

# Numerical Study of Baroclinic Tides in Luzon Strait

SEN JAN<sup>1\*</sup>, REN-CHIEH LIEN<sup>2</sup> and CHI-HUA TING<sup>1</sup>

<sup>1</sup>*Institute of Hydrological and Oceanic Sciences, National Central University,  
300 Jung-da Road, Jung-li 32001, Taiwan, R.O.C.*

<sup>2</sup>*Applied Physics Laboratory, University of Washington,  
1013 NE 40th Street, Box 355640, Seattle, WA 98105-6698, U.S.A.*

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The spatial and temporal variations of baroclinic tides in the Luzon Strait (LS) are investigated using a three-dimensional tide model driven by four principal constituents,  $O_1$ ,  $K_1$ ,  $M_2$  and  $S_2$ , individually or together with seasonal mean summer or winter stratifications as the initial field. Barotropic tides propagate predominantly westward from the Pacific Ocean, impinge on two prominent north-south running submarine ridges in LS, and generate strong baroclinic tides propagating into both the South China Sea (SCS) and the Pacific Ocean. Strong baroclinic tides, ~19 GW for diurnal tides and ~11 GW for semidiurnal tides, are excited on both the east ridge (70%) and the west ridge (30%). The barotropic to baroclinic energy conversion rate reaches 30% for diurnal tides and ~20% for semidiurnal tides. Diurnal ( $O_1$  and  $K_1$ ) and semidiurnal ( $M_2$ ) baroclinic tides have a comparable depth-integrated energy flux 10–20 kW m<sup>-1</sup> emanating from the LS into the SCS and the Pacific basin. The spring-neap averaged, meridionally integrated baroclinic tidal energy flux is ~7 GW into the SCS and ~6 GW into the Pacific Ocean, representing one of the strongest baroclinic tidal energy flux regimes in the World Ocean. About 18 GW of baroclinic tidal energy, ~50% of that generated in the LS, is lost locally, which is more than five times that estimated in the vicinity of the Hawaiian ridge. The strong westward-propagating semidiurnal baroclinic tidal energy flux is likely the energy source for the large-amplitude nonlinear internal waves found in the SCS. The baroclinic tidal energy generation, energy fluxes, and energy dissipation rates in the spring tide are about five times those in the neap tide; while there is no significant seasonal variation of energetics, but the propagation speed of baroclinic tide is about 10% faster in summer than in winter. Within the LS, the average turbulence kinetic energy dissipation rate is  $O(10^{-7})$  W kg<sup>-1</sup> and the turbulence diffusivity is  $O(10^{-3})$  m<sup>2</sup>s<sup>-1</sup>, a factor of 100 greater than those in the typical open ocean. This strong turbulence mixing induced by the baroclinic tidal energy dissipation exists in the main path of the Kuroshio and is important in mixing the Pacific Ocean, Kuroshio, and the SCS waters.

Keywords:

- Numerical model,
- baroclinic tides,
- seasonal,
- fortnightly,
- Luzon Strait.

## 1. Introduction

Luzon Strait (LS) is the primary deep passage connecting the world's largest marginal sea, the South China Sea (SCS), to the northwest Pacific Ocean (Fig. 1(a)). Two prominent submarine ridges running north-south exist in the LS with complex and abruptly changing topography (Fig. 1(b)). The east ridge, the Luzon Island Arc, consists of a series of isles extending from the south-east coast of Taiwan to the north of Luzon. The west ridge,

the Heng-Chun Ridge, extends from the southern tip of Taiwan to the middle reaches of the LS. The east ridge is generally higher and longer than the west ridge.

The Kuroshio and tides are the dominant currents in the LS. The northward-flowing Kuroshio (Fig. 1(a)) sometimes penetrates into the LS at a speed of ~1 m s<sup>-1</sup> with a strong seasonal variation (e.g., Metzger and Hurlbert, 1996; Hu *et al.*, 2000; Centurioni *et al.*, 2004; Tian *et al.*, 2006). Massive westward intrusion of the Kuroshio occurs mostly during October to January (Centurioni *et al.*, 2004). Occasionally, anticyclonic rings detached from the Kuroshio flowing across the LS are observed in the northern SCS (Li *et al.*, 1998). The transport of the Kuroshio

\* Corresponding author. E-mail: senjan@ncu.edu.tw

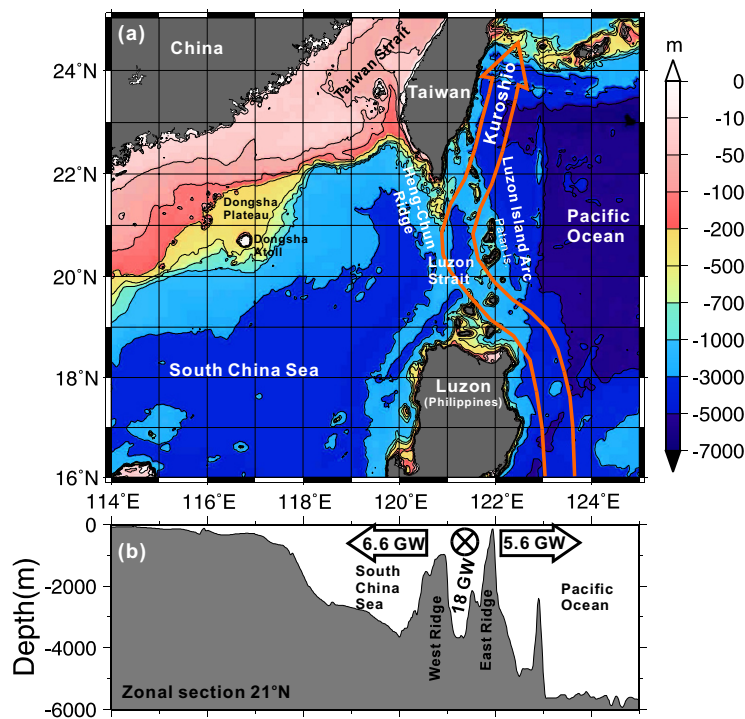


Fig. 1. (a) Bathymetry in the northern South China Sea, northwest Pacific Ocean and Luzon Strait. (b) Topographic transect along 21°N and main results of baroclinic tidal energy fluxes and dissipation. Arrows of 6.6 and 5.6 GW represent the spring-neap mean baroclinic energy fluxes that propagate into the South China Sea and the Pacific Ocean, respectively, from the Luzon Strait. About 18 GW of the baroclinic tide dissipation occurs in the Luzon Strait. ⊗ marks the Kuroshio.

into the SCS is complicated due to the fluctuation of the Kuroshio path. The Kuroshio front in the LS separates the Pacific Ocean and the SCS waters. The mixing of these two water masses depends on the available turbulent processes in the LS. Baroclinic tidal energy generated and dissipated in the LS could be one of the primary turbulence energy sources, as concluded in the present analysis.

Numerical simulations suggest that both diurnal and semidiurnal barotropic tidal currents oscillate at amplitudes of  $0.05\text{--}0.3\text{ m s}^{-1}$  with major axes nearly perpendicular to the two ridges in the LS (Fang *et al.*, 1999; Jan *et al.*, 2002). The tidal sea level amplitudes of the  $O_1$ ,  $K_1$  and  $M_2$  are 0.25, 0.25 and 0.38 m, respectively, measured by a coastal tide gauge at southernmost Taiwan (Jan *et al.*, 2004), suggesting a nearly equal contribution of the three tidal constituents in the LS. The equal importance of the three constituents is closely related to the quasi-resonant response of the diurnal tide in the SCS (Yanagi and Takao, 1998).

Strong baroclinic tides and nonlinear internal waves with vertical isotherm displacement up to 150 m have been observed in the LS and the northern SCS near Dongsha Island (Duda *et al.*, 2004; Ramp *et al.*, 2004; Yang *et al.*, 2004). The barotropic tidal current and topography inter-

action in the LS is considered to be the major generating mechanism for the baroclinic tides (Chao *et al.*, 2007). When baroclinic tides propagate onto a shallow continental slope/shelf, the nonlinear destabilizing effect may steepen and transform baroclinic tides to nonlinear internal waves, which may be further disintegrated into a sequence of solitary waves (Gerkema and Zimmerman, 1995; Lien *et al.*, 2005). Thus the generation of baroclinic tides and the evolution from baroclinic tides to nonlinear internal waves are the focus of contemporary research using in-situ and remote-sensing observations and numerical models (Lien *et al.*, 2005; Zhao and Alford, 2006; Chang *et al.*, 2006). Niwa and Hibiya (2004) used a three-dimensional model to study  $M_2$  baroclinic tides in the western Pacific Ocean. They concluded that 14 GW of the 54 GW westward-propagating  $M_2$  barotropic tidal energy is converted to baroclinic tide in the LS. Nearly 7 GW (50%) of  $M_2$  baroclinic tidal energy dissipates locally and  $\sim 7.4$  GW (50%) propagates into SCS (4.2 GW) and into Pacific Ocean (3.2 GW). Jan *et al.* (2007) studied the effect of the  $K_1$  baroclinic tide generated in the LS on the partial-standing, quasi-resonant barotropic tide in the SCS. The energy of the barotropic  $K_1$  tide in the SCS is significantly reduced due to the barotropic to baroclinic energy conversion in the LS which amounts

Table 1. Parameters for calculating the adjusted height of equilibrium tides.

