Variability of the Observed Deep Western Boundary Current in the South China Sea

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ABSTRACT: The existence of a deep western boundary current (DWBC) in the South China Sea (SCS) was verified by direct observations from three current moorings deployed from September 2015 to September 2018. The average current speeds observed in the DWBC were around 1 cm s$^{-1}$ along the northern boundary and 3 cm s$^{-1}$ along the western boundary. The DWBC demonstrates significant intraseasonal variability in the 30–120-day-period band, which may come from the variability in the Luzon overflow or the eddies in the deep SCS forced by a stable Luzon overflow. In addition, observations found that this DWBC along the northern boundary can reverse its direction meridionally in the spring. Model results suggest that if the Luzon overflow decreases one-third of its typical transport, this current reversal can occur. This behavior can be explained through “relaxation” theory.

KEYWORDS: Abyssal circulation; Intraseasonal variability

1. Introduction

The formation and distribution of deep water in the world’s oceans has been extensively studied (Cox 1989; Broecker 1991; Ganachaud 2003; Reid 2005). The deep western boundary current (DWBC), as the most important component of the deep ocean, has also been widely studied (Johns et al. 1993; Schott et al. 2006; Bacon and Saunders 2010; Toole et al. 2011; Peña-Molino et al. 2012; Meinen et al. 2013; Toole et al. 2017). However, there are few studies on the DWBC in the South China Sea (SCS). The SCS is the largest semi-enclosed deep sea in the northwestern Pacific. A significant quantity (2.5 Sv; 1 Sv = 10$^6$ m$^3$s$^{-1}$) of Northwest Pacific Deep Water enters the deep SCS through the Luzon Strait, which then drives the SCS deep circulation (Qu et al. 2006).

Based on an updated monthly climatology of observed temperature and salinity from the U.S Navy Generalized Digital Environmental Model (GDEM) version 3.0, Wang et al. (2011) calculated the geostrophic currents in the deep SCS assuming the velocity was zero at 2400 m. The most prominent feature is a basin-scale cyclonic circulation with an intensified western boundary current (WBC, Fig. 1a). This DWBC flows southwestward along the northern boundary of the SCS and southward along its western boundary, which is consistent with the classical theory of the DWBC from Stommel (1958) and Stommel and Arons (1960a,b).

With direct current observations consisting of six current meter moorings along a section southeast of the Zhongsha Islands, Zhou et al. (2017) suggested that the mean southward current with around 2.0 cm s$^{-1}$ speed was the DWBC in the SCS and the oscillation period of the current was around 90 days. However, the persistence of DWBC in the SCS has not been discussed. As a comparison, almost 100%, 80%, and 90% of time series show a persistent southward, southeastward, and southward DWBC in the east of bottom Cape Farewell (Bacon and Saunders 2010), east of Northeastern Brazil (Johns et al. 1993), and off Grand Bank (Schott et al. 2006), respectively. The details are also listed in Table 1. In addition, these deep current observations also showed a significant feature that there exists robust intraseasonal variability from 1 month to 120 days.

To study the characteristics of the DWBC in the SCS, two solitary moorings were first deployed in the northern and western boundary regions for 1 year between September 2015 and August 2016. The current direction in the western boundary was found not to be steady, but showed frequent variations. Therefore, subsequently another mooring was deployed closer to the western boundary wall to better capture the boundary current signal. Based on the previous knowledge, we designed our mooring observations to verify the deep boundary current and to see its persistence. Using data from three current moorings, the DWBC in the SCS can be seen to be composed of a southwestward current along the northern boundary and a southward current along the
western boundary. In addition, this study also attempts to describe the intraseasonal variability within this DWBC and possible mechanisms driving these variations.

