

Barotropic dominance of intraseasonal to seasonal variability in the western boundary currents east of Luzon revealed by mooring observations*

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Abstract Recent studies have shown that east of Luzon both the Kuroshio in upper layer and the Luzon Undercurrent (LUC) in the intermediate and deep layers have significant intraseasonal variabilities associated with eddy activities. However, the integrated behavior of the western boundary currents in the full-depth ranges, the dominant variability mode of the western boundary currents east of Luzon and associated physical processes or underlying dynamics are still unknown. This study investigates the questions by using more than two-year mooring observations at 18°N/122.7°E. The results reveal that the dominant variability of the western boundary currents on intraseasonal to seasonal timescales is primarily barotropic in nature. The barotropic velocity component serves as an effective proxy for this variability and is closely linked to variations in the LUC as well as to temperature changes in the intermediate and deep ocean, typically below 1 000 m. During the observation period, the temporal evolution of the western boundary current is characterized by three abnormal events featuring an intensified LUC, which induce southward barotropic transport at 18°N. The third abnormal event with a temperature drop in observed depth range lasted about 6 months, mainly related to the wind stress curl on the basin scale, while the first two abnormal events have a much shorter duration and are likely related to local subsurface mesoscale or submesoscale eddy activity. Both large-scale wind fields and mesoscale processes may lead to a decrease in ocean temperature in the intermediate and deep layers, resulting in a strengthened LUC, dominating the variability of the western boundary current within the entire water column. This study highlights the important role of intermediate and deep ocean changes caused by large-scale wind fields or mesoscale processes in modulating barotropic velocity and the overall variability of western boundary currents.

Keyword: deep ocean; mesoscale eddy; variability; western boundary current; Luzon Island; barotropic

1 INTRODUCTION

Western boundary currents characterized by high velocities, such as Gulf Stream, Kuroshio, East Australian, Brazil, and Somali Current, play crucial roles in the meridional transports of mass and heat globally. The two western boundary currents in the western tropical Pacific, the Kuroshio east of Luzon and the Mindanao Current east of Mindanao in the upper layer, are maintained by the North Equatorial

Current (Nitani, 1972). In the subsurface and intermediate layers, the two western boundary currents, the Mindanao Undercurrent east of Mindanao (Hu and Cui, 1991) and a southward Luzon Undercurrent (LUC, Qu et al., 1997) east of Luzon, are generally thought to feed the eastward

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North Equatorial Undercurrent (Qiu et al., 2013). These western boundary currents east of Philippines regulate the heat budget of the warm pool, and also contribute to global thermohaline circulation by conveying water masses originating from the northern and southern hemispheres (Hu et al., 2011).

Over the past decade, direct current measurements by moorings in the western tropical Pacific have significantly advanced our understanding of the variabilities of the two western boundary currents east of Luzon, the Kuroshio and the LUC. Significant intraseasonal variabilities in the LUC have been found and attributed to mesoscale eddy activities or local atmospheric intraseasonal signals (Hu et al., 2013). Near Luzon Island, based on an approximately 8-month direct current measurement at 122.7°E, 18°N, Hu et al. (2013) first reported a significant 70–80-d variation in the LUC and it was related to the west-moving surface eddies. By analyzing the same mooring observations combined with a high-resolution model output, Wang et al. (2014) argued that the intraseasonal variability in the Kuroshio and LUC were related to ocean mesoscale eddies mainly originating from the North Equatorial Current. Chen et al. (2015) revealed that the Kuroshio at 18°N had a significant strengthening trend from November 2010 to October 2012. Ma et al. (2022) used three mooring observations at 18°N and thought the variability and reversal of the LUC are mainly a result of the first two baroclinic modes. Lien et al. (2014) and Rudnick et al. (2011) highlighted that the important role of mesoscale eddies in modulating the Kuroshio transport near its origin.

However, most existing studies have focused on the variabilities of one of the Kuroshio or the LUC. The integrated behavior of the western boundary currents in the full-depth ranges, the dominant mode of their variabilities and associated physical processes or underlying dynamics are still unknown. In this study, full-depth current measurements at 122.7°E, 18°N combined with other observations were used to investigate the dominant variability mode of western boundary currents and associated physical processes. The work highlights the barotropicity of western boundary currents variabilities, reveals the important role of large-scale forcings and mesoscale eddy activities, and clarifies the important contribution of intermediate and deep ocean to the western boundary currents variabilities and the LUC formation dynamics. This paper is

organized as follows. In Section 2, the data used in this paper are introduced. In Section 3, both barotropic and baroclinic partition and EOF analysis support the barotropic dynamics of western boundary currents variabilities east of Luzon. In Section 4, the possible physical processes responsible for “three abnormal events” featuring an intensified LUC during the observation period are investigated by currents and hydrographic observations, and the main conclusions and discussions are given in Section 5.

2 DATA AND METHOD

2.1 Data

2.1.1 Mooring observation

To observe the multiscale variabilities of western boundary currents east of Luzon, three moorings were deployed at 122.7°E (M1), 123°E (M2) and 123.3°E (M3) along the 18°N section from 26 January 2018 to 21 May 2020 under the support of Northwestern Pacific Ocean Circulation and Climate Experiment (see fig.1 in Ma et al. (2022)). At about 400-m depth, a main float with two 75-kHz Acoustic Doppler Current Profilers (ADCP) for upward and downward current measurements was equipped on each mooring to obtain velocity profiles from the sea surface to ~800-m water depth. All three moorings carried single-point current meters (Aquadopp DW) at 1 200-, 1 400-, and 1 600-m depths, and two additional current meters were also installed at 2 000 and 3 000 m at 122.7°E. Besides, each mooring carried six Seabird Electronics 37CTDs at nominal depths of 600, 800, 1 000, 1 600, 2 000, and 3 000 m, and seven 56T sensors at nominal depths of 450, 550, 650, 750, 850, 900, and 950 m to monitor ocean temperature and salinity. All of the instruments collected good quality data except that the down-looking ADCP at 123°E malfunctioned.