Constituent	Amplitude (m) of equilibrium tide ( $\xi$ )	Elasticity factor ( $\alpha = 1 + k - h$ )	$a (= \alpha \xi)$ for calculating $H_c$ in (2)
O <sub>1</sub>	0.101	0.695	0.070
K <sub>1</sub>	0.142	0.736	0.105
M <sub>2</sub>	0.244	0.693	0.169
S <sub>2</sub>	0.114	0.693	0.079

up to 30% of the incident barotropic tidal energy.

The properties of baroclinic tides and nonlinear internal waves in the SCS are affected by baroclinic tides generated in the LS. A better understanding of the variations of barotropic and baroclinic tidal energetics in the LS is vital to quantify the energetics of baroclinic tides and nonlinear internal waves in the SCS and to identify the turbulence energy available for mixing water masses of the Pacific Ocean and the SCS in the LS (illustrated in Fig. 1(b)). The properties of barotropic and baroclinic tides are strongly inhomogeneous in the LS due to the complicated three-dimensional features of the two submarine ridges (Fig. 1). In-situ study to understand the dynamics of barotropic and baroclinic tides within the LS is expensive and difficult, if not impossible. The present numerical model study is motivated by a wish to study the details of baroclinic tide properties within the LS and provide guidance for future observational study. Seasonal and fortnightly variations and detailed properties of the active roles of the two ridges on generating, dissipating, and mediating energy fluxes have not been discussed in previous studies (e.g., Niwa and Hibiya, 2004; Jan *et al.*, 2007; Chao *et al.*, 2007) and form the primary focus of the present model study. The importance of the west ridge in generating and dissipating baroclinic tidal energy is also evaluated.

We use a three-dimension regional tide model, the same as that used by Jan *et al.* (2007), with realistic stratifications and tidal forcing. The seasonal mean temperature and salinity profiles derived from historical conductivity-temperature-depth (CTD) data collected in the LS are used as the initial condition. Tidal sea levels of four principal constituents (O<sub>1</sub>, K<sub>1</sub>, M<sub>2</sub> and S<sub>2</sub>) on the open boundaries are used to drive the model, individually and in combination. Note that the Kuroshio is not represented in this model study for the model constraint and for clarity in understanding the tidal dynamics in the LS. The Kuroshio exhibits strong seasonal variations within the LS. The density front associated with the Kuroshio and its migration could affect the barotropic to baroclinic energy conversion rate and energy propagation. The relative vorticity of the Kuroshio could modify or trap the baroclinic tidal energy flux (Rainville and Pinkel, 2004).

The effects of the Kuroshio on baroclinic tides are beyond the scope of the present study and should be considered in future investigations.

## 2. Model Description

The three-dimensional baroclinic tide model in this study is a modification of the Princeton Ocean Model described in Blumberg and Mellor (1987). The nonlinear primitive equation model with Boussinesq and hydrostatic approximations is driven by the barotropic tidal forcing. The governing equations have been delineated in Jan *et al.* (2007). The forcing term, i.e. the tidal potential in the horizontal momentum balance, is formulated as:

$$F = gD\nabla_H(\zeta - \beta\eta), \quad (1)$$

where  $g$  is the gravitational acceleration,  $\eta$  is sea level displacement,  $D$  is total water depth ( $D = H + \eta$ ,  $H$ : mean water depth),  $\zeta$  is adjust height of equilibrium tides,  $\beta$  ( $=0.940$  for diurnal tides and  $0.953$  for semidiurnal tides) represents the loading effect due to ocean tides (Foreman *et al.*, 1993), and  $\nabla_H (= (\partial/\partial x)\vec{i} + (\partial/\partial y)\vec{j})$  is the horizontal divergence. Following Pugh (1987), the adjust height of equilibrium tides is defined as

$$\zeta = f_c H_c \cos[\omega_c t + (V_0 + \mu) + m\lambda], \quad (2)$$

where  $f_c$  is nodal factor,  $\omega_c$  is frequency of corresponding tidal constituent,  $V_0$  is initial phase angle of the equilibrium tides,  $\mu$  is nodal angle,  $m = 1$  or  $2$  accounts for diurnal or semidiurnal constituents, respectively,  $\lambda$  is longitude, subscript  $c$  represents tidal constituent, and  $H_c$  ( $=a_1 \sin 2\phi$  or  $a_2 \cos^2 \phi$  for diurnal or semidiurnal tides,  $\phi$ : latitude) is amplitude of equilibrium tides multiplied by the factor  $1 + k - h$  ( $k$  and  $h$  are Love numbers due to the elastic response and the redistribution of the mass of earth). Table 1 lists  $a_1$  and  $a_2$  for diurnal and semidiurnal constituents, respectively.

Since the vertical acceleration is excluded in the vertical momentum equation, processes of nonhydrostatic, nonlinear internal waves, including their conversion from baroclinic tides and interactions with baroclinic tides, are

not resolved in the present model. To parameterize these unresolved processes, we add an artificial linear dissipation term to the horizontal momentum equations similar to that adopted by Niwa and Hibiya (2004). The dissipation term is defined as

$$F_{damp} = -r(\bar{U} - \langle \bar{U} \rangle), \quad (3)$$

where  $\bar{U}$  is the horizontal velocity vector in the Cartesian coordinate,  $\langle \rangle$  represents depth average,  $r$  is a damping coefficient which is set to  $0.2 \text{ day}^{-1}$  as suggested by Niwa and Hibiya (2004).

The model is bounded within  $99.25^{\circ}$ – $135.25^{\circ}$ E and  $2.25^{\circ}$ – $43.25^{\circ}$ N with  $(1/12)^{\circ}$  horizontal resolution. There are 51 uneven  $\sigma$  layers in the vertical with  $\sigma_k = -(0, 0.002, 0.004, 0.006, 0.008, 0.01, 0.012, 0.014, 0.018, 0.022, 0.026, 0.03, 0.034, 0.037, 0.045, 0.053, 0.061, 0.069, 0.077, 0.085, 0.1, 0.116, 0.132, 0.148, 0.179, 0.211, 0.243, 0.274, 0.306, 0.337, 0.369, 0.4, 0.432, 0.464, 0.495, 0.527, 0.558, 0.59, 0.621, 0.653, 0.684, 0.716, 0.748, 0.779, 0.811, 0.842, 0.874, 0.905, 0.937, 0.968, 1)$ , from  $k = 1$  (surface) to 51 (bottom). The bottom topography was established using the revised ETOPO2 (<http://www.ngdc.noaa.gov/mgg/global/relief/ETOPO2/ETOPO2v2-2006/ETOPO2v2c/>) supplement with a 1-min depth archive in the region of  $105^{\circ}$ – $135^{\circ}$ E and  $2.25^{\circ}$ – $35^{\circ}$ N provided by the National Center for Ocean Research (NCOR) of Taiwan.

Figure 2 shows the initial seasonal mean temperature (T), salinity (S) and buoyancy frequency (N) profiles above 500 m depth for summer (July to September) and winter (December to February). The profiles are derived by averaging historical CTD data, provided by NCOR, collected in the vicinity of the LS. The major difference between the two seasonal mean profiles is that the pycnocline is broader and shallower in summer than in winter,  $N_{\max} = 0.018 \text{ s}^{-1}$ . The mixed layer is approximately 50 m and 80 m thick in summer and winter, respectively. Below 500 m, vertical profiles of T, S and N show no significant seasonal variation. Initial fields of T and S are set to be horizontally homogeneous to exclude the currents that might be generated due to the thermal wind relation.

The motionless ocean is subsequently driven by the tidal potential and prescribed tidal sea levels on all open-ocean boundaries through a forced radiation condition similar to that used by Blumberg and Kantha (1985). The tidal sea levels on the open boundaries are computed using harmonic constants compiled in a database (hereafter NAO.99) described in Matsumoto *et al.* (2000). The barotropic and baroclinic open boundary conditions, horizontal/vertical viscosity and diffusivity, and bottom stress formulations are described in Jan *et al.* (2007). Model

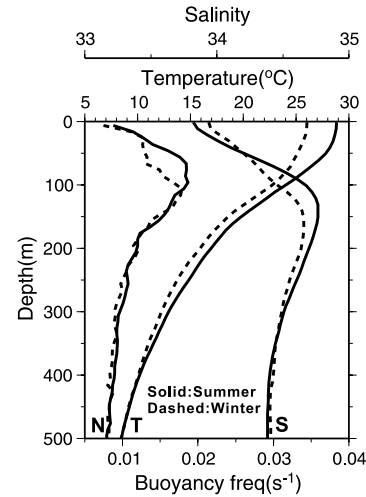


Fig. 2. Seasonal mean summer (solid lines) and winter (dashed lines) temperature, salinity and associated buoyancy frequency profiles for the initial fields of the model.

runs forced by single tidal constituent are set to 15 days; the hourly model results during the last three days, when the model reaches cyclic equilibrium, are analyzed. Model runs forced by the combined four tidal constituents, arbitrarily starting from January 1 00:00 UTC, 2006, are set to 23 days to cover the spring-neap tidal cycle. Three days of hourly model results during the spring tide (day 13–15) and neap tide (day 21–23) are analyzed to quantify the spring-neap variation.