2. Data and model

a. Data

To observe the DWBC in the SCS, three moorings (M1, M2, and M3) using recording current meters (RCMs) were deployed in the deep SCS (Fig. 1a) from September 2015 to September 2018 to monitor the horizontal velocity of the deep flow (Figs. 1b,c). The detailed experimental design information, such as location, sample interval, water depth, and deployment depth, for the moorings are listed in the Table 2. To filter out tidal currents, a 72-h low-pass filter (Godin 1966) was applied to all time series to remove the inertial and major tidal signals and all data were daily averaged. In addition, 56 profiles of the conductivity–temperature–depth systems (CTDs) deeper than 3000 m in the northern SCS collected from 2004 to 2018 were also used in this study. These CTDs were interpolated into 0.25° resolution grid data by the method of Data-Interpolating Variational Analysis (Beckers et al. 2014). The U.S Navy GDEM version 3.0 monthly climatology of observed temperature and salinity, which has a horizontal resolution of 0.25°×0.25°, was also used. It has 78 standard depths from surface to 6600 m, with a vertical resolution varying from 2 m at the surface to 100 m below 1600 m depth. The GDEM version 3.0 was derived from the temperature and salinity profiles extracted from the Mater Oceanographic Observational Dataset edited at the Naval Research Laboratory (NRL) (Fox et al. 2002). The NRL manually examined all the profiles within groups covering small geographic regions and short seasonal or monthly time periods to remove erroneous profiles. This dataset contains about 2.7 million usable profiles, and the data evaluation for the global ocean was reported by Carnes (2009). Wang et al. (2011) demonstrated that the GDEM captures a detailed salinity distribution in the deep SCS. The geostrophic velocity field was calculated from the GDEM v3.0 using the thermal wind relation assuming the velocity was zero at 2400 m (Wang et al. 2011).

b. Inverse reduced gravity model

The flow near the bottom layer in the deep basin was found to be faster than that above it (Huang 2010). Therefore, as an
approximation, it is assumed that the pressure gradient above this layer is negligible. So that, effectively, an inverse reduced gravity model has been applied to study the deep circulation dynamics (Stommel and Arons 1960a; Speer and McCartney 1992; Wang et al. 2018), as illustrated in Fig. 2. The momentum and continuity equations of the reduced gravity model can be described as

\[
\begin{align*}
\frac{\partial h}{\partial t} + hu \frac{\partial u}{\partial x} + hv \frac{\partial u}{\partial y} + u \frac{\partial h}{\partial x} + \frac{\partial hv}{\partial y} &= -g h \frac{\partial \eta}{\partial x} + \frac{\partial^2 h u}{\partial x^2} + \frac{\partial^2 h u}{\partial y^2} - \frac{R}{h_0} \frac{\partial h}{\partial x}, \\
\frac{\partial h}{\partial t} + hu \frac{\partial v}{\partial x} + hv \frac{\partial v}{\partial y} + v \frac{\partial h}{\partial x} + \frac{\partial hv}{\partial y} &= -g h \frac{\partial \eta}{\partial y} + \frac{\partial^2 h v}{\partial x^2} + \frac{\partial^2 h v}{\partial y^2} - \frac{R}{h_0} \frac{\partial h}{\partial y}, \\
\frac{\partial h}{\partial t} + \frac{\partial hu}{\partial x} + \frac{\partial hv}{\partial y} &= -w,
\end{align*}
\]

where \(x\) and \(y\) are the zonal and meridional coordinates, respectively; \(u\) and \(v\) are the horizontal velocity components in the \(x\) and \(y\) directions, respectively; \(f = 2\Omega \sin \theta\) is the Coriolis parameter; \(\Omega\) is the angular frequency of Earth’s rotation; \(\theta\) is latitude; \(g\)’ is the reduced gravity; \(R/h_0\) is the bottom friction coefficient; and \(a_h\) is the horizontal momentum viscosity, with \(R\) set to \(10^{-3}\) m s\(^{-1}\) and \(a_h\) set to \(300\) m\(^2\) s\(^{-1}\). The thickness of the water column \(h\) is defined as

\[
h = h_0 - h + \eta,
\]

where \(h_0 = 3000\) m is the vertical depth of the interface from deepest bottom of the SCS, \(h_0\) is the prescribed bathymetry, and \(\eta\) is the deviation of the interface.