The ADCPs mounted on the moorings recorded hourly velocities with a vertical bin size of 8 m, and the current meters also collected hourly current data. The sampling interval for both CTDs and T sensors was 10 min. First, all of the raw data from ADCPs or current meters were interpolated vertically from 40 m to 3 024 m with an interval of 8 m according to the pressure data from ADCPs and current meters. Then, we used linear regression method to fill in a very small amount of vacant values, and formed gridded hourly current data. Finally, the

gridded hourly data were smoothed by 7-d running mean to remove tidal signals and were converted into daily average data. After averaging all of the raw temperature and salinity data with 10-min resolution to create hourly datasets, the temperature and salinity data were also vertically interpolated from 40 to 3 024 m with an interval of 8 m according to the pressure data for CTDs and ADCPs. Then the data were smoothed by 7-d running mean to remove tidal signals and were converted into daily data. The gridded daily velocities data, temperature and salinity data from 40 to 3 024 m with 374 levels at 122.7°E, 123°E, and 123.3°E were used in this study.

2.1.2 Other dataset

The Roemmich-Gilson Argo monthly data from January 2018 to May 2020 were also used in this study for comparison with hydrographic observations from the three moorings. The monthly Argo data with a resolution of $1^\circ \times 1^\circ$ was based on the analysis of Argo-only derived temperature and salinity fields using an analysis method similar to that described by Roemmich and Gilson (2009), including 58 vertical levels at depths of 2.5–1 975 m with spacing ranging from 10 to 100 m. The data were downloaded from http://sio-argo.ucsd.edu/RG_Climatology.html. For more details concerning the data and methods, readers are referred to Roemmich and Gilson (2009).

2.2 Method

2.2.1 Barotropic and baroclinic velocity from observations

The partitioning of the meridional velocity profile at M1 station (122.7°E, 18°N, water depth: 3 328 m) into barotropic and baroclinic components was performed. The barotropic component is the vertical average of the velocity profile, and the remaining sheared profile represents the baroclinic component, as shown in Eqs.1 & 2 below. Two ADCPs and five current meters at the M1 station provided the daily data from 40 to 3 024 m with an interval of 8 m. For the barotropic component calculation, the observed current velocities at the shallowest level of 40 m and the deepest level of 3 024 m were used to extrapolate the missing velocity values near sea surface and sea floor, respectively.

The specific formula for barotropic velocity by the mooring observations is defined as:

$$v_{BT}(t) = \frac{1}{3328 \text{ m}} \int_{3328 \text{ m}}^{0 \text{ m}} v(z, t) dz$$

$$= \frac{1}{3328 \text{ m}} \left[\int_{40 \text{ m}}^{0 \text{ m}} v(z, t)|_{z=40 \text{ m}} dz + \int_{3024 \text{ m}}^{40 \text{ m}} v(z, t) dz + \int_{3328 \text{ m}}^{3024 \text{ m}} v(z, t)|_{z=3024 \text{ m}} dz \right]. \quad (1)$$

The baroclinic velocity is: $v_{BC}(z, t) = v(z, t) - v_{BT}(t)$. (2)

By definition, vertical integration of baroclinic velocity within the full water depth range is always zero, i.e., $\int_{3328 \text{ m}}^{0 \text{ m}} v_{BC}(z, t) dz = 0$

2.2.2 Barotropic velocity estimated by Sverdrup theory

Classical Sverdrup theory states that Sverdrup transport in the ocean interior, $v_s(x, y, t) = \int_{-H}^0 v(x, y, z, t) dz$, is determined by local wind stress curl via the relation $\beta v_s(x, y, t) = \frac{\text{curl } \tau(x, y, t)}{\rho}$, where ρ is seawater density, β is the meridional gradient of Coriolis parameter. Assuming the volume transport of western boundary currents at 18°N balances the integrated Sverdrup transport in the ocean interior, barotropic velocity of western boundary current can be estimated as:

$$v_{BT_EST}(t) = -\frac{1}{HL_{WBC}} \int_{125^\circ E}^{105^\circ W} v_s(x, y, t)|_{y=18^\circ N} dx$$

$$= -\frac{1}{HL_{WBC}} \int_{125^\circ E}^{105^\circ W} \frac{\text{curl } \tau(x, y, t)}{\rho \beta} |_{y=18^\circ N} dx, \quad (3)$$

where $H=3\,000$ m represents the approximate water depth, $L_{WBC}=1^\circ$ is the assumed width of the western boundary current in this study, and the negative sign denotes the balance of volume transport near the western boundary and those in ocean interior.

This estimation method is dynamically justified due to the rapid barotropic response of the ocean to wind forcing, mediated by fast-propagating barotropic gravity waves.

3 DOMINANT MODE OF WESTERN BOUNDARY CURRENT VARIABILITIES

3.1 Barotropic/baroclinic partitioning

Figure 1 presents the original daily meridional velocity at M1, along with its barotropic and

