Enhanced Diapycnal Mixing in the South China Sea

JIWEI TIAN, Q INGXUAN YANG, AND WEI ZHAO

Physical Oceanography Laboratory, Ocean University of China, Qingdao, China

(Manuscript received 16 August 2007, in final form 2 April 2009)

ABSTRACT

Profiles of current velocity, temperature, and salinity were obtained in the Internal Wave and Mixing Experiment in the South China Sea (SCS), the Luzon Strait, and the North Pacific. The observations are examined for evidence of enhanced diapycnal mixing in the SCS, which reaches $O(10^{-3} \text{m}^2 \text{s}^{-1})$ in magnitude. Results from independent casts reveal that diapycnal diffusivity in the SCS and the Luzon Strait is elevated by two orders of magnitude over that of the smooth bathymetry in the North Pacific, which are typical of background values in an open ocean. The vertical distribution of diapycnal diffusivity is nonuniform in the SCS, exhibiting higher values at depths greater than about 1000 m. This result compares favorably with the direct microstructure measurements at four stations in the SCS. Velocity and density profiles are combined to estimate the internal tide energy flux generated in the Luzon Strait and directed into the SCS. The energy amounts to 10 GW, most of which is rationalized to be the potential energy source for enhanced mixing in the SCS.

1. Introduction

Diapycnal mixing is an important but poorly understood dynamic process in the ocean. The global ocean circulation results from downwelling at a few selected regions in the North Atlantic and the Southern Ocean and upwelling elsewhere in the ocean (Munk and Wunsch 1998). The upwelling occurs because of the decrease in density of the cold, deep water caused by a downward mixing of heat across the thermocline. After upwelling into the upper ocean, the water can flow back to the original location of deep-water formation in the North Atlantic, thus completing the process of global ocean circulation. Therefore, diapycnal mixing could be a vital mechanism controlling the distribution of physical properties, and the concentration of nutrients, dissolved gases, and particulate matter in the water (Mackinnon and Gregg 2003). In particular, turbulent mixing in marginal seas bordering the world oceans makes an important contribution not only to local physical processes, but also to global ocean circulation and heat transport. The latter process could in turn impact significantly on the climate through water exchange between the marginal seas and the adjacent oceans (Rahmstorf 2003).

Munk and Wunsch (1998) speculated that upward advection of cold water is balanced by a mechanically driven downward diffusion of heat, resulting in the observed stratification below the main thermocline. The magnitude of diapycnal mixing necessary to maintain abyssal stratification was estimated by Munk and Wunsch (1998) to be equivalent to a globally averaged diapycnal diffusivity of $10^{-4} \text{m}^2 \text{s}^{-1}$. In recent years, intense localized mixing with diapycnal diffusivities higher than $10^{-5} \text{m}^2 \text{s}^{-1}$ have been found to occur over rough topographies, such as seamounts (Nabatov and Ozmidov 1988; Kunze and Toole 1997; Lueck and Mudge 1997), ridges (Polzin et al. 1997; Ledwell et al. 2000), canyons (St. Laurent et al. 2001; Carter and Gregg 2002), and hydraulically controlled passages between basins (Roemmich et al. 1996; Polzin et al. 1996; Ferron et al. 1998). In these areas, turbulent mixing has been inferred to be 100–1000 times larger than that of the ocean interior. However, both direct microstructure measurements (Gregg 1987; Gregg 1989; Polzin et al. 1995) and results from tracer release experiments (Ledwell et al. 1993; Ledwell et al. 1998) suggest that diapycnal diffusivities in much of the abyssal ocean are an order of magnitude smaller than the value predicted by most models of ocean circulation (Munk and Wunsch 1998). Recently it has been argued that wind-driven upwelling in the Southern
Ocean could reduce the requirement for globally averaged diapycnal diffusivity to $3 \times 10^{-5} \text{ m}^2 \text{s}^{-1}$ (Webb and Suginobara 2001). Nevertheless, substantially higher diffusivity is still required for some regions.

Compared with the open oceans, less attention has been paid to the marginal seas. A study by Carter et al. (2005) revealed the existence of a complex mixing environment on the Monterey Bay continental shelf. Depth- and time-averaged dissipation rates and diapycnal diffusivities were higher than those of the observations made over other continental shelves with smooth topography, but they were below those influenced by prominent topographic features. The most surprising finding was the presence of large and downslope-propagating nonlinear internal solitary-like waves of elevation with high aspect ratios. Mackinnon and Gregg (2003) conducted observations on the outer New England shelf in the late summer of 1996 as part of the Coastal Mixing and Optics Experiment. They found that dissipation rates and diffusivity were both highly episodic. The average dissipation rate within the bottom boundary layer was $1.2 \times 10^{-7} \text{ W kg}^{-1}$, and it varied in magnitude with the strength of the near-bottom flow from the barotropic tide and low-frequency internal waves. They considered that half of the thermocline dissipation was due to the strong shear and strain within six solibores, which lasted cumulatively less than 1 day, but contained 100-fold elevation in dissipation and diffusivity. The midcolumn dissipation correlated strongly with shear originated from low-frequency internal waves.