The model domain includes the SCS central basin bounded by \(115^\circ\) and \(122^\circ\)E in longitude and \(12^\circ\)–\(22^\circ\)N in latitude with a horizontal resolution of \(1/6^\circ \times 1/6^\circ\). The deep basin circulation is forced by uniform upwelling and the deep overflow (1.5 Sv). The 1.5 Sv is chosen based on the observations from the Luzon Strait (Zhao et al. 2014). Parameter \(w\) represents the uniform upwelling and is set to \(2 \times 10^{-4}\) m s\(^{-1}\), so that the total upwelling over the area of model domain approximately equals to 1.5 Sv of total transport of the Luzon overflow, as needed to satisfy mass conservation. Several numerical experiments are carried out to investigate the role of overflow in the abyssal circulation dynamics. The simple model is run for 10 years to reach a quasi-steady state at first, then begin the experiments from 1 to 360 days. Luzon overflow is set to 1.5 Sv for a full year with real topography generally except specific statements. The simple model generally demonstrated a cyclonic circulation in the deep SCS (Fig. 3a), which is quite similar to the diagnosed circulation pattern based on GDEM (Wang et al. 2011) and the analysis based on Stommel–Arons theory (Stommel and Arons 1960a). The numerical experiments are as below:

Case 1: Control run;

Case 2: Luzon overflow satisfies the equation \(1.5\) Sv \(- 0.6\) Sv \(\times \sin[2\pi/(80\) days]\], where \(t\) is from 1 to 360 days for a full year, to confirm the hypothesis that the 30–120 day oscillation was from the Luzon overflow, 0.6 Sv is chosen based on the time series of the observation from (Zhou et al. 2014), 80 days is chosen because it is a significant oscillation period of observed velocities;

Case 3: Luzon overflow satisfies the equation

\[
\begin{align*}
\{ &1.5\) Sv \(- 0.6\) Sv \(\times \sin[\pi/(30\) days]\] \text{ for } 90 < t < 120 \text{ days} \\
&1.5\) Sv \text{ otherwise}
\end{align*}
\]

to explore how the 1-month decrease of Luzon overflow is linked to the current reversal in spring.

c. MIT General Circulation Model (MITgcm)

The MITgcm (Marshall et al. 1997) was also applied to simulate the deep SCS circulation. The model domain includes the SCS central basin bounded by \(99^\circ\)–\(140^\circ\)E in longitude and

\[
\nabla p = 0 \quad \vec{u} = 0 \quad \rho
\]

\[
\text{surface}
\]

\[
h
\]

\[
\vec{u} \neq 0 \quad \rho + \Delta p \quad h_0
\]

\[
\text{bottom}
\]

\[
\text{Fig. 2. Schematic of the inverse reduced gravity model configuration.}
\]

|$TABLE 2. Experimental mooring design.| Mooring | Water depth (m) | Mooring duration | Sample interval (min) | Location | Instrument depth (m) |
|---|---|---|---|---|---|
| M1 | 3530 | Sep 2015–Jun 2017 | 20 | 19.7°N, 119.5°E | 2000 |
| M2 | 4103 | Sep 2015–Aug 2016 | 20 | 16.5°N, 115.7°E | 2025 |
| M3 | 4149 | Jun 2017–Sep 2018 | 20 | 16.6°N, 115.6°E | 3226 |
5°S–30°N in latitude with a horizontal resolution of 1/12° × 1/12°. Thirty vertical levels have resolutions varying from 5 m at the surface to 200 m below 500 m depth. The model was forced by the surface flux from the European Centre for Medium-Range Weather Forecasts (ECWMF) ERA-40 (Uppala et al. 2005) for the 23-yr period of 1993–2015. The monthly mean climatological temperature and salinity data from GDEM were applied as the initial condition. The daily open boundary conditions were from Simple Ocean Data Assimilation (SODA) based on Parallel Ocean Program (Smith et al. 1992).

