S0278-4343\(98\)00042-9](https://doi.org/10.1016/S0278-4343(98)00042-9).
- Huess, V., and O. B. Andersen, 2001: Seasonal variation in the main tidal constituent from altimetry. *Geophys. Res. Lett.*, **28**, 567–570, <https://doi.org/10.1029/2000GL011921>.
- Hunter, J. R., 1975: A note on quadratic friction in the presence of tides. *Estuarine Coastal Mar. Sci.*, **3**, 473–475, [https://doi.org/10.1016/0302-3524\(75\)90047-X](https://doi.org/10.1016/0302-3524(75)90047-X).
- Jan, S., C.-S. Chern, J. Wang, and S.-Y. Chao, 2007: Generation of diurnal K₁ internal tide in the Luzon Strait and its influence on surface tide in the South China Sea. *J. Geophys. Res.*, **112**, C06019, <https://doi.org/10.1029/2006JC004003>.
- , R.-C. Lien, and C.-H. Ting, 2008: Numerical study of baroclinic tides in Luzon Strait. *J. Oceanogr.*, **64**, 789–802, <https://doi.org/10.1007/s10872-008-0066-5>.
- Kang, S. K., J. Chung, S. Lee, and K. Yum, 1995: Seasonal variability of the M₂ tide in the seas adjacent to Korea. *Cont. Shelf Res.*, **15**, 1087–1113, [https://doi.org/10.1016/0278-4343\(94\)00066-V](https://doi.org/10.1016/0278-4343(94)00066-V).
- , S.-R. Lee, and H.-J. Lie, 1998: Fine grid tidal modeling of the Yellow and East China Seas. *Cont. Shelf Res.*, **18**, 739–772, [https://doi.org/10.1016/S0278-4343\(98\)00014-4](https://doi.org/10.1016/S0278-4343(98)00014-4).
- , M. G. G. Foreman, H.-J. Lie, J.-H. Lee, J. Cherniawsky, and K.-D. Yum, 2002: Two-layer tidal modeling of the Yellow and East China Seas with application to seasonal variability of the M₂ tide. *J. Geophys. Res.*, **107**, 3020, <https://doi.org/10.1029/2001JC000838>.
- Lagerloef, G. S., G. T. Mitchum, R. B. Lukas, and P. P. Niiler, 1999: Tropical Pacific near-surface currents estimated from altimeter, wind, and drifter data. *J. Geophys. Res.*, **104**, 23 313–23 326, <https://doi.org/10.1029/1999JC900197>.
- Leeuwenburgh, O., O. B. Andersen, and V. Huess, 1999: Seasonal tide variations from tide gauges and altimetry. *Phys. Chem. Earth*, **24A**, 403–406, [https://doi.org/10.1016/S1464-1895\(99\)00049-6](https://doi.org/10.1016/S1464-1895(99)00049-6).
- Leffler, K. E., and D. A. Jay, 2009: Enhancing tidal harmonic analysis: Robust (hybrid L1/L2) solutions. *Cont. Shelf Res.*, **29**, 78–88, <https://doi.org/10.1016/j.csr.2008.04.011>.
- Li, T., and B. Wang, 2005: A review on the western North Pacific monsoon: Synoptic-to-interannual variabilities. *Terr. Atmos. Oceanic Sci.*, **16**, 285–314, [https://doi.org/10.3319/TAO.2005.16.2.285\(A\)](https://doi.org/10.3319/TAO.2005.16.2.285(A)).
- Louis, J. F., 1979: A parametric model of vertical eddy fluxes in the atmosphere. *Bound.-Layer Meteor.*, **17**, 187–202, <https://doi.org/10.1007/BF00117978>.

- Müller, M., 2012: The influence of changing stratification conditions on barotropic tidal transport. *Cont. Shelf Res.*, **47**, 107–118, <https://doi.org/10.1016/j.csr.2012.07.003>.
- , J. Y. Cherniawsky, M. G. G. Foreman, and J.-S. von Storch, 2012: Global M_2 internal tide and its seasonal variability from high resolution ocean circulation and tide modeling. *Geophys. Res. Lett.*, **39**, L19607, <https://doi.org/10.1029/2012GL053320>.