The harmonic constants calculated from the simulated surface tides are compared with NAO.99. Figure 3 shows the model-simulated co-tidal charts for the four constituents under the summer stratification as an example. The distributions of the co-phase line for  $O_1$  and  $K_1$  off northern Luzon Island are likely indicative of amphidroms (Figs. 3(a) and (b)). For  $M_2$  and  $S_2$ , there are degenerated amphidroms north of Taiwan (Figs. 3(c) and (d)), which are consistent with those shown in Lefevre *et al.* (2000). The averaged sea level root-mean-square discrepancies for summer (winter), which considers both amplitude and phase differences, are 2.2 (2.1), 2.9 (3.1), 2.5 (2.4) and 1.0 (1.0) cm respectively for  $O_1$ ,  $K_1$ ,  $M_2$  and  $S_2$  as compared with the sea level calculated from NAO.99 at depths greater than 200 m in the vicinity of the LS ( $115^{\circ}$ – $127^{\circ}$ E,  $18^{\circ}$ – $23^{\circ}$ N). The associated goodness of fit for summer (winter) relative to the sea level calculated from NAO.99, similar to the percentage of accuracy (POA) defined in Lefevre *et al.* (2000), is 95.4 (95.4), 93.7 (93.0), 98.7 (98.8) and 98.8 (98.8)% for  $O_1$ ,  $K_1$ ,  $M_2$  and  $S_2$ , respectively, suggesting that barotropic tides are reasonably reproduced in the model. The difference between the harmonic constants derived from the simulated



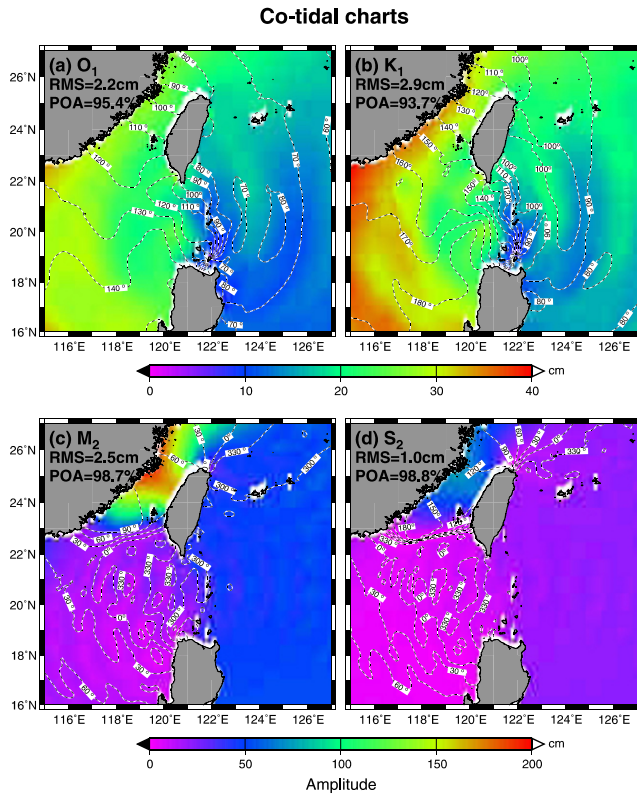


Fig. 3. Co-tidal charts for the model-simulated (a)  $O_1$ , (b)  $K_1$ , (c)  $M_2$  and (d)  $S_2$  constituents in the vicinity of the Luzon Strait. Initial field is summer stratification. Root-mean-square discrepancy (RMS) for the simulated tidal sea level relative to that calculated from NAO.99 in depths  $> 200$  m and the associated percentage of accuracy (POA) are inserted on each panel.

surface tides in summer and winter is not significant. Further fine-tuning might improve the accuracy, but that is not the major subject of this study.

The associated depth-averaged tidal current ellipses of the four constituents are shown in Fig. 4. The characteristics of barotropic tidal currents in the LS have been delineated in many papers, e.g., Fang *et al.* (1999), Lefevre *et al.* (2000) and Niwa and Hibiya (2004), and are not repeated here. Briefly, the current amplitudes are of similar magnitude  $\sim 0.2 \text{ m s}^{-1}$  for the  $O_1$ ,  $K_1$  and  $M_2$  and relatively small,  $0.1 \text{ m s}^{-1}$ , for the  $S_2$  in the LS. The barotropic tidal currents are much weaker,  $< 0.1 \text{ m s}^{-1}$ , in the deep Pacific Ocean and the deep northern SCS. The intensification of the barotropic tidal currents in the LS is mainly due to the narrowing and shoaling topographic effects when tidal waves propagate westward from the deep western Pacific into the SCS through the two submarine ridges in the LS. The characteristic of diurnal tides, which are quasi-resonant, partial-standing waves in the SCS with a meridional nodal band roughly across the LS

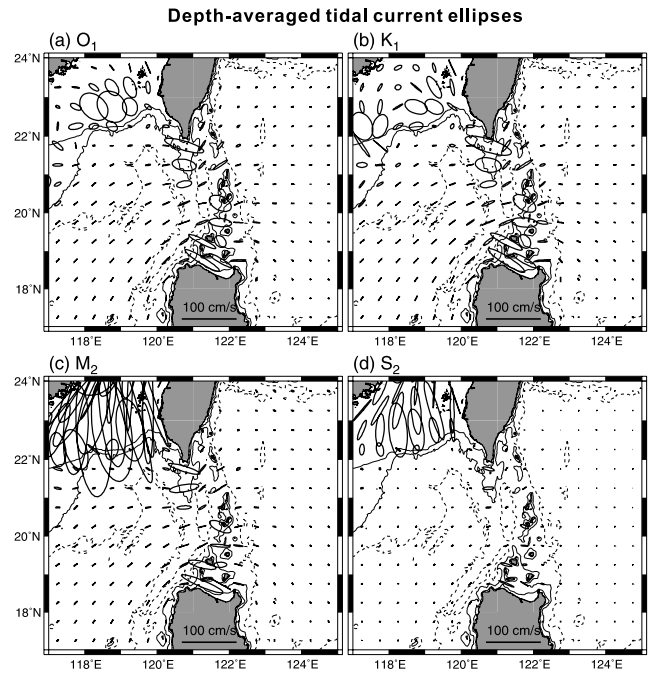


Fig. 4. Depth-averaged tidal current ellipses derived from model-simulated (a)  $O_1$ , (b)  $K_1$ , (c)  $M_2$  and (d)  $S_2$  constituents under summer stratification. 200 m (bold line), 1000 m (thin line) and 3000 m (dashed line) isobaths are appended.

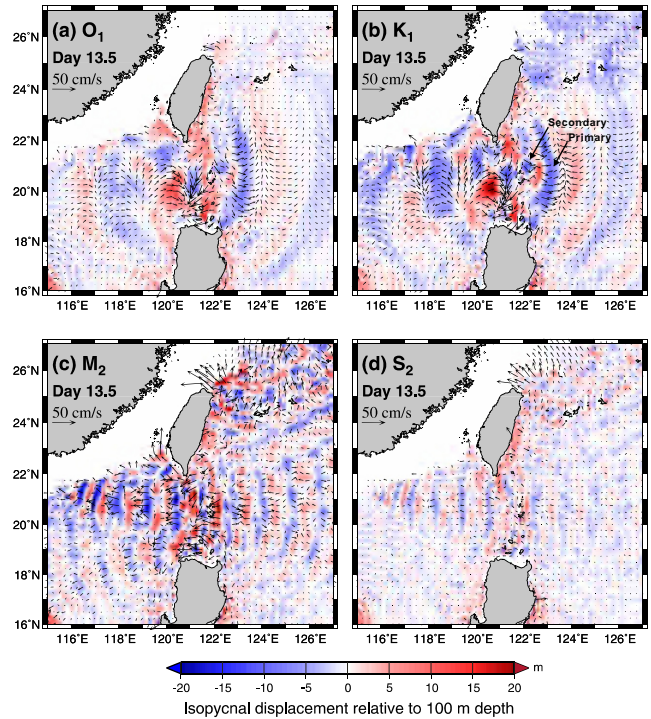


Fig. 5. Instantaneous (day 13.5) isopycnal vertical displacement relative to 100 m depth and corresponding tidal current vectors at 100 m depth derived from model results of (a)  $O_1$ , (b)  $K_1$ , (c)  $M_2$  and (d)  $S_2$  constituents under summer stratification.