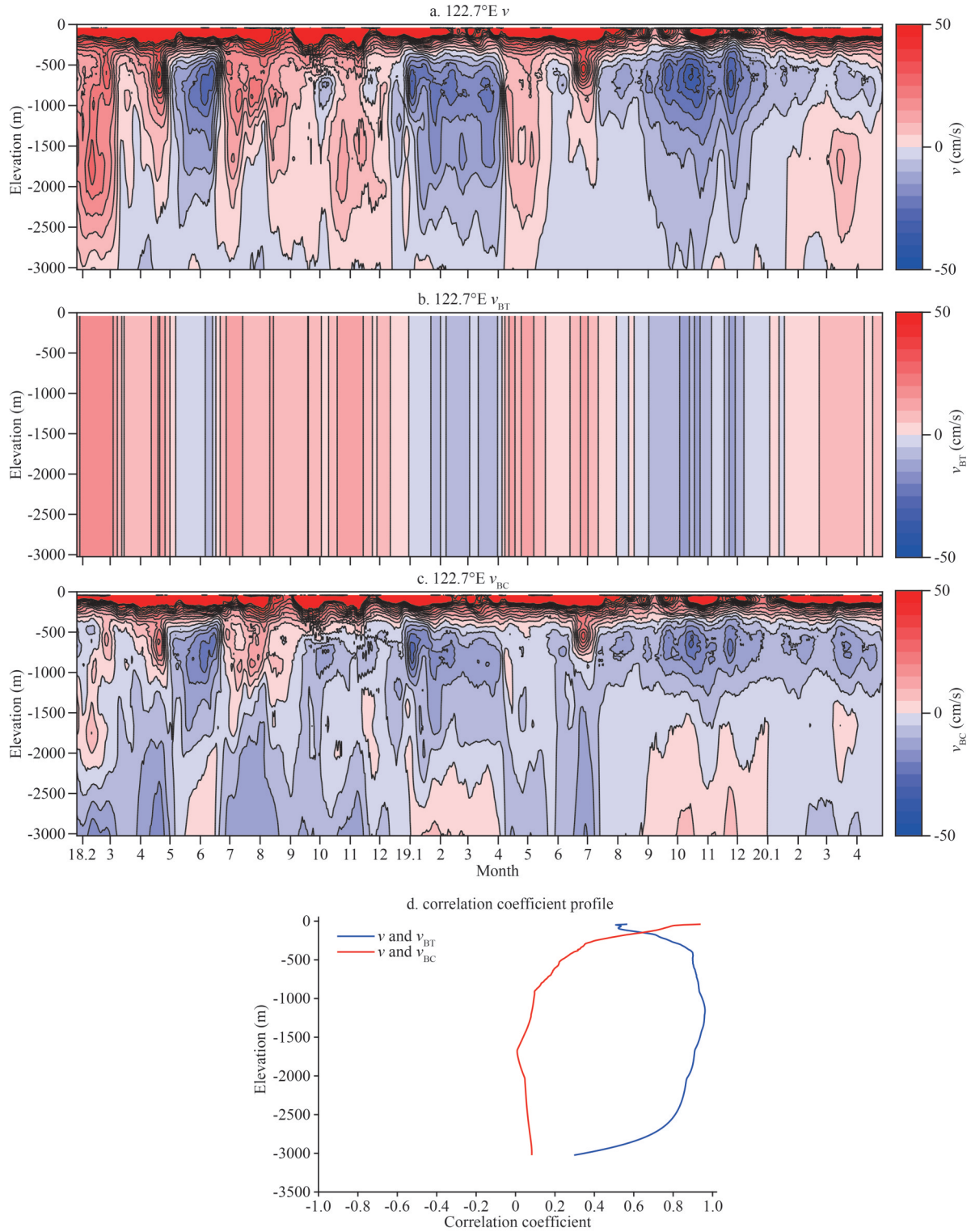


Fig.1 Original daily meridional velocity (a), barotropic (b), and baroclinic components (c) at M1 (18°N, 122.7°E) station from 26 January 2018 to 27 April 2020; profile of correlation coefficients between original meridional velocity and the barotropic component (blue line), and between original meridional velocity and baroclinic component (red line) calculated at their corresponding levels (d)

The time annotations on x-axis represent month, each tick means the first day of each month, and years of 2018, 2019, and 2020 are also indicated.

baroclinic components. The Kuroshio is predominantly confined to upper ~300 m (Fig.1a), and it is markedly stronger than that at the other two stations (figure not shown) because the M1 station is the closest to the velocity core of the Kuroshio (fig.3 in Yuan et al. (2024)).

Below the Kuroshio, southward current intermittently occurred with typical cores at about 600–700-m depth. The southward current generally extended even to 3 000-m depth, exhibiting vertical coherence below ~300 m (Fig.1a), and it was called as the LUC (Hu et al., 2013). A prominent feature observed is the occurrence of strong LUC events during three distinct periods, from May to mid-June 2018, from mid-December 2018 to March 2019, and from August 2019 to mid-January 2020 (Fig.1a). The three periods during which strong LUC occurred are hereafter referred to as the “three abnormal events”, which have good consistency with the three cooling events in the temperature of deep ocean in Section 4.2.

The three abnormal events were clearly discernible from the barotropic velocity component, and the times with southward barotropic meridional velocity component exhibits a strong correspondence with the times when the strong LUC occurred (Fig.1b). The mean barotropic velocity over the observation period was 3.1 cm/s northward. This translates to a mean northward volume transport of 10.3 Sv at 18°N assuming a width of the western boundary current of 100 km and the local water depth of 3 328 m. Notably, this mean northward barotropic transport at 18°N reversed to southward during each of the three abnormal events featuring an intensified LUC.

The baroclinic component reveals a distinct vertical velocity shear at approximately 300–400-m depth, featuring a northward baroclinic current in the upper layer associated with the Kuroshio and a generally southward baroclinic current in the lower layer associated with the LUC (Fig.1c). In contrast to the total observed LUC (Fig.1a), the southward baroclinic current below ~300–400 m remained relatively stable throughout the entire observation period, and it occasionally changed in short time when some extreme strong eddies passed by the M1 station (Fig.1c). Crucially, the three abnormal events are not distinctly captured by the baroclinic component alone (Fig.1c). Therefore, the partitioning of the meridional velocity profile indicates that barotropic component plays a more significant role in the variability of western

boundary currents at this location.

To further quantify the representativeness of barotropic and baroclinic components, the vertical profiles of correlation coefficients between the original meridional velocities and the barotropic/baroclinic components at corresponding depths were calculated (Fig.1d). Barotropic component is highly correlated ($R>0.8$) with observed velocities across a broad vertical range from 300 to 2 500 m (blue line in Fig.1d). In contrast, a high correlation with the baroclinic component is confined predominantly to the surface layer (red line in Fig.1d). Consequently, the barotropic velocity time series may serve as an effective proxy for the integrated variability of the western boundary current system east of Luzon.

3.2 Dominant variability mode

The dominant modes of variability in the western boundary current variabilities at M1 were identified using conventional Empirical Orthogonal Functions (EOF) analysis. Figure 2 displays the anomalies of original daily meridional velocity at M1, alongside the portions of the velocity anomalies explained by the first and second EOF modes respectively. The spatial patterns (EOF eigenvectors), and the principle component time series of the first two EOF modes are shown in Fig.3a & b. For comparison, the temporal evolution of the barotropic velocity (Fig.1b) and the depth-averaged (40–296 m) baroclinic velocity (Fig.1c) are also reproduced in Fig.3c.

The first EOF mode accounts for 74.2% of the total variance. The three abnormal events featuring an intensified LUC are distinctly evident in both the original velocity anomalies (Fig.2a) and the pattern reconstructed from the first EOF mode (Fig.2b). Spatially, the first EOF mode exhibits synchronous variability of the same sign throughout the entire water column (Figs.2b, 3a), indicating a dominant barotropic structure. This visual interpretation is confirmed quantitatively. The principle component time series of the first EOF mode (Fig.3b) is nearly identical to the time series of the barotropic velocity ($R=0.99$, Figs.1b & 3c). Thus, the first EOF mode essentially represents the barotropic variability of the western boundary current system, and the time series of the barotropic component can serve as a very good proxy for western boundary current variability. An interesting feature of the first mode's spatial pattern is that the strongest variability occurs between 500- and 1 000-m depth, the typical layer of the LUC core, rather than in the upper layer