The South China Sea (SCS), which connects with the Pacific through the Luzon Strait, is one of the largest marginal seas of the Pacific. Large numbers of internal waves are induced by the prominent bathymetry in the Luzon Strait (Niwa and Hibiya 2004; Jan et al. 2007; Tian et al. 2003; Tian et al. 2006), which propagate into the SCS with great energy flux to furnish the turbulent mixing there. Extensive research pertaining to the barotropic tide, internal waves, and solitons in the SCS has been conducted during the Asian Seas International Acoustics Experiment (ASIAEX). With the deployment of a moored array across the shelf break in the northeast SCS during April–May 2001, Beardsley et al. (2004) revealed that the barotropic tidal currents in this area were mixed (diurnal $O_1$ and $K_1$, semidiurnal $M_2$ and $S_2$). These barotropic energy fluxes were compared with the locally generated diurnal internal tide and high-frequency internal solitary-type waves generated by the $M_2$ flow through the Luzon Strait. In the same area, Duda et al. (2004) observed incident nonlinear internal waves from the east with large amplitudes, which changed amplitude, horizontal length scale, and energy when shoaling. The internal tides were at times sufficiently nonlinear to break down into bores and groups of high-frequency nonlinear waves. Further, Ramp et al. (2004) found that the highly nonlinear internal waves (or solitons) were generated near the Batan Islands in the Luzon Strait and propagated west across deep water to the shelf break. They also reported that these so-called transbasin waves had amplitudes ranging from 29 to greater than 140 m, and were among the largest such waves ever observed in the world’s oceans. Internal solitons northeast of Dongsha Island on the continental slope of the northern SCS were investigated with the current profiler and thermistor chain moorings by Yang et al. (2004); they pointed out that most of the observed internal solitons were first baroclinic mode depression waves. Values were given for the largest horizontal current velocity, vertical displacement, temperature variation induced by the internal solitons, and the nonlinear phase speed.

In recent years, a series of works concerning internal wave and energy flux in the SCS have been reported. Using ADCP measurements taken in the SCS, Lien et al. (2005) suggested that strong internal tides generated in the Luzon Strait propagated as a narrow tidal beam into the SCS. Amplified by the shoaling continental slope near Dongsha Island, they then became nonlinear and evolved into high-frequency nonlinear internal waves. Lien et al. (2005) estimated that the diapycnal diffusivity would be about $10^{-3} \text{ m}^2 \text{s}^{-1}$ within a 200 km $\times$ 200 km area centered at Dongsha Island in the 10-m-thick interface. Chang et al. (2006) showed that both divergences of energy and energy flux of nonlinear internal waves were strong along and across the waves’ prevailing westward propagation path. The energy flux of nonlinear internal wave was the largest on the Dongsha Plateau, while those to the west and north of the plateau can be ignored. The divergence of the average energy flux of the nonlinear internal wave along the wave path on the plateau was examined, giving a corresponding diapycnal diffusivity of $4 \times 10^{-3} \text{ m}^2 \text{s}^{-1}$ in 10-m thickness. Klymak et al. (2006) reported the evolution of an energetic nonlinear internal wave packet as it propagated through the deep central basin of the SCS, toward the western slope and shelf. The estimated amplitudes and phase speeds of the wave packet were 170 m and $2.9 \pm 0.1 \text{ m s}^{-1}$. The Korteweg–de Vries (KdV) fit and the satellite-derived estimate of the extent of horizontal wave implied a westward energy flux of 4.5 GW for each crest.