The model was run for 100 years to reach a steady state, and then run for another 23 years as case 4. The climatological mean result of case 4 was shown in Fig. 3b. It shows that the deep circulation in the SCS is also cyclonic and accompanied by a significant DWBC, suggesting that the MITgcm can be applied in the deep SCS circulation. The daily outputs of the final year in the model were used for analysis in this study.

3. Results

a. Verification of a persistent WBC in the deep SCS

Fifty-six CTD profiles deeper than 3000 m from 2004 to 2018 were applied to see the deep circulation in the SCS. As shown in Fig. 4, the spatial distribution of the salinity at 3000 m layer suggested that the deep circulation is cyclonic in SCS, accompanying a southwestward current along the northern boundary and a southward current along the western boundary. This new dataset suggests that the cyclonic circulation is robust in the deep SCS.

Because the basin widens after the overflow passes over the Luzon Strait, the DWBC in the northeastern SCS is quite weak (Wang et al. 2011). The M1 mooring was deployed at the inflection point where the northern continental slope changes from the meridional direction along 119.5°E to the zonal direction along 19.7°N (Figs. 1a,b). The time series of current from this mooring (Fig. 5a) demonstrates an obvious prevailing southwestward flow. Over the entire 660-day observational record, the meridional velocity (Vob) was southward for almost 482 and 513 days at 2000 and 2500 m depth, representing southward velocity on 73.0% and 77.7% of the time series, respectively. The time averaged Vob was \(-0.57 \pm 1.52\) and \(-1.06 \pm 1.59\) cm s\(^{-1}\) and the range fluctuated between \(-4.35\) and \(+3.53\) cm s\(^{-1}\) and between \(-4.42\) and \(+4.42\) cm s\(^{-1}\) at 2000 and 2500 m depth, respectively. Similarly, the zonal velocity (Uob) was westward for almost 411 and 485 days at 2000 and 2500 m depth, representing westward velocity on 62.3% and 73.5% of the time series, respectively. The time averaged Uob was \(-0.40 \pm 1.83\) and \(-0.87 \pm 1.99\) cm s\(^{-1}\) at 2000 and 2500 m depth, respectively.
and the range fluctuated between \(-4.12\) and \(+6.27\) cm s\(^{-1}\) and between \(-5.20\) and \(+6.15\) cm s\(^{-1}\) at 2000 and 2500 m depth, respectively. These observations verify the existence of a deep boundary current near the northern boundary and demonstrate that the dominant direction of the DWBC along the northern boundary at 2000 and 2500 m depth was southwestward.

During the same observed period from September 2015 to August 2016 as M1, another mooring (M2) was deployed at 16.5°N, 115.7°E around 32 km from the western boundary wall to verify that a southward current exists along the deep western boundary (Figs. 1a,c). However, this mooring failed to detect a stable southward current, finding southward flow on less than 40% (129, 132, and 135 days of 345 days at 2025, 3025, and 3525 m layers, respectively) of the time series and northward flow at all other times (Fig. 5b). As an attempt to observe the southward WBC, this mooring was then moved closer to the western boundary and placed at 16.6°N, 115.6°E (relabeled M3 for clarity), around 20 km from the western boundary wall during the period from June 2017 to September 2018. The time series of deep current at M3 are shown in Fig. 6, the mean meridional velocity statistics at M3 are \(-0.94 \pm 2.43\), \(-2.69 \pm 3.43\), and \(-3.36 \pm 3.40\) cm s\(^{-1}\) at 2226, 3226, and 3726 m, respectively. As expected, over the entire 453 days of the observational record, the meridional velocity was southward at the three current meter depths for almost 286, 359, and 377 days. So that southward flow was seen on 63.1%, 79.2%, and 83.2% of the days of the data record for the upper, middle, and bottom layers at mooring M3, respectively. The above observations verify that there is a stable southward current in the western boundary, the differences between M2 and M3 suggest that the WBC is very narrow and closer to the wall, thus, the width of the DWBC is roughly around 20–32 km from the boundary wall.