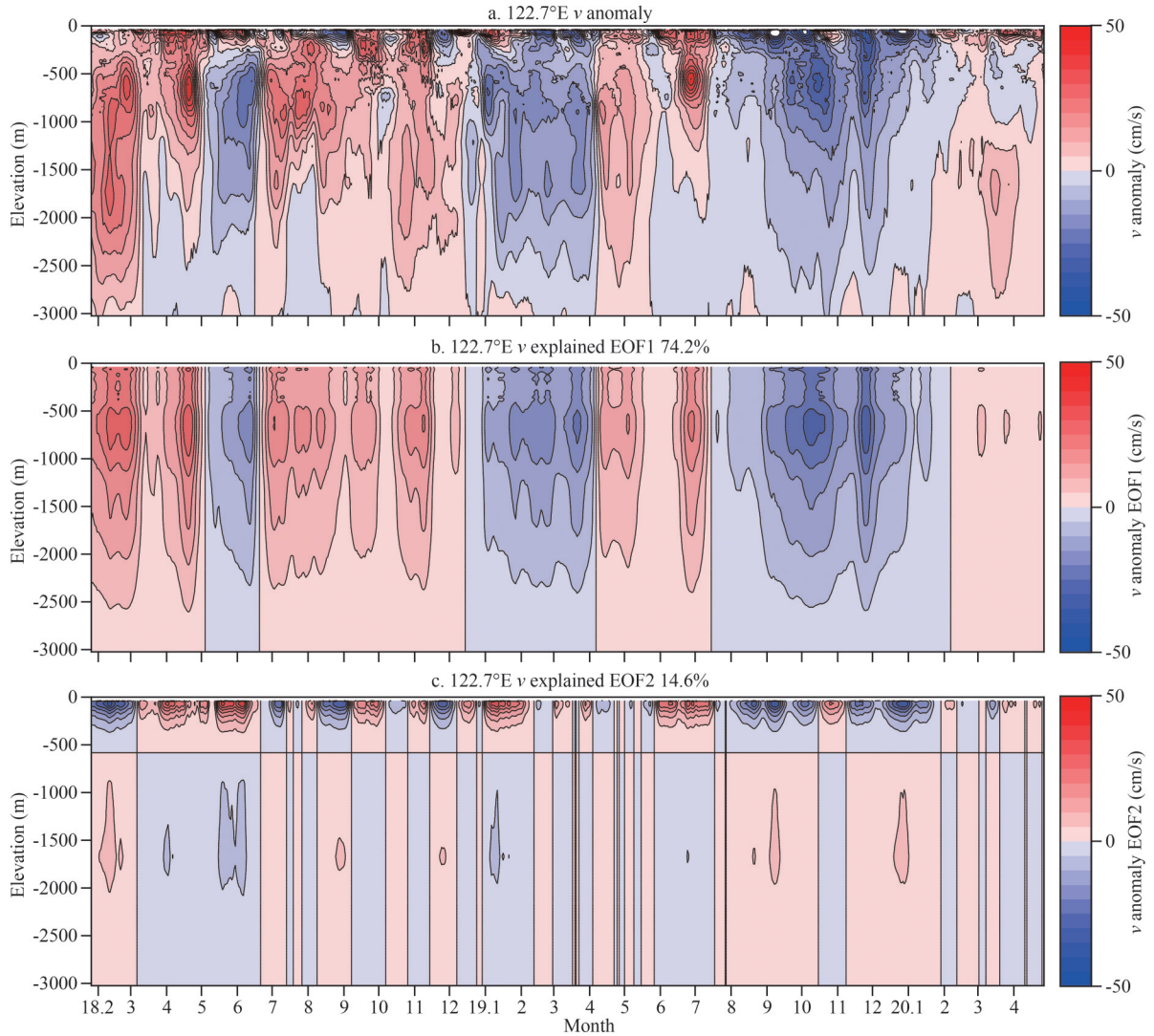


Fig.2 Original daily meridional velocity anomalies at M1 station from 26 January 2018 to 27 April 2020 (a); meridional velocity anomalies explained by the first EOF mode (b), and meridional velocity anomalies explained by the second EOF mode (c)

occupied by the Kuroshio. This indicates that, within this EOF mode, the LUC contributes more significantly to the overall variance than the Kuroshio. In fact, the vertical profile of the standard deviations for the original observed meridional velocities (Fig.2a) has two maximum values of 0.23 m/s near 60-m depth and 0.15 m/s near 600-m depth and have a minimum value of 0.12 m/s near 400 m (figure not shown), showing substantial variability within the 500–1 000-m depth range.

The second EOF mode explains 14.6% of the total variances. Its spatial pattern displays a dipole structure with opposing variations above and below about 600-m depth, with the strongest signals concentrated in the upper 300 m (Figs.2c & 3a). As

expected, the principle component time series of the second EOF mode (Fig.3b) is highly correlated with the temporal evolution of the baroclinic velocity averaged in the upper 300 m (Fig.3c, $R=0.96$). This confirms that the second EOF mode primarily represents the baroclinic component associated with the Kuroshio in the upper ocean.

Quantitatively, the variances of the barotropic and baroclinic components from the original meridional velocity and those from the velocity anomalies explained by the first and second EOF modes were calculated (Table 1). The sum of the variances of observed daily meridional velocity across all 374 vertical levels (40–3 024 m) in Fig.1a is $3.9097 \text{ m}^2/\text{s}^2$. This total variance can be