The Luzon Strait is a passage not only for energy flux, but also water transport. Existing evidence suggests that the Pacific water enters the SCS via the deeper layer while it exits via the intermediate layer of Luzon Strait (Tian et al. 2006). Deep-water mass of the SCS, which
emanates from the Pacific deep water, spreads along the seafloor and constitutes the primary branch of the SCS circulation. In a thermodynamic steady state, the closure of the SCS circulation is accomplished by vertical mixing and upwelling. Mixing with the warmer overlying water heats up the colder water at depth and lowers its density, resulting in the intrusion of the Pacific deep waters into the SCS through the Luzon Strait (Qu et al. 2006). The notable impact of this intrusion on the Pacific circulation is important, and could be inferred by the work published by Furue and Endoh (2005). Their numerical results showed that about 8 Sv of the deep–bottom water flows into the Pacific, most of the inflow returns to the Southern Ocean at the middepth of about 1000 m, and the residual of 2–3 Sv (1 Sv = 10^6 m^3 s^{-1}) upwells to the near surface in the subpolar region of the Pacific. However, their result was based on the premise that the Luzon Strait was closed in the model. If we open the Luzon Strait, about 2 Sv of deep–bottom water of the North Pacific will enter the SCS through the Luzon Strait, and then return to the North Pacific after upwelling to the intermediate layer driven by mixing in the SCS (Tian et al. 2006). Thus, the SCS acts as an agent, who on one hand transports the Pacific deep water upward, and then releases it to the Pacific on the other. As a result, the upward transport of the North Pacific deep–bottom water in the subpolar region of the Pacific may be reduced to 1–2 Sv. A simple calculation given below demonstrates the influence of the SCS mixing on the Pacific as well. The average diffusivity of the SCS and Pacific (consider the SCS and Pacific as a whole) might be expressed as $K_p = (K_{p,SCS}A_{SCS} + K_{p,Pac}A_{Pac})/A_{Tot}$, where $A_{SCS}$, $A_{Pac}$, and $A_{Tot}$ denote the area of the SCS, the Pacific, and their sum. $K_{p,SCS}$ and $K_{p,Pac}$ are the diffusivities of the SCS and the Pacific. Here, taking $K_{p,Pac} = 10^{-5}$ m^2 s^{-1}, $K_{p,SCS} = 10^{-2}$ m^2 s^{-1}, and $A_{Pac} = O(100A_{SCS})$, the SCS mixing thus could be as important as the entire mixing in the Pacific. This suggests that mixing in the SCS plays a key role in maintaining the strength of abyssal water transport and the Pacific circulation.

In the text that follows, the experimental part, including the study area, water masses, instrumentation, and data collection procedures, will be described first in section 2. An overview of the parameterization is given in section 3, along with data processing routines and methods used to quantify turbulent dissipation and diffusivity. The characteristics of turbulent dissipation and mixing are presented in section 4. An estimation of the internal tide energy flux generated in the Luzon Strait is given in section 5. Subsequently, in section 6, results from available methods will be compared with direct microstructure observations, and this will be followed finally by a summary and discussion in section 7.

2. Field experiment

a. Study area and water masses

Internal Wave and Mixing Experiment (IWME) was initiated for the first time, which focused on the intensity, spatial distribution, and mechanisms of both the internal wave and mixing in the SCS, the Luzon Strait, and the North Pacific, and abyssal water transport through the Luzon Strait. The experiment consisted of five cruises conducted from April 2004 to December 2006 in study areas, which were characterized by smooth abyssal plains in the Pacific, rough ridges in the Luzon Strait, and steep slopes in the SCS (Fig. 1). The $M_2$ barotropic tide was almost normally incident onto the depth of the Luzon Strait where the ridges were located, and the region of incidence was considered as one of the hottest spots of internal tide generation (Niwa and Hibiya 2004). Water transport between the SCS and the Pacific was active through the Luzon Strait. A sandwiched vertical structure of the Luzon Strait transport had been confirmed by direct observation conducted in October 2005, and a westward abyssal water transport of 2 Sv was reported by Tian et al. (2006).

There are three typical water masses in the experiment region of this study: the North Pacific Tropical Water (NPTW), the North Pacific Intermediate Water (NPIW), and the North Pacific Deep Water (NPDW). The first two water masses are discernible evidently in the $\theta - S$ diagram, corresponding to the upper and lower vertexes, respectively (Fig. 2). In the Pacific, NPTW and NPIW correspond respectively to a salinity maximum of 35 and 34.15. The water column shows a reversed “S” shape. While in the SCS, the salinity maximum of NPTW shows a decrease of 0.3, to 34.7, and the salinity minimum of NPIW shows an increase of 0.3, to 34.45. As a result, the water column shows a weaker reverse $S$ than that in the Pacific. The characteristics of NPTW and NPIW change remarkably from the Pacific to the SCS, to which the mixing occurring in the SCS might have made a significant contribution.