- Pacanowski, R., and S. Philander, 1981: Parameterization of vertical mixing in numerical models of tropical oceans. *J. Phys. Oceanogr.*, **11**, 1443–1451, [https://doi.org/10.1175/1520-0485\(1981\)011<1443:POVMIN>2.0.CO;2](https://doi.org/10.1175/1520-0485(1981)011<1443:POVMIN>2.0.CO;2).
- Pawlowicz, R., B. Beardsley, and S. Lentz, 2002: Classical tidal harmonic analysis including error estimates in MATLAB using T_TIDE. *Comput. Geosci.*, **28**, 929–937, [https://doi.org/10.1016/S0098-3004\(02\)00013-4](https://doi.org/10.1016/S0098-3004(02)00013-4).
- Perlin, A., J. N. Moum, J. M. Klymak, M. D. Levine, T. Boyd, and P. M. Kosro, 2005: A modified law-of-the-wall applied to oceanic bottom boundary layers. *J. Geophys. Res.*, **110**, C10S10, <https://doi.org/10.1029/2004JC002310><https://doi.org/10.1029/2004JF000124>.
- Pugh, D. T., and J. M. Vassie, 1976: Tide and surge propagation offshore in the dowsing region of the North Sea. *Dtsch. Hydrogr. Z.*, **29**, 163–213, <https://doi.org/10.1007/BF02226659>.
- Snyder, R. L., M. Sidjabat, and J. H. Filloux, 1979: A study of tides, setup and bottom friction in a shallow semi-enclosed basin. Part II: Tidal model and comparison with data. *J. Phys. Oceanogr.*, **9**, 170–188, [https://doi.org/10.1175/1520-0485\(1979\)009<0170:ASOTSA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1979)009<0170:ASOTSA>2.0.CO;2).
- St-Laurent, P., F. J. Saucier, and J.-F. Dumais, 2008: On the modification of tides in a seasonally ice-covered sea. *J. Geophys. Res.*, **113**, C11014, <https://doi.org/10.1029/2007JC004614>.
- Stammer, D., and Coauthors, 2014: Accuracy assessment of global barotropic ocean tide models. *Rev. Geophys.*, **52**, 243–282, <https://doi.org/10.1002/2014RG000450>.
- van Haren, H., 2000: Properties of vertical current shear across stratification in the North Sea. *J. Mar. Res.*, **58**, 465–491, <https://doi.org/10.1357/002224000321511115>.
- Tani, S., and Coauthors, 2011: General Bathymetric Chart of the Oceans (GEBCO 2008). International Hydrographic Office/Intergovernmental Oceanographic Commission, accessed 3 August 2011, <http://www.gebco.net>.
- Wang, B., and Z. Fan, 1999: Choice of South Asian summer monsoon indices. *Bull. Amer. Meteor. Soc.*, **80**, 629–638, [https://doi.org/10.1175/1520-0477\(1999\)080<0629:COSASM>2.0.CO;2](https://doi.org/10.1175/1520-0477(1999)080<0629:COSASM>2.0.CO;2).
- Wyrtki, K., 1961: Physical oceanography of the Southeast Asian waters. Scripps Institution of Oceanography Rep., 195 pp.
- Yanagi, T., and T. Takao, 1998: Clockwise phase propagation of semi-diurnal tides in the Gulf of Thailand. *J. Oceanogr.*, **54**, 143–150, <https://doi.org/10.1007/BF02751690>.
- , S. I. Sachoemar, T. Takao, and S. Fujiwara, 2001: Seasonal variation of stratification in the Gulf of Thailand. *J. Oceanogr.*, **57**, 461–470, <https://doi.org/10.1023/A:1021237721368>.
- Yu, Z., and P. S. Schopf, 1997: Vertical eddy mixing in the tropical upper ocean: Its influence on zonal currents. *J. Phys. Oceanogr.*, **27**, 1447–1458, [https://doi.org/10.1175/1520-0485\(1997\)027<1447:VEMITT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1997)027<1447:VEMITT>2.0.CO;2).
- Zu, T., J. Gan, and S. Y. Erofeeva, 2008: Numerical study of the tide and tidal dynamics in the South China Sea. *Deep-Sea Res. I*, **55**, 137–154, <https://doi.org/10.1016/j.dsr.2007.10.007>.