(Yanagi and Takao, 1998; Jan *et al.*, 2007), also contributes to the barotropic tidal current strengthening. The major axes of tidal current ellipses of all four constituents are approximately perpendicular to the orientation of the two ridges in the LS, in favor of the generation of baroclinic tides.

It is interesting that the simulated amplitude and phase of the four tidal constituents in Fig. 3 appear to fluctuate with wavelengths similar to the horizontal wavelengths of underlying baroclinic tides around LS. Indeed, both the satellite Topex/Poseidon sea surface height and numerical results show that the strong baroclinic tides generated in the LS could modify the spatial distribution of amplitude and phase of the barotropic tide locally and even remotely in the SCS. Niwa and Hibiya (2004) attributed the short-wavelength ( $\sim 100$  km) fluctuations superposed on the simulated co-amplitude and co-phase lines (their figure 3) and along the Topex/Poseidon tracks (their figure 10) to the underlying  $M_2$  baroclinic tides. Similar fluctuations with wavelength  $\sim 200$  km are also found in the simulated  $K_1$  co-tidal chart around the LS (figure 4 in Jan *et al.*, 2007).

### 3. Results and Discussions

#### 3.1 General properties

Figure 5 illustrates instantaneous isopycnal vertical displacements relative to the equilibrium depth at 100 m and corresponding current velocities at day 13.5 for the four tidal constituents under the summer stratification. The amplitudes of the modeled baroclinic tides in the LS are  $\sim 20$  m for  $K_1$  and  $M_2$ ,  $\sim 10$  m for  $O_1$ , and  $\sim 5$  m for  $S_2$ . The complicated spatial distributions of baroclinic tidal wave patterns in the LS suggest a strong spatial inhomogeneity in the generation and propagation of baroclinic tides. The baroclinic tidal currents, as fast as  $0.25 \text{ m s}^{-1}$ , are dominant in the western Pacific Ocean and northern SCS, where the barotropic tidal currents are relatively small (cf. Fig. 4). Because of the meridional variation of the topography of the two ridges, both the zonal and meridional variations of baroclinic tidal energetics occur in the LS, which is especially clear for the diurnal tides (Figs. 5(a) and (b)). Notably, the  $M_2$  baroclinic tides that propagate into the northern SCS have two major generating sites in the LS. One is in the northern LS, where the baroclinic tides are excited and propagate westward toward the Dongsha Island roughly along  $21^\circ\text{N}$ . The other is in the southern reaches of the LS, where the excited baroclinic tides propagate southwestward. These patterns are shown better below, in the form of the baroclinic tidal energy flux (Fig. 8). The two groups of baroclinic tides interfere with each other in the northern South China Sea. East of the LS, the prevailing eastward-propagating  $K_1$  baroclinic tide is followed by a secondary wave with

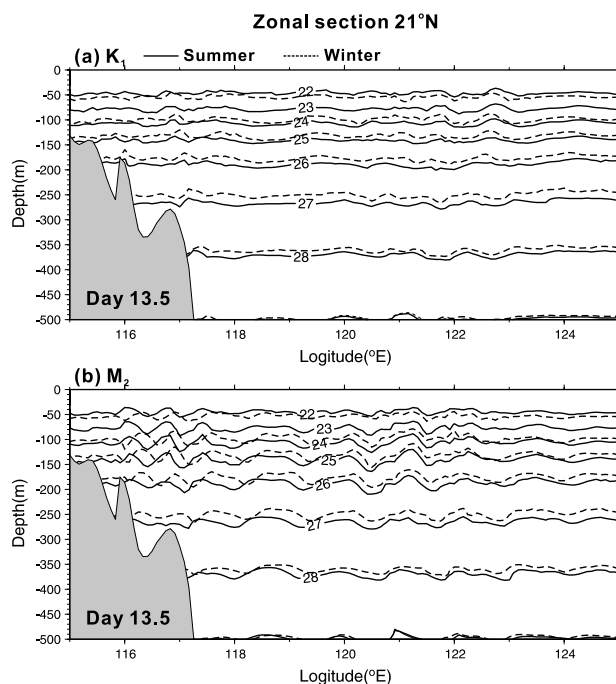


Fig. 6. Model-produced instantaneous (day 13.5) density transect along  $21^\circ\text{N}$  for (a)  $K_1$  and (b)  $M_2$  constituents under summer (solid lines) and winter (dashed lines) stratifications.

smaller amplitude, e.g., as indicated by arrows in Fig. 5(b), two successive wave troughs, separated  $\sim 100$  km apart, at  $123^\circ\text{E}$  and  $122^\circ\text{E}$  along  $21^\circ\text{N}$ , respectively. Detailed analysis of the model results suggests that the former is generated at the east ridge and the latter at the west ridge. The horizontal wavelengths of diurnal and semidiurnal baroclinic tides are 300 km and 120 km, respectively, estimated subjectively from the distance between two successive prevailing wave crests (or troughs) near the LS. Near Dongsha Island, the wavelength of the semidiurnal internal tide decreases to  $\sim 90$  km due to the shoaling of the bottom topography (Fig. 5(c)).

The model-produced isopycnal fields along  $21^\circ\text{N}$ , the center of the baroclinic tidal beam, for  $K_1$  and  $M_2$  in the two seasons are compared in Fig. 6. The seasonal difference of isopycnal fluctuations occurs mostly above 300 m depth and the westward wave propagation is slightly faster in summer than in winter. Beam-like vertical structures of baroclinic tides emanate from the two major ridges within  $120.5^\circ$  to  $122^\circ\text{E}$  and the northern SCS. The higher vertical mode is seen around the LS, especially  $M_2$ . Propagating away from the generation site, the beam-like feature disappears quickly, suggesting the dominance of low vertical modes of baroclinic tides in the northern SCS and Pacific Ocean as high-mode baroclinic tides are dissipated. The zonal phase speed is estimated to be  $2.94$

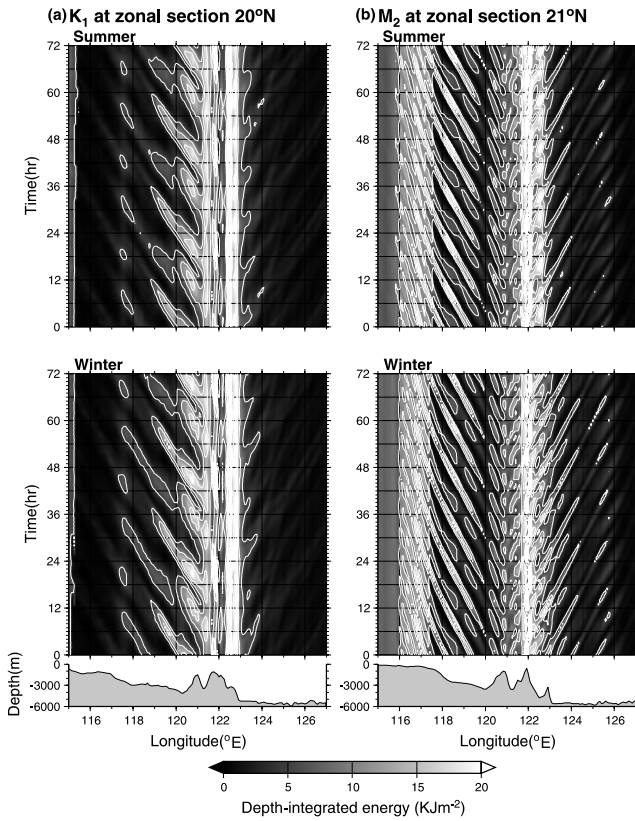


Fig. 7. Temporal and zonal variations of depth-integrated baroclinic tide energy for (a)  $K_1$  along  $20^\circ\text{N}$  and (b)  $M_2$  along  $21^\circ\text{N}$  in summer and winter. Relevant sectional bottom topography is inserted below. Contour interval is  $2.5 \text{ kJ m}^{-2}$ .

to  $3.40 \text{ m s}^{-1}$  for semidiurnal and diurnal constituents (Fig. 5). For the  $K_1$  baroclinic tide, the phase speed propagating into the SCS in winter is 4% slower than that in summer, and the phase speed propagating into the Pacific Ocean in winter is 15% slower than that in summer (Fig. 5(a)). For the  $M_2$  baroclinic tide, the zonal phase speed propagating into the SCS in winter is 14% slower than that in summer, and the phase speed propagating into the Pacific Ocean in winter is 5% slower than that in winter. Near the Dongsha Plateau, the phase speed of the  $M_2$  baroclinic tide decreases significantly,  $\sim 2 \text{ m s}^{-1}$  in summer and  $1.85 \text{ m s}^{-1}$  in winter, close to the horizontal phase speed of the first-mode internal tides:  $1.96 \text{ m s}^{-1}$  in summer and  $1.8 \text{ m s}^{-1}$  in winter. The results of the  $M_2$  baroclinic phase speed and seasonal variations are consistent with those found by Zhao and Alford (2006).