b. Instrumentation and data collection

The instruments used in IWME included a 300-kHz lowered acoustic Doppler current profiler (LADCP) manufactured by RD Instruments, Inc., a Seabird 9–11 Plus CTD by SeaBird Electronics, Inc., a Turbulence Ocean Microstructure Acquisition Profiler (TurboMAP)-2 from Alec Electronics Co., Ltd., and a Vertical Microstructure Profiler (VMP)-2000 from RGL Consulting, Ltd. The first two instruments measured the current velocity, temperature, salinity, and pressure. TurboMAP and VMP offered us the turbulent dissipation rate directly at the microstructure scale.
The vertical bin size was set to 8 m, the number of layers was set to 13, and the sampling frequency was set to 1 Hz for LADCP, which collected current velocity with an estimated uncertainty of 1 cm s$^{-1}$. The resolution was 24 Hz, the accuracy of conductivity was 0.0003 S m$^{-1}$, and the temperature was 0.001°C for CTD. LADCP was mounted on a CTD package and the two were deployed together. TurboMAP-2 sampled turbulent shear with an accuracy of 5% and a depth rating of 500 m. VMP-2000 sampled shear with an accuracy of 2% and a depth rating of 2000 m. The analysis of the microstructure data and the estimation of dissipation rate for TurboMAP-2 and VMP-2000 essentially followed the procedure described by Wolk et al. (2002).

An extensive set of oceanographic measurements, consisting of a total of 500 profiles, was collected during five cruises. In the present work, measurements made in two sections (Fig. 1), consisting of 47 stations (97 profiles), were analyzed. Data from other stations were not included in the analysis because they were outside the scope of this study. The following was noteworthy: (i) four profiles of turbulent dissipation rate by VMP-2000 were collected in the SCS (These data extended to 1500 m and thus enabled us to test the parameterization result), and (ii) there were 15 stations densely distributed along the Luzon Strait section with a station spacing of only about 20 km. Of particular interest were the five repeat-occupation stations at which the data records were typically 30 h and the time interval between two successive profiles was only about 3 h. These measurements offered us the opportunity to study the internal tide energy flux that emanated from the Luzon Strait and then propagated toward the SCS. Analysis of the dataset thus helped reveal the energy source for mixing in the SCS. At each station occupation, LADCP and CTD profiles were collected, and this observation was found to be adequate to resolve the problems we will describe later.

3. Finescale parameterizations and data processing

a. Finescale parameterizations

Henyey et al. (1986) presented a simple analytical model for turbulent kinetic energy dissipation that was
in good agreement with their numerical ray-tracing experiments. Scaling for turbulence based on internal wave–wave interaction theory was first validated against oceanic data by Gregg (1989). The most recent incarnation depends on both shear and strain variance as
\[ \epsilon = \epsilon_{30} \left[N, \Phi_{\text{shear}}(m), \Phi_{\text{strain}}(m)\right] L(\theta, N), \]  
\[ \epsilon_{30} = 6.73 \times 10^{-10} \left(\frac{N}{N_0}\right)^2 \left(\frac{0.1}{k_c}\right)^2 \left(\frac{1 + 1/R}{4/3}\right) \left(\frac{2}{R - 1}\right)^{1/2} \]  
(b). Data processing
The shear profile was first derived from the raw LADCP velocity profile, with a 10-m-depth grid. Then, each profile was divided into 320-m segments with 160-m overlapping lengths, starting from the bottom. Each segment was centered at equally spaced depths and was normalized by the segment-averaged \( N \). A linear fit was removed from each segment and then windowed at both vertical wavenumbers less than \( k_c \), the ratio of the average spectral levels is \( \Phi_{\text{shear}}/\Phi_{\text{GM shear}} = 0.1/k_c \). Therefore, the expression of Eq. (1) becomes
\[ \epsilon = \epsilon_0 \frac{(V_{\text{shear}}^2)^{1/2}}{(V_{\text{GM shear}}^2)^{1/2}} h(R)L(\theta, N), \]  
where \( \epsilon_0 = 2.5 \times 10^{-5} \text{ W kg}^{-1} \). Following the formulation by Osborn (1980), diapycnal diffusivity is given by \( K_g = \gamma (\epsilon/N^2) \), where \( \gamma \) is the mixing efficiency, taken to be 0.2 here.