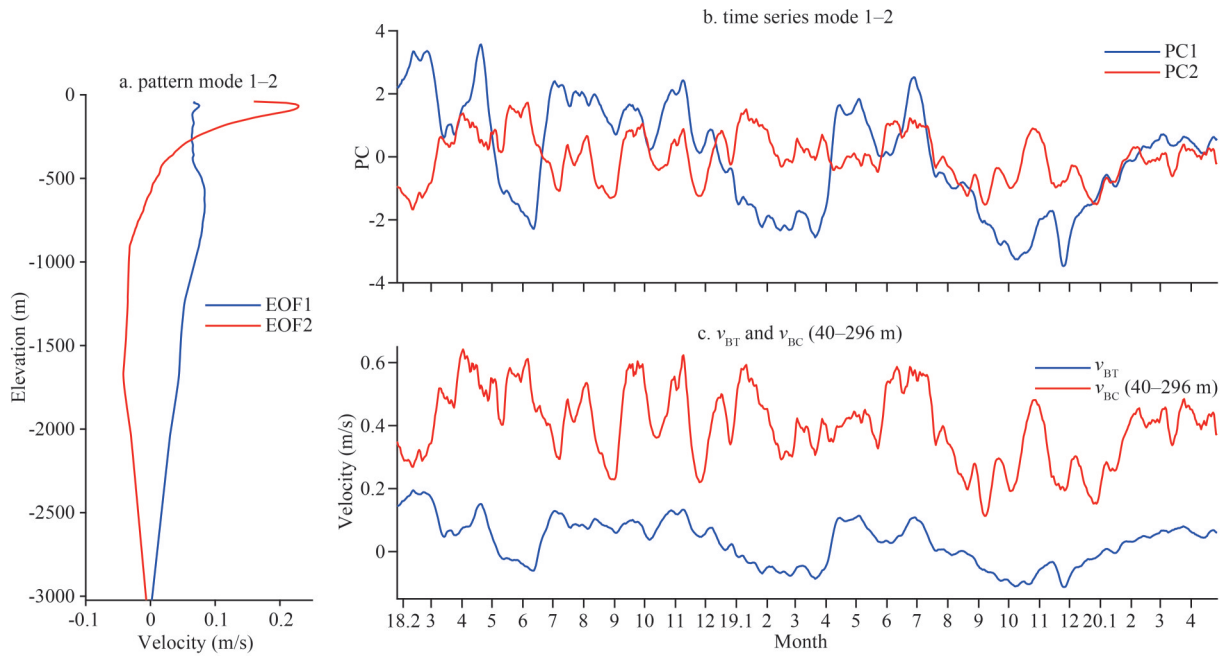


Fig.3 The spatial patterns of the first (blue line) and second (red line) EOF modes (a); the time series of the first (blue line) and second (red line) EOF modes (b); the evolutions of the barotropic velocity (m/s, blue line) and the baroclinic velocity averaged between 40 and 296 m (c)

PC: principle component.

Table 1 Variances of the barotropic and baroclinic components from the original meridional velocity and the velocity anomalies explained by the first two EOF modes

	Total variance (m^2/s^2)	Variance of barotropic component (m^2/s^2)	Variance of baroclinic component (m^2/s^2)	Twice covariance of BT and BC components (m^2/s^2)
Original daily velocity	3.909 7	1.941 7	1.642 9	0.325 1
Velocities explained by EOF1	2.902 5 (74.2%)	1.893 5	0.679 1	0.329 9
Velocities explained by EOF2	0.571 4 (14.6%)	0.007 9	0.558 9	0.004 5

BT: barotropic, BC: baroclinic.

decomposed into the variance of barotropic component in Fig.1b ($1.941 7 \text{ m}^2/\text{s}^2$, 50% contribution), the variance of baroclinic component Fig.1c ($1.642 9 \text{ m}^2/\text{s}^2$, 42% contribution) and twice the covariance between them ($0.325 1 \text{ m}^2/\text{s}^2$, 8% contribution).

A parallel calculation was made for the velocity anomalies explained by the first EOF mode in Fig.2b. The first EOF mode explains the variances of $2.902 5 \text{ m}^2/\text{s}^2$ (74.2% of original total variances), and it captures nearly the entire barotropic signal ($1.893 5 \text{ m}^2/\text{s}^2$), a small part of the baroclinic signal ($0.679 1 \text{ m}^2/\text{s}^2$), and two times of their covariances ($0.329 9 \text{ m}^2/\text{s}^2$, Table 1). If the first EOF mode is taken as the representative mode of meridional velocity variability, it can be concluded that the variability of the western boundary current east of Luzon is fundamentally barotropic in nature. In

contrast, the second EOF mode explained the variances of $0.571 4 \text{ m}^2/\text{s}^2$ (14.6% of original total variance). As expected, this mode is almost entirely composed of baroclinic components ($0.558 9 \text{ m}^2/\text{s}^2$), especially in the upper 300 m. Therefore, the barotropic component not only contributes a larger share of the total variance than the baroclinic component, but is also almost entirely captured by the first EOF mode.

To clarify the timescale of the western boundary current variability addressed in this study, spectral analysis on three time series: the barotropic velocity (Fig.1b), the Kuroshio velocity (averaged from 50 to 300 m) and the LUC velocity (averaged from 500 to 2 500 m) were performed (figure not shown). The analysis reveals that the observed currents exhibit significant spectral peaks near the periods of 230–240 d. As will be shown in the next section, the

three abnormal events with a strengthened LUC are linked to physical processes on intraseasonal or seasonal timescales. Therefore, the variability of western boundary currents discussed herein occurs primarily on intraseasonal to seasonal timescales.

In summary, barotropic component plays a dominant role in the variability of western boundary currents east of Luzon on the intraseasonal to seasonal timescales. It is concluded that the variability of this current system has a significant barotropic nature on these timescales. The southward barotropic velocities during the observation period were predominantly the result of the three abnormal events featuring an intensified LUC. The underlying dynamics and forcings of these three abnormal events will be the focus of the next section.

4 POSSIBLE EXPLANATION FOR “THREE ABNORMAL EVENTS” FEATURING AN INTENSIFIED LUC

The most prominent feature in the mooring observations at M1 is the occurrence of strong LUC during three distinct periods: May to mid-June 2018, mid-December 2018 to March 2019, and August 2019 to mid-January 2020 (Fig.1a), which is called “three abnormal events”. The three abnormal events are investigated in this section to elucidate their potential causes and illuminate underlying dynamics governing western boundary current variability.

4.1 Sverdrup transport balance

The preceding section established the barotropic nature of western boundary current variability east of Luzon, with the three abnormal events corresponding to pronounced negative anomalies in barotropic velocity. Therefore, I examine whether the three abnormal events can be explained within the framework of classical Sverdrup theory. Assuming the volume transport of western boundary currents at 18°N balances the integrated Sverdrup transport in the ocean interior, barotropic velocity of western boundary current can be estimated by wind stress curl in the Pacific Ocean. Daily wind fields both NCEP (National Centers for Environmental Prediction) and ERA (ECMWF Reanalysis, ECMWF-European Centre for Medium-Range Weather Forecasts) reanalyses were used to compute wind stress following the friction coefficient in Large and Pond (1981). Wind stress curl was calculated using a centered finite-difference scheme.

Sverdrup transport at 18°N was calculated by integrating wind stress curl along the latitude in the ocean interior, and the estimated barotropic velocity was obtained by Eq.3 in Section 2. It is worth noting that the method to estimate barotropic velocity is valid due to rapid barotropic response to the ocean to the wind forcing, mediated by fast-propagating barotropic gravity wave velocity.