### 3.2 Energy propagation

We evaluated the seasonal variations of the two dominant diurnal and semidiurnal tides. The temporal and zonal variations of the total depth-integrated baroclinic tide

Table 2. Horizontal baroclinic energy speeds estimated from Fig. 7. DS: Dongsha Plateau; SCS: northern South China Sea; PO: northwest Pacific Ocean.

Run	Stratif.	$C_{gx} \text{ (m s}^{-1}\text{)}$		
		DS	SCS	PO
$K_1$ at $20^\circ\text{N}$	Summer	—	2.91	2.59
	Winter	—	2.78	2.55
$M_2$ at $21^\circ\text{N}$	Summer	1.69	2.91	3.06
	Winter	1.56	2.84	2.91

energy for  $K_1$  and  $M_2$  along  $20^\circ\text{N}$  and  $21^\circ\text{N}$  are shown in Fig. 7. The two latitudes are chosen where wave crests are running nearly north-south in Figs. 5(b) and (c). The zonal component of the horizontal energy propagation speed of the baroclinic tide is computed as the averaged inverse slope of the peak energy line on Fig. 7 and is summarized in Table 2. The  $K_1$  baroclinic energy propagates westward at  $\sim 2.9 \text{ m s}^{-1}$  in summer and  $\sim 2.8 \text{ m s}^{-1}$  in winter west of the LS, and propagates eastward into the Pacific Ocean at  $\sim 2.6 \text{ m s}^{-1}$  in summer and winter. The  $M_2$  baroclinic energy propagates eastward at a uniform speed,  $\sim 3.0 \text{ m s}^{-1}$ , from the LS to as far as  $126^\circ\text{E}$ . In the deep northern SCS basin, the mean zonal speed of westward-propagating  $M_2$  baroclinic energy is  $\sim 2.9 \text{ m s}^{-1}$ . The westward propagation speed decreases dramatically ( $1.69$  and  $1.56 \text{ m s}^{-1}$  in summer and winter) west of the continental slope at  $118^\circ\text{E}$ . Most of the  $M_2$  baroclinic tidal energy is dissipated east of  $116^\circ\text{E}$  (Fig. 7(b)), which is consistent with a previous conclusion by Chang *et al.* (2006). Figure 7 also suggests the complicated baroclinic tide generation, which comprises both barotropic and baroclinic processes in the LS.

The eigen-mode speed  $C_1$  of the first vertical mode  $M_2$  baroclinic tides, computed using the initial vertical stratification profiles, is  $2.82 \text{ m s}^{-1}$  in summer and  $2.60 \text{ m s}^{-1}$  in winter. Alford *et al.* (2006) showed that  $C_n^2 = C_g C_p$ , where  $C_n$  is speed of the  $n$ -th vertical mode,  $C_g$  is the group speed, and  $C_p$  is the phase speed. For linear internal tides,  $C_g = C_p(\omega^2 - f^2)\omega^{-2} \approx 0.85 C_p$  at the semidiurnal frequency. The modeled group speed is faster than the theoretical gravest mode speed and close to the phase speed. The discrepancy is probably due to the effect of nonlinearity and the superposition of eigen-modes.

When the  $M_2$  baroclinic tide enters the shallow Dongsha Plateau, the group speed decreases to  $1.69 \text{ m s}^{-1}$  in summer and  $1.56 \text{ m s}^{-1}$  in winter. Large-amplitude nonlinear internal waves are found near Dongsha Plateau and may be converted from the shoaling internal tides (Lien *et al.*, 2005). Long-term mooring observations of nonlinear internal waves on the Dongsha Plateau show that the propagation speed of nonlinear internal waves is

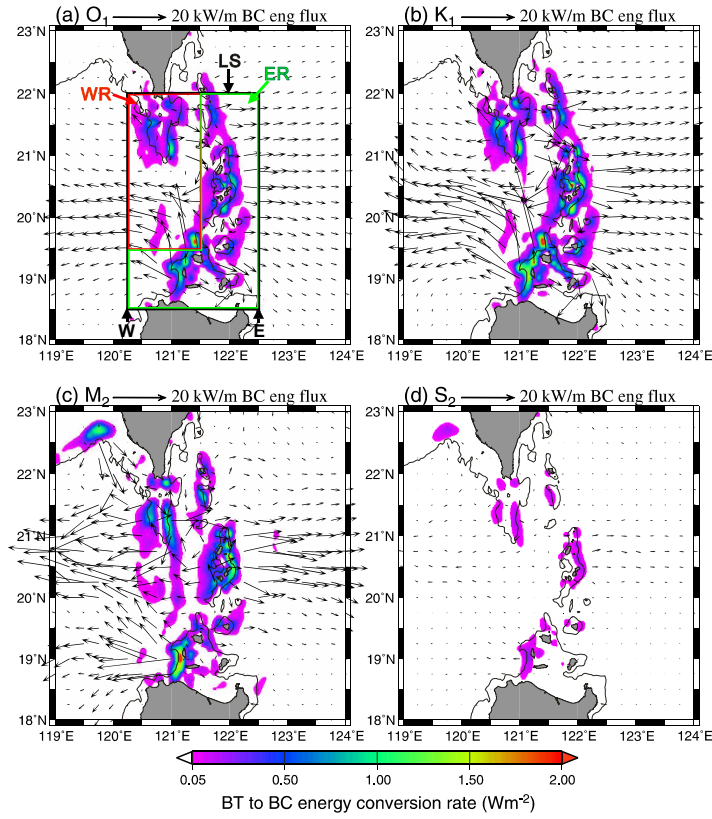


Fig. 8. Spatial distribution of depth-integrated barotropic-to-baroclinic energy conversion rate and baroclinic energy flux averaged over the last three days of simulation for (a)  $O_1$ , (b)  $K_1$ , (c)  $M_2$  and (d)  $S_2$  constituents in summer. 1000 m isobath is appended for reference.

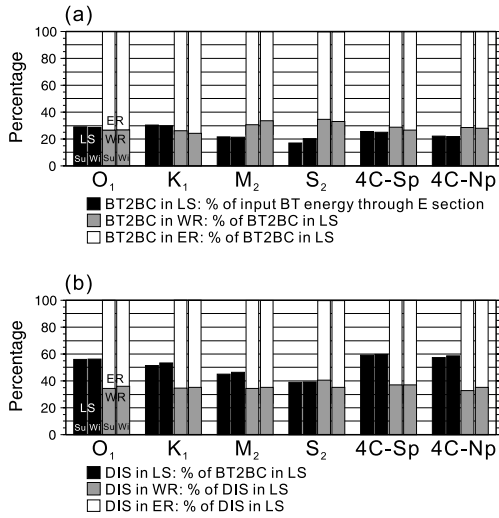


Fig. 9. (a) Percentages of barotropic-to-baroclinic conversion energy (BT2BC) in rectangular box (LS) to input barotropic energy through section E in Fig. 8 and BT2BC in the regions covering west and east ridges (WR and ER in Fig. 8) to BT2BC in LS. (b) Percentages of dissipation energy (DIS) to BT2BC in LS and DIS in regions covering WR and ER to DIS in LS.

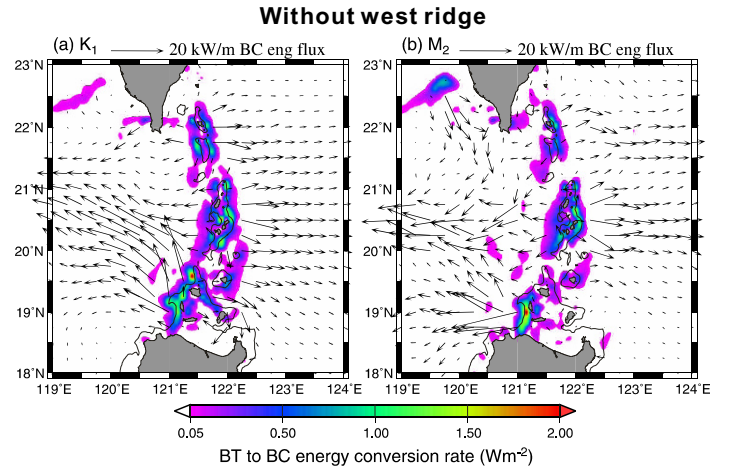


Fig. 10. As Fig. 8 but with ridge height of west ridge relative to 3000 m depth multiplied by a factor 0.1 for (a)  $K_1$  and (b)  $M_2$ . 1000 m isobath is appended for reference.



fastest  $\sim 2 \text{ m s}^{-1}$  in the summer and slowest  $\sim 1.6 \text{ m s}^{-1}$  in winter. Zhao and Alford (2006) also found the seasonal variation of  $M_2$  baroclinic tide propagation speed, faster in summer  $2.5 \text{ m s}^{-1}$  and slower  $2.2 \text{ m s}^{-1}$  in winter, averaged between the LS and Dongsha Plateau, computed using the climatology stratification.