FIG. 2. Relation of potential temperature vs salinity (with \( \sigma_p \) contours overlain) along section I.
ends with 10% $\sin^2$ tapers for it. Subsequently, the shear in each segment was Fourier transformed (32 points), and its vertical wavenumber power spectral density $S(V_z/N)$ was computed. Because of the smoothing associated with range averaging, finite differencing, interpolation, and instrument tilting, one spectral correction function

$$S_{\text{correc}} = \frac{1}{\sin^2 \left( \frac{8}{\lambda_z} \right) \sin^2 \left( \frac{10}{\lambda_z} \right)}$$

was used for high-wavenumber attenuation (Polzin et al. 2002), where 8 m represents the vertical bin size for LADCP, and 10 m the vertical depth interval for the shear profile. The $N$-normalized shear variance $\langle V_z^2/N^2 \rangle$ was calculated as the corrected power spectral density integrated between a maximum vertical wavelength of 320 m and a variable minimum vertical wavelength that ranged from 50 to 90 m to avoid contamination by instrument noise at higher wavenumbers, $N$ was divided into 320-m segment, identical to the shear segment following Polzin et al. (1995). The strain profile was divided into 320-m segment, identical to the shear profile, and each segment was windowed at both ends with 10% $\sin^2$ tapers. Then, strain spectra $S(\xi_z)(k_z)$ were calculated through Fourier transforming each strain segment. A spectral correction $S_{\text{correc}} = \sin^2(2/\lambda_z)$ was required for the first differencing inherent in the gradients, where 2 m represents the vertical depth interval for the strain profile. The strain variance $\langle \xi_z^2 \rangle$ was obtained by integrating strain spectra over the same wavenumber band as shear variance,$$\langle \xi_z^2 \rangle = \int_{1/320}^{\max k} S(\xi_z)(k_z) \, dk_z.$$The GM model strain variances were computed over the same wavenumber band

$$GM(\xi_z^2) = \frac{\pi E_0 b j_s}{2} \int_{1/320}^{\max k} \frac{k_z^2}{(k_z + k_{z*})^2} \, dk_z.$$Diapycnal diffusivity values were obtained when applying the shear and strain spectra to Eq. (1). At each repeat-occupation station, the final diffusivity was derived by averaging the results of each profile. The averaged spectra of shear and strain below 1000-m depth in the SCS are shown in Fig. 3, along with the corresponding GM spectra. It is obvious that the observed shear and strain spectra at these depth bins have levels above GM spectra. Over these depth intervals the spectra elevation is large, which indicates that in terms of shear and strain, the SCS below 1000-m depth is typically not in an open-ocean state.

4. Dissipation and mixing

The spatial variability of turbulent mixing derived from the parameterizations in the study region is illustrated by Fig. 4. The result reveals a mixing pattern that appears to be correlated to the underlying bathymetry, and shows an increase of diffusivity with decreasing height above the bottom. Turbulent diffusivities on the order of $10^{-3} \text{ m}^2 \text{s}^{-1}$ are observed over rough topography, within a maximum thickness of about 2000 m above the bottom in the middle area of the SCS and the Luzon Strait. Enhanced mixing in such a thick layer is striking. All previous findings show that diffusivity values ranged between $10^{-4} \text{ m}^2 \text{s}^{-1}$ and $10^{-2} \text{ m}^2 \text{s}^{-1}$ at a depth of 500 m above the sea bottom in the Brazil Basin (Polzin et al. 1997), Scotia Sea (Naveira Garabato et al. 2004), and Romanche Fracture Zone (Polzin et al. 1996). Diffusivities are elevated by about two orders of magnitude in the SCS, where $K_p$ increases from about $10^{-4} \text{ m}^2 \text{s}^{-1}$ at 1000 m to $10^{-3} \text{ m}^2 \text{s}^{-1}$ to $10^{-2} \text{ m}^2 \text{s}^{-1}$ near the seafloor. It is clear that enhanced mixing is dominant in the SCS and the Luzon Strait. A slight decrease in the diffusivity exists between the SCS basin and the Pacific,
which is considered as a transition region. Here the microstructure intensity is little lower than its counterpart in the SCS and the Luzon Strait, although high values exceeding \(10^{-3}\) m\(^2\) s\(^{-1}\) actually occur near the flanks of the subsurface ridges. In contrast, the observations indicate that there is a background diapycnal diffusivity of only \(10^{-5}\) m\(^2\) s\(^{-1}\) over most of the smooth abyssal plain in the Pacific, although values of \(10^{-4}\) m\(^2\) s\(^{-1}\) are found in some patches below 1000-m depth. Peak values of \(10^{-3}\) m\(^2\) s\(^{-1}\) cannot be ignored near steep slopes and abrupt ridges in the SCS and the Luzon Strait, though they cannot be shown in Fig. 4. In addition, some high values of \(10^{-3}\) m\(^2\) s\(^{-1}\) occur in the uppermost 500 m of the continental shelf of the SCS and the transition region.