Figure 4a compares the estimated barotropic velocity from two wind products with that from mooring observations at 18°N . The two estimation series based on NCEP and ERA wind forcings show generally good agreement, depicting northward barotropic velocity interrupted by periods of southward flow during August–November 2018 and September–December 2019 (Fig.4a). Because the latitude of 18°N is located in the subtropical gyre circulation, the wind stress curl is generally negative values across the basin. Positive wind stress curls occurring during the autumn and winter of 2018 and 2019 (Fig.4b–c) are associated with the seasonal meridional migration of the basin-scale wind fields. Comparison with the observed barotropic velocity shows that the first abnormal event (May–June 2018) and the second abnormal event (January–March 2019) are not captured by the Sverdrup estimates. In contrast, the third abnormal event (September–December 2019) shows remarkable consistency between estimated and observed barotropic velocities (Fig.4a).

Moreover, the two prominent southward velocity peaks observed in October 2019 and November/December 2019 (Figs.1a, 2a–b, & 4a) are particularly well reproduced by the estimated barotropic velocity, although the magnitude of the latter is slightly underestimated. Therefore, it can be concluded that the third abnormal event is primarily driven by the basin-scale wind stress curls at 18°N through the Sverdrup transport balance. The third abnormal event persisted for about 6 months, during which the entire water column from the surface to bottom flowed anomalously southward (Fig.2a), consistent with a steady wind-driven barotropic response.

In summary, the third abnormal event is mainly caused by anomalous wind stress curl patterns during the autumn and winter of 2019 at 18°N in the Pacific ocean interior, which generated northward transport in the ocean interior that were compensated by sustained southward current in the whole water column in the western boundary. The next part utilizes hydrographic data to demonstrate

that the temperature anomaly in the third abnormal event is different from those in the first and second abnormal events, providing independent support for distinct dynamic mechanism of the third abnormal event.

4.2 Hydrographic signature and subsurface eddy

Figure 5 shows temperature records from all three mooring stations, while Fig.6 shows the velocity anomalies at M1 station and the temperature anomalies across the mooring array. Three pronounced cooling episodes are evident in the records, particularly below 1 000 m, occurring

from May to July 2018, January to April 2019, and August 2019 to January 2020. These cooling periods align precisely with the three abnormal events identified in the velocity data (Figs.5 & 6c–e). However, the characteristics of the first two abnormal events were of shorter duration and exhibited significant vertical fluctuations of the isothermal surfaces in intermediate and deep layers. In contrast, the third event persisted a longer duration with weaker vertical fluctuation of the isothermal structure (Fig.5). Additionally, during the first two events, cooling was primarily confined to the lower layer, with even an increase of temperature in the

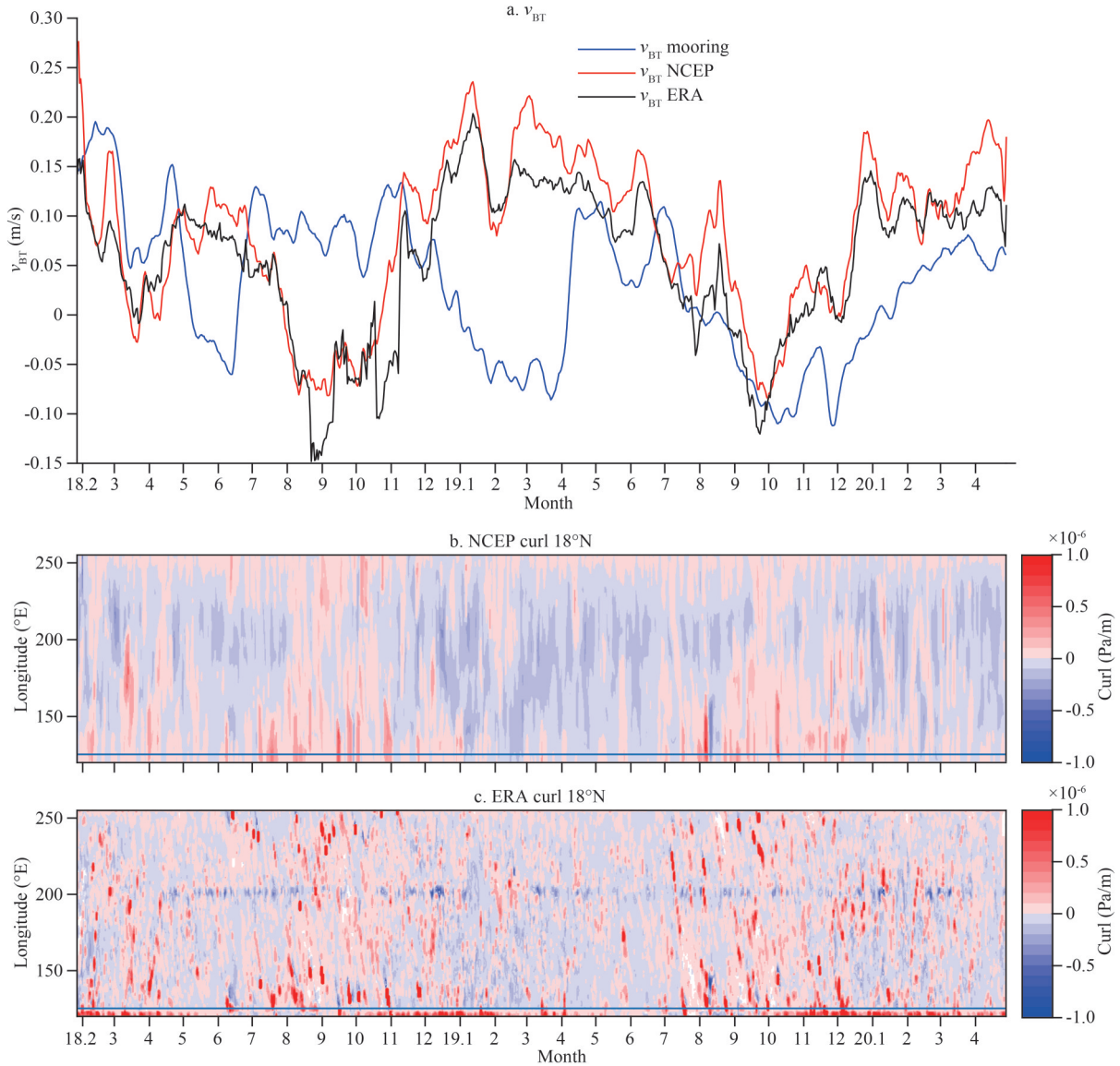


Fig.4 Comparing mooring observed barotropic velocity (m/s, blue line) with estimated barotropic velocity of western boundary currents east of Luzon by using NCEP (red line) and ERA (black line) wind fields along 18°N from 155°E to 105°W (a); the time-longitude plots for the NCEP (b) and ERA (c) wind stress curl along 18°N

The blue lines in (b) and (c) indicate the position of 125°E.