### 3.3 Energetics

The energetics of the model-produced baroclinic tides are discussed as follows. Following the formulation in Niwa and Hibiya (2004), the depth-integrated barotropic ( $F_{bt}$ ) and baroclinic ( $F_{bc}$ ) energy fluxes and the barotropic-to-baroclinic (bt-to-bc) energy conversion rate ( $E_{bt2bc}$ ) averaged over the last three days of simulations are calculated by

$$F_{bt} = \frac{1}{T} \int_0^T \left( \langle \bar{U} \rangle \int_{-H}^{\eta} (g\rho_0(\zeta - \beta\eta) + P') dz \right) dt, \quad (4)$$

$$F_{bc} = \frac{1}{T} \int_0^T \left( \int_{-H}^{\eta} \bar{U}' P' dz \right) dt, \quad (5)$$

$$E_{bt2bc} = \frac{1}{T} \int_0^T \left( g \int_{-H}^{\eta} \rho' w_{bt} dz \right) dt, \quad (6)$$

where  $\rho_0$  represents the initial background density stratification,  $\bar{U}'$  is the baroclinic velocity vector in the Cartesian coordinate,  $P'$  is perturbation pressure,  $T$  ( $=3 \text{ d}$ ) is period of average,  $\rho'$  is perturbation density and

$$w_{bt} = \langle U_x \rangle \left( \sigma \frac{\partial D}{\partial x} + \frac{\partial \eta}{\partial x} \right) + \langle U_y \rangle \left( \sigma \frac{\partial D}{\partial y} + \frac{\partial \eta}{\partial y} \right) + \left( \sigma \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial t} \right) \quad (7)$$

is the Cartesian vertical velocity relevant to the barotropic flow. Results obtained under the summer stratification are shown in Fig. 8. The baroclinic energy of  $S_2$  is the weakest of all constituents. Spatial distributions of the conversion rates and baroclinic energy fluxes suggest distinct effects of the east and west ridges on generating diurnal and semidiurnal baroclinic tides. Considering the barotropic tidal current and topography indicated in Fig. 4, and examining the region of high area-integrated bt-to-bc conversion rate in Fig. 8, the strong tidal current and topography interaction occurs mostly over the east ridge around the Batan Islands.

The baroclinic tidal energy fluxes of  $O_1$  and  $K_1$  propagate eastward into the Pacific Ocean in a tidal beam  $\sim 300 \text{ km}$  wide, comparable to the meridional extent of the east ridge in the LS, centered at  $20.5^\circ\text{N}$  (Figs. 8(a)

and (b)). The energy conversion from the barotropic tides to the baroclinic tides occurs along the entire east ridge at a rate of  $>0.05 \text{ W m}^{-2}$  along most of the ridge. The energy flux of  $K_1$  is slightly greater than that of  $O_1$ , with an average flux of  $\sim 10 \text{ kW m}^{-1}$  immediately east of the LS. Propagating westward into the SCS from the LS, the baroclinic tidal energy fluxes of  $O_1$  and  $K_1$  emanate mostly from the southern end of the east ridge in a narrower ( $150 \text{ km}$ ) tidal beam centered at  $20^\circ\text{N}$ . The westward energy flux is about  $15 \text{ kW m}^{-1}$ , greater than that of the eastward energy flux; it propagates northward first between the two ridges, and then turns to the west. The meridionally integrated eastward baroclinic energy fluxes of  $O_1$  and  $K_1$  into the Pacific Ocean are comparable to the meridionally integrated westward energy fluxes into the SCS, with different tidal beam widths. The northern part of the west ridge is a secondary generation site ( $\sim 26\%$ ) of diurnal baroclinic tides.

The conversion of the  $M_2$  barotropic tide to the baroclinic tide occurs on both the east and west ridges at a rate similar to that of diurnal tide. The eastward-propagating  $M_2$  baroclinic energy has a tight beam width of  $125 \text{ km}$ , between  $20$  and  $21.25^\circ\text{N}$  centered at  $20.5^\circ\text{N}$  (Fig. 8(c)). The maximum eastward energy flux into the Pacific Ocean reaches  $\sim 20 \text{ kW m}^{-1}$  with an average of  $10 \text{ kW m}^{-1}$  immediately east of the LS. The  $M_2$  baroclinic energy flux west of the LS into the SCS is generated at two primary sites, the east ridge south of  $21^\circ\text{N}$  and the west ridge north of  $21^\circ\text{N}$ . These two energy fluxes merge into a nearly  $125 \text{ km}$  tidal beam propagating westward. A sizable fraction of the baroclinic energy that emanated from the southern part of east ridge propagates southwestward centered at  $19^\circ\text{N}$  (Fig. 8(c)).

The modeled baroclinic tidal energy flux in the LS is nearly ten times that in the deep basins at the western Pacific and near Hawaii,  $1\text{--}2 \text{ kW m}^{-1}$ , estimated using historical hydrographic and mooring measurements (Alford, 2003). It is comparable to the observed strong semidiurnal internal tidal energy flux,  $O(10) \text{ kW m}^{-1}$ , radiating from the Hawaiian Ridge (Lee *et al.*, 2006).

The energy budgets averaged over the last three days of the model simulations in the LS, i.e., the rectangular box in Fig. 8(a), for all four constituents in two seasons are analyzed and summarized in Table 3. The 3-day averaged, depth-integrated baroclinic dissipation rate is computed as the difference of the divergence of baroclinic energy flux and the bt-to-bc energy conversion rate, i.e.  $\nabla \cdot F_{bc} - E_{bt2bc}$ . The sectional-integrated barotropic energy in the east section ranges from approximately  $27$  to  $45 \text{ GW}$  ( $1 \text{ GW} = 10^9 \text{ W}$ ) for the three major constituents,  $O_1$ ,  $K_1$  and  $M_2$  and exhibit little seasonal variation. Both diurnal and semidiurnal barotropic tidal energy fluxes propagate westward from the Pacific Ocean, across the LS, and into the SCS.

Table 3. Energetics in the Luzon Strait. TS: initial temperature and salinity profiles of summer (Su) or winter (Wi); Tid: tidal stage for spring (Sp) or neap (Np) tides; BTE: meridional-section integrated barotropic energy at the east (E) and west (W) sections of the rectangular box in Fig. 8; BT2BC: area-integrated barotropic-to-baroclinic energy conversion rate in the Luzon Strait (LS) and the regions over the west (WR) and east (ER) ridges; BCE: meridional-section integrated baroclinic energy at E and W sections; ~DIS: area-integrated approximately energy dissipation (divergence of the baroclinic energy flux – BT2BC) over LS, WR and ER. 4C means the four constituents composed together. Negative values of BTE and BCE mean westward. Units are GW ( $10^9$  W). All quantities are averaged over three days of simulation.

Run	TS	Tid	BTE		BT2BC			BCE		~DIS		
			W	E	LS (%BTE-E)	WR (%LS)	ER (%LS)	W	E	LS (%BT2BC)	WR (%LS)	ER (%LS)
$O_1$	Su	—	–28.43	–27.11	7.89 (29.1)	2.10 (26.6)	5.79 (73.4)	–1.56	1.73	4.41 (55.9)	1.52 (34.5)	2.89 (65.5)
	Wi	—	–28.05	–26.80	7.65 (28.5)	2.06 (26.9)	5.59 (73.1)	–1.54	1.65	4.30 (56.2)	1.55 (36.0)	2.75 (64.0)
$K_1$	Su	—	–34.00	–36.92	11.24 (30.4)	2.92 (26.0)	8.32 (74.0)	–2.73	2.43	5.79 (51.5)	2.01 (34.7)	3.78 (65.3)
	Wi	—	–33.39	–37.84	11.33 (30.0)	2.74 (24.2)	8.59 (75.8)	–2.75	2.26	6.03 (53.2)	2.12 (35.2)	3.91 (64.8)
$M_2$	Su	—	–32.45	–44.01	9.53 (21.7)	2.93 (30.7)	6.60 (69.3)	–3.02	2.30	4.30 (45.1)	1.48 (34.4)	2.82 (65.6)
	Wi	—	–30.51	–45.68	9.72 (21.3)	3.27 (33.6)	6.45 (66.4)	–2.88	2.03	4.50 (46.3)	1.59 (35.3)	2.91 (64.7)
$S_2$	Su	—	–3.27	–7.06	1.21 (17.1)	0.42 (34.7)	0.79 (65.3)	–0.39	0.36	0.47 (38.8)	0.19 (40.4)	0.28 (59.6)
	Wi	—	–3.59	–6.40	1.30 (20.3)	0.43 (33.1)	0.87 (66.9)	–0.39	0.41	0.51 (39.2)	0.18 (35.3)	0.33 (64.7)
4C	Su	Sp	–177.32	–196.09	50.35 (25.7)	14.50 (28.8)	35.85 (71.2)	–10.73	9.28	29.75 (59.1)	11.00 (37.0)	18.75 (63.0)
	Wi	Sp	–175.28	–199.57	49.86 (25.0)	14.27 (28.6)	35.58 (71.4)	–10.40	8.99	29.77 (59.7)	11.05 (37.1)	18.72 (62.9)
	Su	Np	–42.69	–50.00	11.07 (22.1)	2.96 (26.7)	8.10 (73.3)	–2.63	2.00	6.33 (57.2)	2.07 (32.7)	4.26 (67.3)
	Wi	Np	–41.43	–50.97	11.23 (22.0)	3.16 (28.1)	8.07 (71.9)	–2.53	2.04	6.58 (58.6)	2.32 (35.3)	4.26 (64.7)