The variations of turbulent diffusivity, dissipation rate, and buoyancy frequency, with depth in the four regions denoted respectively as A, B, C, and D in Fig. 4, are illustrated in Fig. 5, and also are summarized in Table 1. Buoyancy frequency exhibits the same trend in the four regions, that is, ranging between \(10^{-4}\) s\(^{-1}\) in the upper layer and about \(10^{-6}\) s\(^{-1}\) in the deep layer. A small difference is observed in the Pacific, for which the buoyancy frequency is a little greater than those of the rest in the deep layer. The heightened levels of turbulence with averaged diffusivities of about \(10^{-3}\) m\(^2\) s\(^{-1}\) or greater occur throughout most of the water columns, accompanied by elevated values of the dissipation rate from about \(10^{-8}\) W kg\(^{-1}\) to \(10^{-7}\) W kg\(^{-1}\) in the SCS and the Luzon Strait (Figs. 5a,b,d). The mean vertically integrated dissipation rates \(E = \int_{-H}^{0} \rho c_v dz\) are found to range from \(3.1 \times 10^{-2}\) W m\(^{-2}\) to \(9.21 \times 10^{-2}\) W m\(^{-2}\) for these regions. There is a remarkable drop in the
magnitude of the diffusivity over the region of the smooth seafloor in the Pacific (Fig. 5c), with an averaged diffusivity of about $10^{-5}$ m$^2$ s$^{-1}$ and dissipation rate of about $10^{-9}$ W kg$^{-1}$. The drop in diffusivity is accompanied by a corresponding drop of $E$ to $3.6 \times 10^{-3}$ W m$^{-2}$. Some lower-diffusivity values occur in the upper layer in the SCS and Pacific (Figs. 5a,c), and this is caused more by a decrease of the $\epsilon$ value rather than an increase of the value of $N^2$. The situation is different from the case in the Scotia Sea, where reduction in $K_p$ was mainly related to the occurrence of high stratification (Naveira Garabato et al. 2004). Dissipation on the order of a $10^{-8}$ W kg$^{-1}$ range was found over the northeast Pacific and northeast Atlantic Oceans (Toole et al. 1994), Brazil Basin (Polzin...
et al. 1997), Romanche Fracture Zone (Polzin et al. 1996), and Scotia Sea (Naveira Garabato et al. 2004) in a layer thickness of about 500 m. Contrary to this, in the study regions A, B, and D, high dissipation values with an order greater than $10^{-8}$ W kg$^{-1}$ occupy most of the water column, which is the thickest ever reported as far as we known.

### 5. Internal tide energy flux

To further reveal the energy source for the enhanced mixing in the Luzon Strait and revealing internal tide energy flux generated in the Luzon Strait was conducted along section II (Fig. 1). The Luzon strait is an exclusive, deep passage connecting the SCS and the Pacific. East of the strait, the ocean floor is relatively flat, whereas bathymetry near the Luzon Strait is more variable. Thus, the Luzon Strait presents itself as a significant conversion machine, transforming barotropic energy to baroclinic energy. A total of 65 full-depth profiles of horizontal velocity $(u, v)$, temperature $T$, salinity $S$, and pressure $P$ were collected at 15 stations (Fig. 1) using LADCP and CTD. Five of the 15 stations were reoccupied in order to more effectively investigate internal tide energy flux. At each of these five stations, a total of 11 LADCP/CTD profiles were collected at 3-h intervals between successive profiles.

For internal gravity waves, baroclinic energy flux is the product of the pressure perturbation $\bar{p}$ and baroclinic perturbation velocity $v = (\bar{u}, \bar{v}, \bar{w})$, that is, $\mathbf{F}_E = C_g E = \langle \bar{p} v \rangle_\sigma$, where $C_g = \partial \omega / \partial k$ is the group velocity, $E$ is the wave energy, and $\langle \cdot \rangle_\sigma$ denotes an average over one wave phase. The baroclinic velocity profiles $\bar{u}'(z, t)$ were inferred by integrating the LADCP shear profiles vertically and seeking an integral constant. The semi-diurnal and diurnal signals $[\bar{u}^M, (z, t), \bar{u}^K_i(z, t)]$ of baroclinic perturbation velocity can be derived from $\bar{u}'(z, t)$ as

$$\bar{u}'(z, t) = u_0(z) + \bar{u}^M(z, t) + \bar{u}^K(z, t),$$

$$= u_0(z) + \sum_{i=1, 2} A_i \cos(\omega_i t - \varphi_i),$$

where $i = 1, 2, 3, \ldots, 11$ indicate the successive profiles obtained at different times at each repeat-occupation station, and the symbols $A_i, \omega_i, \varphi_i, (j = 1, 2)$ represent respectively the amplitude, frequency, and initial phase of the baroclinic semi-diurnal and diurnal perturbation velocity. The above coefficients were determined with a least squares method, which then yielded solutions for the perturbation velocity. The pressure perturbation $\bar{p}$ was computed by integrating the hydrostatic balance $0 = - (\partial \bar{p} / \partial z) - \nabla^2 \xi$ to obtain