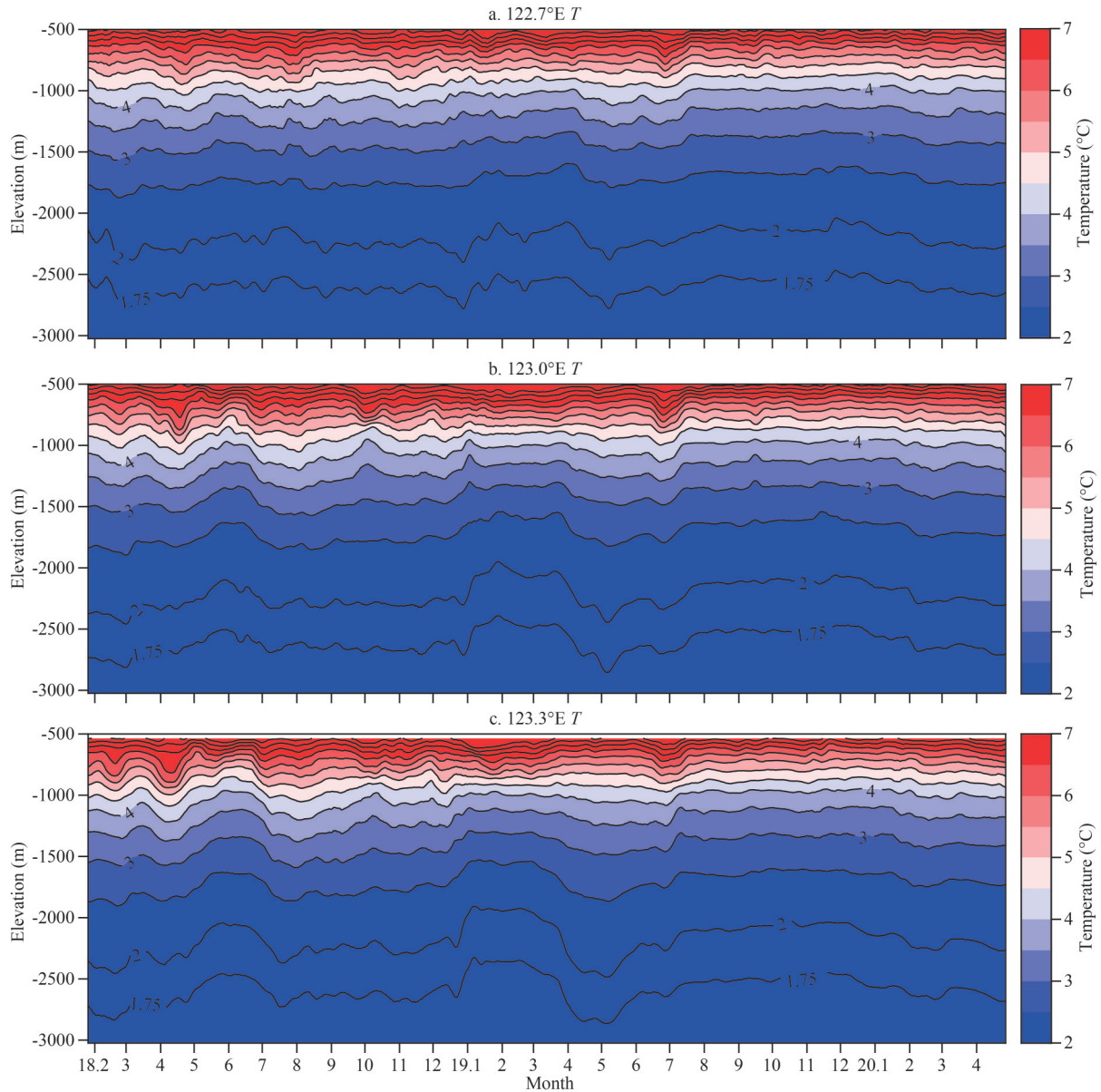


Fig.5 Mooring observed daily temperature at M1 (a), M2 (b), and M3 (c) stations from 26 January 2018 to 27 April 2020

upper ocean. The third cooling event, however, involved synchronous cooling throughout the entire observed water column (Figs.5 & 6). These differences suggest that the first two abnormal events resulted from relatively short-term processes in the intermediate and deep layers, potentially linked to ocean eddies or wave activities. The third abnormal event, conversely, appears to have influenced the entire water column consistently, with comparable cooling amplitudes observed at all three stations, which is consistent with the explanation of the Sverdrup transport theory proposed in Section 4.1.

Regarding the first two abnormal events, the

amplitude of temperature decrease varied considerably among the three stations, being largest at M3 (the easternmost station) and the smallest at M1 (Fig.6). This spatial pattern implies a westward propagation of the anomalous signal from the ocean interior. A detailed analysis of the zonal and meridional velocity anomalies during the first abnormal event reveals the passage of a westward-propagating subsurface cyclonic eddy. Prior to 20 June 2018, the southern flank of this eddy passed the M1 station, generating a southeastward anomalous flow. The eddy center reached the M1 longitude on 20 June 2018 when the cyclonic eddy caused an eastward velocity anomaly, with the