Therefore, the existence of a WBC in the deep SCS is supported by the velocity observations at M1 and M3. The average speed of this DWBC is around 1 cm s\(^{-1}\) along the northern boundary and 3 cm s\(^{-1}\) along the western boundary. The robust skewed frequency distribution of current at M1 and M3 shows more details about the stable DWBC (Fig. 7). Regardless of meridional velocity or zonal velocity, it shows a positive skewed distribution with a median of \(-2\) to \(-1\) cm s\(^{-1}\) at M1, and a negative skewed distribution with a median of \(-2\) to \(0\) cm s\(^{-1}\) at M3. In addition, the meridional velocities at M1 and M3 for all vertical layers are almost in the same direction consistently suggests that the DWBC is relatively barotropic and can exist over a rather thick layer (Figs. 5 and 6). Note that the DWBC also has a baroclinic component because the observed current speed changes with depth.

b. Intraseasonal variability of the DWBC

The observations at both the M1 and M3 moorings also demonstrate robust intraseasonal variability of the DWBC. Wavelet spectral analysis demonstrates quantitatively that the time series of the current speed at both M1 and M3 have a significant intraseasonal signal in roughly the 30–120 day period band (Fig. 8). It is interesting to note that both the southwestward current in northern boundary and the southward current in western boundary demonstrate intraseasonal oscillation of 30–120 days, giving us a hint that a linkage between them may exist. Furthermore, since the deep water in

![Fig. 5. Horizontal velocity time series at (a) M1 and (b) M2 (unit: cm s\(^{-1}\)).](image)

![Fig. 6. Horizontal velocity time series at M3 (unit: cm s\(^{-1}\)).](image)

![Fig. 7. Histograms of velocity components for each 1 cm s\(^{-1}\) bin at (a) M1 (2500 m depth) and (b) M3 (3226 m depth). Red bars represent zonal velocity, and blue bars represent meridional velocity.](image)
the SCS has its origin from the Luzon overflow, we hypothesized that the 30–120 day oscillation was from the Luzon overflow.

To test this hypothesis, we impose a Luzon overflow with an 80 day oscillation component in the reverse reduced gravity model (case 2). The time series along the boundary for this case is presented in Fig. 9a. As we can see, the 80 day overflow oscillation propagates into the SCS, suggesting that a source-driven mechanism for the 30–120 day oscillation in the deep SCS is a possibility. In addition, the amplitude of the 80 day signal in the deviation of the interface (\(\eta\)) along the boundary region (Fig. 10) is much larger than in the interior SCS, indicating that the intraseasonal signal from the Luzon overflow mainly propagate into the SCS along the boundary.

In addition, recent study suggested that the intraseasonal variations of deep SCS circulation may be caused by surface eddy (Zhang et al. 2013). We compared the current meter data with satellite altimeters data during the observed period. Unfortunately, the correlation between them is quite weak (correlation coefficient less than 0.15), suggesting that the intraseasonal variations found here are not likely associated with the upper-layer eddy (figures not shown). The results further remind us to see whether there are some deep layer eddies to drive the intraseasonal variability of the DWBC.

Although the Luzon overflow is stable, the MITgcm experimental results suggest that there are many eddies in the deep SCS (Fig. 11a). These eddies are basically generated near the inlet of deep Luzon overflow in the eastern boundary, then move westward, and finally dissipate in the western boundary, with a life cycle about 80 days from generation to dissipation. Further, the velocity anomalies show obvious movement from east to west (Fig. 11b) and demonstrate significant intraseasonal variations with a period of 30–120 days. It is notable that the Luzon overflow is stable in this experiment, but its intraseasonal variability is robust, suggesting that the intraseasonal variation in the deep SCS can also be caused by the eddy in the deep SCS itself.