$$\bar{p}(z) = -\rho_0 \int_z^0 \bar{b}(z') dz' + \bar{p}(0),$$

$$= -\rho_0 \int_z^0 \bar{b}(z') dz' + \frac{\rho_0}{H} \int_{-H}^0 \bar{b}(z') dz' dz,$n

$$= \rho_0 \int_z^0 \nabla^2 (\xi^2)(z') dz' - \frac{\rho_0}{H} \int_{-H}^0 \nabla^2 (\xi^2)(z') dz' dz,$$

where $\rho_0$ is the density averaged over the entire water column.

### Table 1. Averaged diapycnal diffusivity (m$^2$ s$^{-1}$) at different water depth ranges in the four regions referred to in Fig. 4.

<table>
<thead>
<tr>
<th>Region</th>
<th>Averaged Diapycnal Diffusivity (m$^2$ s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Region A</td>
<td>$4.9 \times 10^{-4}$</td>
</tr>
<tr>
<td>Region B</td>
<td>$1.0 \times 10^{-3}$</td>
</tr>
<tr>
<td>Region C</td>
<td>$1.0 \times 10^{-5}$</td>
</tr>
<tr>
<td>Region D</td>
<td>$8.2 \times 10^{-4}$</td>
</tr>
</tbody>
</table>

**Fig. 6.** Zonal baroclinic energy flux of (left) diurnal internal tide and (right) semi-diurnal internal tide.
displacement from baroclinic measurements. As a result, 11 instantaneous $p_u$ profiles were available at each repeat occupation station for semidiurnal and diurnal internal tides, respectively. We then averaged five continual profiles for estimating the final energy flux profile of semidiurnal internal tide, and nine profiles for the diurnal internal tide. Final energy fluxes of both semidiurnal and diurnal internal tides are shown in Fig. 6. In all profiles, there is a low-mode character to the energy flux, with much of the energy confined in two depth intervals, that is, one in the upper portion and the other in the lower portion of the water column. “Pulses” of the zonal energy flux are most prominent in the depth range between the surface and the depth of 500 m, with each pulse occupying an about 500-m vertical interval. Much weaker activity occurs below the 500-m depth. The energy fluxes of the diurnal and semidiurnal internal tide are clearly pointed toward the west at most of the stations, indicating that the energy radiation is mostly directed toward the SCS. The levels of energy flux in these profiles range typically from 10 W m$^{-2}$ in the upper 300 m to 1 W m$^{-2}$ in the lower depth. The depth-integrated westward energy fluxes range from 2 to 10 kW m$^{-1}$, with an average value of 7 kW m$^{-1}$ toward the SCS. With a width integral, the westward energy fluxes of diurnal and semidiurnal internal tide amount to 10 GW in total, of which 4.6 GW is contributed by diurnal internal tide and 5.4 GW by the semidiurnal internal tide.

6. Results comparison

a. Compare results with available methods

Based on the internal tide energy calculated in section 5, a basin-averaged abyssal $K_\rho$ can be calculated from the energy budget. We assume 80% of the internal tide energy is available for supporting the mixing in the SCS, termed $F_{Ea}$. As a result, a basin-averaged $K_\rho$ of $1.3 \times 10^{-5}$ m$^2$ s$^{-1}$ can be deduced using a simple model

$$K_\rho = 0.2 \frac{F_{Ea}}{\rho V N^2},$$

where $\rho$ is the averaged in situ density of the SCS water, $V$ is the deep-water volume of the SCS below 1500 m, and $N^2$ is the squared average buoyancy frequency below 1500 m. Values of these variables are listed in Table 2.

On the other hand, from a heat budget based on estimates of the abyssal water transport through the Luzon Strait, a basin-averaged $K_\rho$ can be obtained. The deep Pacific water, which is colder than 2.6°C and denser than 1027.58 kg m$^{-3}$, enters the SCS and becomes trapped in the deep layer at a depth that is too deep for this water to flow out of the SCS through the Taiwan Strait, the Karimata Strait, or the Mindoro Strait. A temperature steady state in the deep ocean requires a balance between the upward advection of cold water and a downward diffusive heat flux (Munk and Wunsch 1998). A heat budget balance leads to the equation

$$\rho QC_p \left( \theta_u - \theta_i \right) = -A \rho \frac{\partial \theta}{\partial z},$$

where $Q$ is a volume transport of abyssal water into the SCS through the deep layer of the Luzon Strait (Tian et al. 2006); $C_p$ is the specific heat capacity of seawater; $\theta_i$ and $\theta_u$ are, respectively, the potential temperature of the