About 30% of the westward incident diurnal barotropic tidal energy converts to the baroclinic tide, and about 21% of the incident semidiurnal barotropic tidal energy converts to baroclinic tide (Fig. 9). Note that the incident barotropic tidal energy here means the zonal component of energy flux through the east section E in Fig. 8, because it is the most effective component in generating baroclinic tides over the meridional ridges. The model conversion rate for  $M_2$  is close to the result of Niwa and Hibiya (2004). Nearly 19 GW ( $O_1 + K_1$ ) and 11 GW ( $M_2 + S_2$ ) of diurnal and semidiurnal baroclinic tides, respectively, are generated in the LS. About 30% is generated at the west ridge and 70% at the east ridge (Fig. 9). About 4 GW of baroclinic diurnal tides propagates into the SCS and 4 GW into the Pacific Ocean. About 3 GW of baroclinic semidiurnal tides propagates into the SCS and 3 GW into the Pacific Ocean. Nearly 50% of baroclinic tides generated in the LS is dissipated locally (Fig. 9): 10 GW for baroclinic diurnal tides and 5 GW for the baroclinic semidiurnal tides (Table 3). Compared with

the results of baroclinic tide study along the Hawaiian ridge, Klymak *et al.* (2006) found 8–25% of baroclinic tidal energy, 2–4.5 GW, is dissipated locally, and the rest propagates away. Our model result of 50% local dissipation of baroclinic tidal energy is consistent with the result of Niwa and Hibiya (2004), and is likely caused by the effect of double ridges. Baroclinic tides generated at one ridge interact with the other ridge within a short distance, <100 km, from their generation sites. The dissipated baroclinic tidal energy may be the source of the local turbulence mixing, which is potentially important for mixing water masses in the Pacific Ocean, Kuroshio, and the SCS.

The 3-day averaged, LS area-integrated barotropic tide energy calculated from the divergence of the barotropic energy flux indicates that the  $K_1$  barotropic energy is  $\sim$ –12 GW (negative means energy loss) which is balanced by the bt-to-bc energy conversion ( $\sim$ 11 GW) and the dissipation due to friction ( $\sim$ 1 GW). A similar estimate of  $M_2$  indicates that  $\sim$ 18 GW of barotropic tide

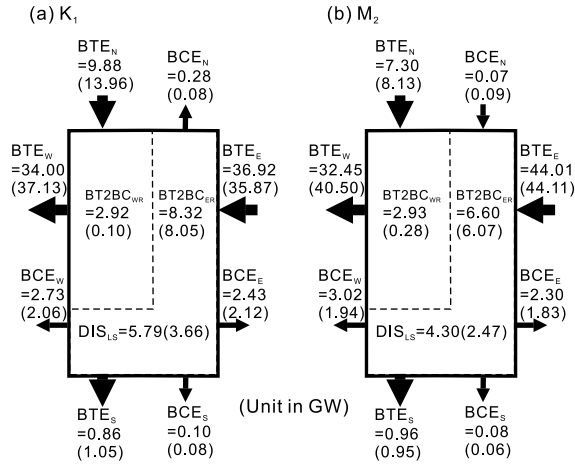


Fig. 11. Diagram showing 3-day averaged, area-integrated baroclinic energy dissipation (DIS) and barotropic-to-baroclinic conversion energy (BT2BC) over west ridge (WR) and east ridge (ER) and meridional sectional area-integrated barotropic (BT) and baroclinic (BC) tide energy through east (E), west (W), north (N) and south (S) boundaries of the LS for (a)  $K_1$  and (b)  $M_2$  on topography with or without the west ridge. Numbers in parentheses indicate corresponding energetics calculated from model runs without the west ridge. All quantities are GW.

energy is lost in the LS, which is balanced by  $\sim 10$  GW of bt-to-bc conversion and  $\sim 8$  GW of local barotropic dissipation. These estimates are schematically shown in Fig. 11 below.

### 3.4 Role of west ridge

Chao *et al.* (2007) identified the west ridge in the middle portion of the LS as a damper of incoming  $M_2$  baroclinic tide from the east ridge; while in the northern portion of the LS, the west ridge becomes a secondary generation site for  $M_2$  baroclinic tide. Based on the model results, Jan *et al.* (2007) concluded that the east ridge is the major source of  $K_1$  baroclinic tide in the SCS. To further evaluate the effect of the west ridge in generating and dissipating baroclinic tides we conducted two additional experiments with the height of the west ridge relative to 3000 m depth reduced by 90%, i.e., deepening the west ridge artificially to  $\sim 3000$  m, leaving all other model settings unchanged for  $K_1$  and  $M_2$ . Hereafter we refer to this case as “without west ridge” for convenience. Figure 10 illustrates the 3-day averaged, depth-integrated bt-to-bc conversion energy and baroclinic energy fluxes for  $K_1$  and  $M_2$ . If one compares Figs. 8 and 10, one sees that an unrealistic deep west ridge suppresses the generation of baroclinic tides along the northern west ridge. Without the west ridge, the spatial distribution of  $K_1$  baroclinic energy flux east of the east ridge (Fig. 10(a)) differs little

from that with the west ridge (Fig. 8(b)); the  $M_2$  baroclinic energy beam changes from one (Fig. 8(c)) to two separated beams (Fig. 10(b)). West of the east ridge, the spatial structure of baroclinic energy fluxes for both  $K_1$  and  $M_2$  differ significantly from those shown in Fig. 8. The westward-propagating  $K_1$  beam in Fig. 8(b) is shifted  $0.5^\circ$  north and becomes narrower in Fig. 10(a). The west ridge apparently serves as a block, keeping the westward  $K_1$  energy beam centered at  $20.5^\circ$ N. The width of the westward  $M_2$  baroclinic tidal energy beam (Fig. 10(b)) narrows to about two thirds of that in Fig. 8(c), which implies that the west ridge is more important in generating westward propagating  $M_2$  baroclinic tides than  $K_1$  baroclinic tides.

Figure 11 compares the area-integrated bt-to-bc energy, sectional area-integrated baroclinic tidal energy fluxes, and dissipation of baroclinic tidal energy calculated from the model runs with and without the west ridge for  $K_1$  and  $M_2$  under the summer stratification. The eastward baroclinic energy through section E in Fig. 8(a) decreases to 2.12 and 1.83 GW for  $K_1$  and  $M_2$  without the west ridge, which corresponds to 12.7% and 20.4% reduction of that with the west ridge. Through the west section W in Fig. 8(a), the westward  $K_1$  and  $M_2$  baroclinic energy drops to 2.06 and 1.94 GW, corresponding to a decrease of 20.8% and 35.7% of that with the west ridge. The west ridge plays a more significant role in modulating the  $M_2$  baroclinic tide than modulating the  $K_1$  baroclinic tide in SCS. Without the west ridge, the area-integrated baroclinic tidal energy dissipation in the LS reduces to 36.7% (5.79 vs. 3.66 GW) and 42.5% (4.30 vs. 2.47 GW) of that with the west ridge for  $K_1$  and  $M_2$  baroclinic tides. The higher reduction rates of baroclinic tidal energy and dissipation for  $M_2$  indicate that the west ridge is more efficient in generating and dissipating  $M_2$  baroclinic tides than  $K_1$  baroclinic tides. The reduction of the baroclinic tidal energy dissipation rate in the LS without the west ridge suggests the importance of the double ridges in the LS in dissipating baroclinic tides.