Summing up the above description, we provide two possible explanations of the intraseasonal variability of deep SCS. First, once the Luzon overflow has intraseasonal variability as observed, it can induce the intraseasonal variability in the deep SCS as expected. Second, as the Luzon overflow is stable, it can also produce an intraseasonal variability by driving out deep eddies. Actually, all these mechanisms are still debating. The intraseasonal variability of deep ocean could also be related to the atmospheric Madden–Julian oscillation (Matthews et al. 2007), the monsoon and El Niño–Southern Oscillation (Wang et al. 2012), ocean instabilities (Srinivasan et al. 2009), or barotropic Rossby wave (Warren et al. 2002). This unsolved question is left for future study based on more observations and three-dimensional high-resolution models.

c. Current reversal of the northern DWBC in spring

Although the dominant direction of the DWBC along the northern boundary is southwestward, it is interesting to note that the meridional velocity at mooring M1, reverses its direction from southward to northward in April or May each year, and the current speed reaches its positive maximum of \(~4.4\) cm s\(^{-1}\) (Fig. 5a) at that time. To investigate possible mechanisms for this reversal, looking for some possible signals shows that the Luzon overflow appears to attain its seasonal minimum in the March to April time frame (Zhou et al. 2014).
Is it possible that the decrease of Luzon overflow is linked to the current reversal in spring? To explore this question, a numerical experiment (case 3) is designed to consider variations of the Luzon overflow (from Zhou et al. 2014).

In the initial phase of case 3, the deep Luzon overflow decreases from 1.5 to 1.0 Sv between days 90 and 105 (Fig. 12a, blue line). As a result, the deep southward velocity at M1 increased from 2 to 4 cm s$^{-1}$ accordingly but with a 5 day lag (Fig. 12a, green line and Figs. 13a–c). Subsequently, in the second phase of case 3, as the Luzon transport increases from 1.0 Sv back up to 1.5 Sv (from day 105 to day 120), the southward meridional velocity first responds by weakening (Fig. 13d) and then reversing direction to northward about 15 days (from day 120 to day 135) later (Figs. 13e,f), noting that the 15 days is roughly equal to the time that the current would take to flow from the Luzon Strait trough to the M1 mooring site in deep SCS (Fig. 9). Finally, as the Luzon overflow remains fixed, the deep current turns back to southward (Fig. 13g). These results generally simulate the flow reversal seen in the observations, suggesting that the variations in the of Luzon overflow, specifically, its weakening, are important for the deep boundary current reversal.

To give a more robust conclusion, a total of 195 experiments were designed to test how variations in the strength of the mean overflow affect the deep current reversal. The Luzon overflow ($Q_{LS}$) in these experiments was set to

$$Q_{LS} = S - A \sin(\pi t/30).$$

The mean Luzon overflow ($S$) was varied from 0.1 to 1.5 Sv at 0.1 Sv increments and the decrease in its amplitude ($A$) was varied from 0 to 0.6 Sv at 0.05 Sv increments. With the mean $S$ fixed as a constant, say 1.2 Sv, the meridional flow at M1 would reverse to northward only when $A$ exceeded 0.4 Sv. As $A$ continued to increase pass 0.4 Sv, the maximum meridional velocity also increase. On the other hand, the larger the Luzon overflow is, the greater the amplitude that is required to reverse the direction of the meridional flow (Fig. 13i). The quantitative relationship between $S$ and $A$ that induces a current reversal roughly satisfies $A = S/3$.

This may be explained by a theory analogous to “wind relaxation” (Chao et al. 1995). At the beginning, the Luzon Strait overflow possesses a fixed transport. However, as the overflow flows southwestward and tend to “pile up” water against the deep basin wall. When the Luzon overflow decreases to a certain degree, it triggers a northeast current along the wall boundary. Because of the westward shift induced by beta ($\beta$) effect, the current eventually flows to the northwest. We can call this mechanism “overflow relaxation.”

In addition, the meridional velocity will only reverse its direction to northward at a small area near M1 (blue half-ellipse area in Fig. 13a). Outside of the half-ellipse area, the decrease
in meridional velocity lags by several days of the decrease in the Luzon overflow, and then increases as the Luzon overflow increases. Take sites T0 (21.1°N, 118.8°E, between M1 and M3) and M3 as examples, the meridional velocity at T0 (M3) decreases as the Luzon overflow decreases but with a 9 (15) day lag, and further increases as the overflow increases (Figs. 12b,c). These results suggest that the reversal of current direction in the spring is very localized.