![Fig. 7. Turbulent dissipation rate profiles measured by VMP-2000 at 21°N, 118°E. The original observation (gray line) and the depth-averaged result (black line) are indicated. Note the logarithmic scale for the $x$ axis.](image-url)
inflowing Pacific deep water and the water upwelling to the SCS intermediate layer by mixing and then flowing back to the Pacific; $A$ is the area of the SCS; and $\partial \theta / \partial z$ is the mean vertical gradient of potential temperature. Values of these variables are listed in Table 3. A basin-averaged $K$ is then estimated to be $1.8 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$.

b. Compare results with microstructure observations

Diffusivity profiles inferred from the parameterization were compared with the concurrent microstructure measurements by VMP-2000 in the SCS. The VMP-2000 sampled turbulent shear at 512 Hz as it descended at a speed of roughly 0.6 m s$^{-1}$, achieving a vertical full-wavelength resolution of 3 mm. Electronic noise, instrument vibration, and small amounts of turbulence generated at the leading edge of the guard assembly applied a shear dissipation rate noise level of $1 \times 10^{-10} \text{ W kg}^{-1}$ to our measurements. Four VMP-2000 profiles extending to 1500-m depth in the SCS are used here. Figures 7 and 8 show the instantaneous dissipation rate profile and the corresponding spectra at different depth intervals at 21°N, 118°E. The final averaged results for the four diapycnal diffusivity profiles are shown in Fig. 9. The two result sets show a similar overall trend. Although all are within the same order of magnitude, there are quantitative differences by factors of 2–5 between the individual profiles. The comparison shows a satisfactory agreement between the observation and parameterization, with biases of $2 \times 10^{-3}$ m$^2$ s$^{-1}$, $1 \times 10^{-3}$ m$^2$ s$^{-1}$, $5 \times 10^{-4}$ m$^2$ s$^{-1}$, and $3 \times 10^{-4}$ m$^2$ s$^{-1}$ for the four pairs of diffusivity profiles, respectively. The bias $B$ involved here between vectors $A_1$ and $A_2$ is determined as $B = [(1/n) \sum_{i=1}^{n} (A_{1i} - A_{2i})^2]^{1/2}$, where $n$ is the total number of elements in vectors $A_1$ and $A_2$. The fact that all of the three independent calculations are able to yield closely similar results adds to our confidence in the conclusions reached in this study.

7. Summary and discussion

An analysis of current velocity and hydrographic observation in IWME has revealed that the heightened...
level of turbulent mixing appears to exist below 1000-m depth in the SCS. The mixing reaches an averaged order of $10^{-3}$ m$^2$ s$^{-1}$ in magnitude, although some weaker values are evident in the upper depth. The Pacific is occupied with less vigorous mixing. As in the Luzon Strait, the turbulent mixing there is strong as a whole, especially in the middle part below 500-m depth, where the maximum reaches the order value of $10^{-2}$ m$^2$ s$^{-1}$. The energy source, which induces the enhanced turbulent mixing existing in the SCS, is estimated to be 10 GW, based on calculating the internal tide energy flux generated in the Luzon Strait and then propagating toward the SCS. Assuming that 80% of this energy is dissipated in the SCS to furnish the energy need for turbulent mixing there, the averaged diapycnal diffusivity is inferred and found to be comparable with the parameterization result. Another independent test involving the heat budget balance in the SCS also gives results in agreement with those of the parameterization. It is notable as well that a comparison between the parameterization results and the observations made at the four stations in the SCS again shows a good agreement.

Our analysis suggests that the pattern and the intensity of the basin-scale abyssal circulation in the SCS are controlled by small-scale turbulent mixing, whose impact has yet to be recognized. Upwelling in the SCS driven by mixing is indispensable to maintaining the abyssal water transport through the Luzon Strait. Furthermore, the inflowing waters from the deep Pacific have been warmed up before leaving the basin, which implies that the SCS contributes to the water mass transformations in the Pacific. In summary, this paper illustrates the important role played by the enhanced mixing in the SCS on Pacific circulation. A generalized qualitative relationship between the two is put forward in this paper, but further observations and numerical modeling works are still required to put this relationship on a more quantitative basis.

**Acknowledgments.** The authors wish to thank Eric Kunze, Louis St. Laurent, and an anonymous reviewer.
for valuable suggestions and comments on the manuscript. This work is supported by the Natural Science Foundation of China (Project 40776005 and 40890153).

REFERENCES