### 3.5 Properties of combined constituents

Model simulations of the combined four tidal constituents, which are closer to reality, are performed to compare the spring-neap variations. The depth-integrated bt-to-bc conversion rate and baroclinic energy fluxes are shown in Fig. 12. Again, there is no significant seasonal variation, but strong spring-neap variation. The westward-propagating barotropic energy entering the LS amounts to as much as 200 GW, approximately 25% of which is converted to baroclinic energy in the LS during the spring tide; the incident barotropic energy drops greatly to 11 GW, 22% of which is converted to baroclinic energy during the neap tide (Table 3). During the spring tide, over most of the west and east ridges,  $>0.05 \text{ W m}^{-2}$  of

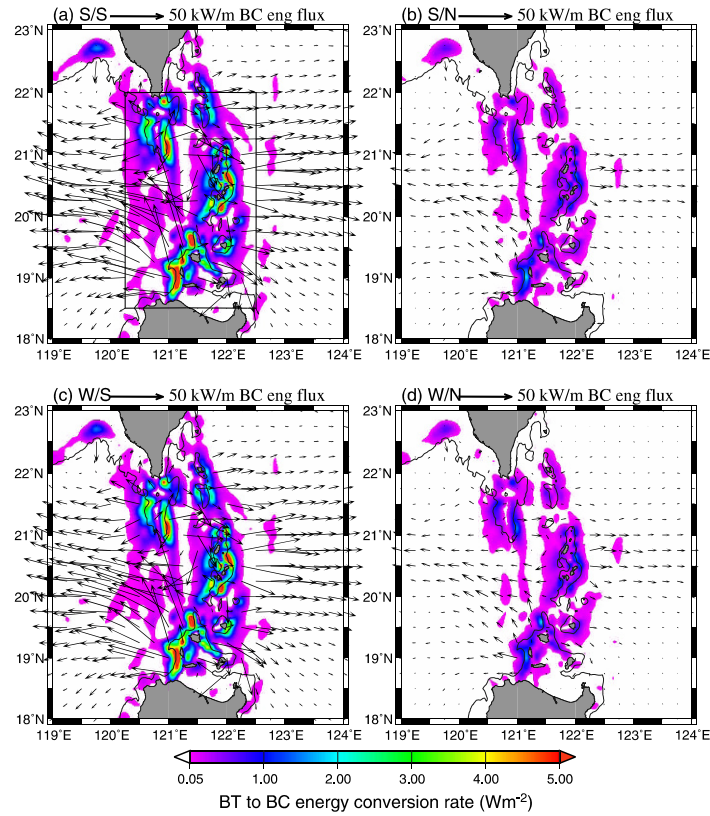


Fig. 12. As Fig. 5 but for the four constituents combined together during (a) spring (b) neap tides under summer stratification.

baroclinic tidal energy is generated, with some hot spots of  $>5 \text{ W m}^{-2}$ . Two nearly 300 km wide tidal beams emanate from the Luzon strait into the SCS and the Pacific Ocean, centered at  $\sim 20.5^\circ\text{N}$  with an average energy flux of  $\sim 50 \text{ kW m}^{-1}$  close to the generation site.

During the spring tide, integrated over the LS,  $\sim 50$  GW of baroclinic tidal energy is generated in a spring tide, 28% at the west ridge,  $\sim 10$  GW propagates into the SCS,  $\sim 9$  GW propagates into the Pacific Ocean, and  $\sim 30$  GW is dissipated locally, 40% of it at the west ridge (see Table 3 and Fig. 9). During the neap tide,  $\sim 11$  GW of baroclinic tidal energy is generated,  $\sim 5$  GW propagates away, and  $\sim 6$  GW is dissipated locally. Although the magnitudes of energetics show a factor of five in spring-neap variations, the ratios of the area-integrated bt-to-bc energy conversion rate to the incident barotropic energy, and the baroclinic dissipation to the bt-to-bc conversion rate remain nearly the same in the LS (Fig. 9). When averaged over the spring-neap cycle, the meridionally integrated baroclinic tidal energy flux is 6.6 GW into the South China Sea and 5.6 GW into the Pacific Ocean (Fig. 1(b)). Approximately 58% of baroclinic tidal energy,  $\sim 18$  GW, generated in the Luzon Strait dissipates locally. For comparison, most of baroclinic tidal energy generated in other topographic features in the ocean propagates away

and little is dissipated locally. The strong baroclinic tidal energy dissipation in the LS is more than five times that estimated in the vicinity of the Hawaiian ridge, which might be attributed to the interaction between baroclinic tides and the complicated double-ridge topography.

### 3.6 Turbulence dissipation

Nearly 18 GW of baroclinic tidal energy is dissipated in the LS. The turbulence kinetic energy dissipation rate per unit water mass,  $\varepsilon$ , averaged within the LS is  $\sim 10^{-7} \text{ W kg}^{-1}$ , a factor 100 greater than that in a typical open ocean. The vertical eddy diffusivity can be computed following the Osborn method (Osborn, 1980) as  $K_\rho = 0.2\varepsilon N^{-2}$ . Given the depth averaged stratification  $N = 0.004 \text{ s}^{-1}$ , a bulk estimate of  $K_\rho$  is  $O(10^{-3}) \text{ m}^2\text{s}^{-1}$ , which is 100 times that in the abyssal ocean with typical internal wave field (Toole *et al.*, 1994). The estimated vertical diffusivity is similar to that computed from the hydrographic data (Qu *et al.*, 2006).

In the close vicinity of generation sites, baroclinic tides may dissipate along the tidal beam via shear instability (Lueck and Mudge, 1997; Lien and Gregg, 2001). Lien and Gregg (2001) found  $\varepsilon = O(10^{-6}) \text{ W kg}^{-1}$  and  $K_\rho > 0.01 \text{ m}^2\text{s}^{-1}$  in a thin 50-m layer along the  $M_2$  baroclinic tidal beam within 4 km from the shelf break



where baroclinic tides are generated. Although the shear instability can dissipate high-mode internal tides effectively, but it occurs only in a small fraction of the water column and dissipates only a small percentage of the total baroclinic tidal energy. For example, only 8–25% of baroclinic tides generated along the Hawaiian ridge is dissipated locally (Klymak *et al.*, 2006).

Identifying the dynamic mechanisms responsible for the strong dissipation rate (50%) of baroclinic tides generated in the LS is beyond the scope of this numerical model study. Intuitively, we expect that the complex double ridge topography, within ~100 km, interacts strongly with baroclinic tides in the LS. When baroclinic tides are generated at one ridge and propagate onto the other ridge, the shoaling topography could enhance the baroclinic tidal shear in the interior and near the bottom. Small-scale processes such as hydraulic jumps might occur over the rough topography. Baroclinic tides may become nonlinear and convert into internal solitary waves (Lien *et al.*, 2005). Lien *et al.* (2005) estimated a ~16% conversion rate from baroclinic tides to internal solitary waves on a shoaling continental slope in the northern SCS. Since internal solitary waves are not resolved in the present numerical model, the conversion into internal solitary waves is represented as the local loss of baroclinic tidal energy. All these small-scale processes could be responsible for the large dissipation rate of baroclinic tides in the LS. Further numerical models or observation campaigns are needed to understand the exact dissipation mechanisms of baroclinic tides in the LS.

#### 4. Summary

The temporal and spatial variations of baroclinic tides in the LS have been investigated using a three-dimensional tide model driven by four principal tidal constituents,  $O_1$ ,  $K_1$ ,  $M_2$  and  $S_2$ . Simulations were performed with four individual constituents and with the combined tide separately, and initialized with different seasonal vertical stratifications.

The model results allow us to conclude that there is little seasonal variation in the baroclinic energy budget. The baroclinic tidal propagation speed, including group, phase, and eigen-mode speed, is about 10% greater in summer than in winter. About 20 GW of baroclinic diurnal tide and ~10 GW of baroclinic semidiurnal tide are generated in the LS, with ~30% on the west ridge and 70% on the east ridge. The semidiurnal baroclinic tide has a narrow energy beam of ~125 km compared to ~300 km for the diurnal baroclinic tide. The semidiurnal baroclinic tidal energy flux emanating westward from the east ridge and that from the northern west ridge combine to form a strong westward narrow tidal beam propagating across the SCS basin onto the Dongsha Plateau. This semidiurnal baroclinic energy flux is likely the energy

source for the large-amplitude nonlinear internal waves found in the SCS. Nearly 50% of the baroclinic tidal energy, ~18 GW on average, generated in the LS dissipates locally due to the unique double-ridge effect in the LS. This is distinctly more dissipative than baroclinic tides generated in other topographic features in the ocean, which is often <25%. The strong baroclinic tidal energy dissipation in the LS is more than five times that estimated in the vicinity of the Hawaiian ridge.

The spring-neap variation of baroclinic energy budget is significant. In the spring tide, nearly 50 GW of the combined baroclinic tidal energy is generated in the LS, 20 GW propagates away, and 30 GW dissipates locally; while in neap tide, merely 11 GW is generated, ~4.5 GW propagates away, and ~6.5 GW dissipates locally. The strong dissipation of baroclinic tidal energy may provide energy for the mixing of water masses among the Pacific Ocean, Kuroshio, and the SCS. Our results suggest that the mixing should exhibit significant spring-neap variation. Within the LS, the average turbulence kinetic energy dissipation rate  $\varepsilon$  is  $O(10^{-7})$  W kg<sup>-1</sup> and the turbulence diffusivity is  $O(10^{-3})$  m<sup>2</sup>s<sup>-1</sup>, a factor 100 greater than those in the typical open ocean. Note that the Kuroshio is likely to play an important role in modulating the baroclinic tide's generation, trapping, dissipation, and propagation (Chao *et al.*, 2007). Excluding the Kuroshio helps clarify our model results concerning the barotropic and baroclinic energy budgets, but the model is undoubtedly as over simplification. Further investigations including the Kuroshio effect are needed.

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