4. Summary and discussions
The results from three current moorings demonstrate that there exists a narrow, relatively barotropic western boundary current in the deep SCS, a dominant southwestward current and a significant southward current were found along the north boundary and the west boundary, respectively. The robust
stable DWBC found here is quite similar to world oceanic DWBCs (Johns et al. 1993; Schott et al. 2006; Bacon and Saunders 2010), this finding may be helpful to understand deep circulation processes of the marginal seas.

Because the flow at the M2 and M3 demonstrate different current pattern, we surmise that the width of the DWBC in that area is between 20 and 32 km. By simple estimation from the current moorings, the transport of the DWBC is about 1.2 to 1.92 Sv (assuming the thickness of the DWBC is 2000 m and the mean velocity is 3 cm s\(^{-1}\) from M3), which is consistent with the 1.65 ± 0.48 Sv observed by Zhou et al. (2017). According to Liu and Gan (2017), the deep SCS has a residence time of ~40 years, which is much shorter than the renewal time of the deep Pacific water. Thus, we estimate that the deep SCS could transport 1.56 × 10\(^{18}\) to 2.49 × 10\(^{18}\) kg \(M = Q_{\text{DWBC}} \times T_{\text{renew}} \times \rho\), where \(M\) is water mass, \(Q_{\text{DWBC}}\) is the transport of the DWBC, \(T_{\text{renew}}\) is the renewal time of 40 years, and \(\rho\) is water density of 1030 kg m\(^{-3}\)) water from the deep Pacific every 40 years. This is a considerable quantity of deep water and provides a crucial clue explaining the role of the SCS in the renewal of the deep water masses. The unique deep features of the SCS circulation found in this study may be important for understanding deep variations in the world’s oceans. Note this transport quantity calculated here may be overestimated because the DWBC may entrain some regional deep water, as indicated from the differences between the SCS and the Pacific Ocean water masses.

A power spectral analysis of the western boundary current time series finds significant intraseasonal variability in the 30–120 day band. The intraseasonal variability of the deep Luzon Strait transport can propagate into the interior of the SCS. Our study indicates two possible mechanisms to drive the intraseasonal variability: One is the propagation of intraseasonal variability in the Luzon overflow and the other is the deep layer eddies in the SCS driven by a stable Luzon overflow. But in fact, even though M2 does not locate at the path of the DWBC, the current at M2 also presents significant intraseasonal variability (Fig. 4b). This intraseasonal variation in the deep SCS may be associated with other factors such as surface mesoscale ocean eddies, topographic Rossby waves, and so on (Rudnick...

In addition, the southward current is also found to exist along the northern boundary nearly all year, although it reverses direction in April or May. Our model results suggest that the reversal of the southward current can be attributed to variations in the mean Luzon overflow. As the overflow decreases to two-thirds of its original value in spring, the deep boundary current reverses after roughly 15 days. A “relaxation” theory is put forth that may explain this reversal. Because we have only one single mooring observation for 2 years along the northern boundary, the feature of current reversal and its associated theory should be further examined in the future.

In general, the DWBC in the SCS is a complicated system with large variability associated with not only Luzon overflow and topography, but also eddies and internal ocean dynamics (Tian and Qu 2012). For example, Wang et al. (2011) suggested that the deep SCS circulation was mainly forced by the Luzon overflow, and controlled by salinity variations associated with the abyssal topography. Zhang et al. (2013) showed that an upper-layer eddy pair may greatly influence the deep circulation oscillation in the northeastern SCS. Shu et al. (2014, 2016) indicated that the eddies can also drive bottom-trapped topography Rossby waves on the slope. Thus, more comprehensive diagnosis and analysis of the deep SCS circulation needs both the development of ocean general circulation model simulation skill and the continuation of the in situ observations.

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REFERENCES