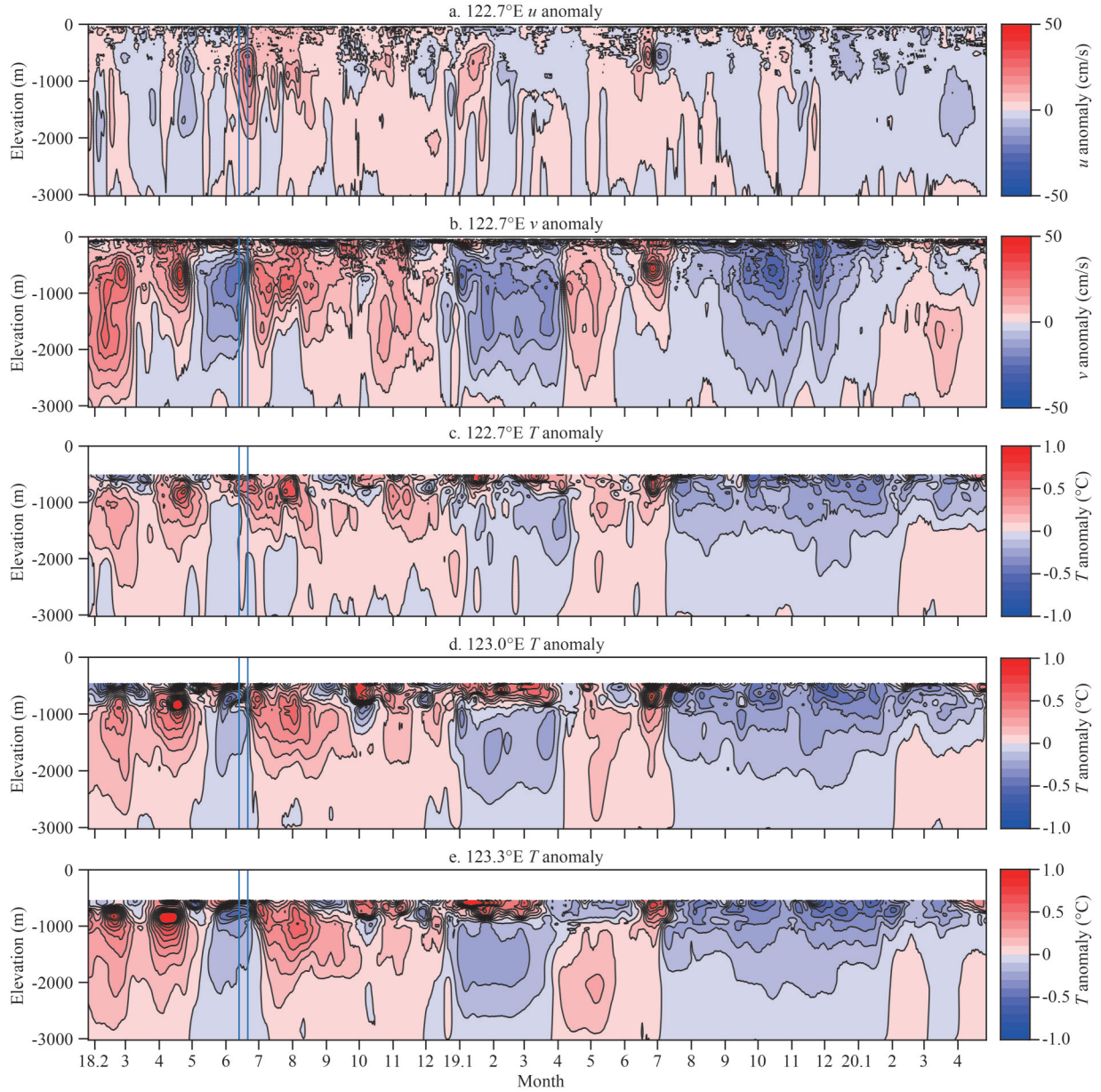


Fig.6 Mooring observed daily zonal (a) and meridional (b) velocity anomalies at M1 station, and mooring observed daily temperature anomalies at M1 (c), M2 (d), and M3 (e) stations from 26 January 2018 to 27 April 2020

The two days of 12 June 2018 and 20 June 2018 are indicated (blue lines).

maximal zonal velocity and negligible meridional component. After 20 June 2018, as the cyclonic eddy continued westward, the M1 station experienced a northeastward abnormal flow (Fig.6a–b). Therefore, I conclude that the first abnormal event, characterized by its short duration, was primarily induced by the passage of a subsurface eddy passing through the intermediate and deep layers.

The evolution of velocity vectors during the second abnormal event is more complex, and it is difficult to find the definitive cause for this

abnormal event. Nevertheless, it is evident that this abnormal event is also related to the dynamic processes of the intermediate and deep layers (below 1 000 m), because temperature in the intermediate and deep layers decreases, while temperature in the upper layer increases (Fig.6). In fact, the maximum vertical displacements of the isotherms during both the first and second events occurred at $\sim 2\,000$ m (Fig.5), indicating that the energy sources for these anomalies resided in the intermediate and deep ocean.

4.3 Connection with the intermediate and deep ocean

As established in Section 3, barotropic velocity serves as a robust representation of the western boundary current variability. To explore its thermodynamic connections, a correlation analysis between the barotropic velocity and temperature at different depths was conducted (Fig.7). The analysis reveals a remarkably high correlation between barotropic velocity and temperature evolution below 1 000 m (Fig.7a). Figure 7b compares the normalized barotropic velocity with the normalized, depth-averaged temperature from 1 000 to 3 024 m at M1, M2, M3, and their spatial average. The correlation coefficients are 0.81, 0.90, 0.85, and 0.89, respectively, confirming a strong linkage between barotropic velocity and deep-ocean thermal structure. Notably, the consistency between the strong LUC and negative temperature anomalies is more pronounced during the first two events than during the third (Fig.7b), further corroborating their distinct dynamical origins as proposed earlier.

The important role of intermediate and deep ocean temperature has been further elucidated by Tian et al. (2026). Their analysis demonstrated substantial differences in the calculated meridional geostrophic velocity depending on the choice of reference level (no-motion level). The strength of the LUC derived geostrophically using the 3 000-m reference level agrees well with direct measurements, whereas it is significantly underestimated when a 1 500-m level is adopted (see their fig.13). This

finding underscores that realistic representation of the deep ocean below 1 500 m is essential for accurately quantifying the LUC and its variability. Consequently, the dominant variability of the western boundary currents east of Luzon, approximated by the barotropic component, is primarily governed by temperature variations in the intermediate and deep ocean (below 1 000 m), rather than by those in the upper and subsurface layers.

Among the three moorings, the temperature at M3 (the easternmost station) exhibits the highest correlation with the barotropic velocity at M1. This spatial distribution suggests abnormal signals typically originate from ocean interior, propagate westward, and are ultimately constrained by the western boundary. Dynamic calculations indicate that a temperature decrease at M3 is associated with a southward velocity anomaly, manifesting as a strong LUC. These interior temperature decreases can be forced by basin-scale wind stress curl anomalies (as in the third abnormal event), or by the passage of subsurface ocean mesoscale eddies (as in the first abnormal event). In addition, the steep topography in the vicinity of M3 may facilitate the trapping of topography Rossby waves, providing another potential mechanism for influencing the intermediate and deep layers. The role of topography warrants further investigation in the future.

The T - S diagram of station M3 (18°N, 123.3°E) during the three abnormal events offer further insights into water mass characteristics (Fig.8).

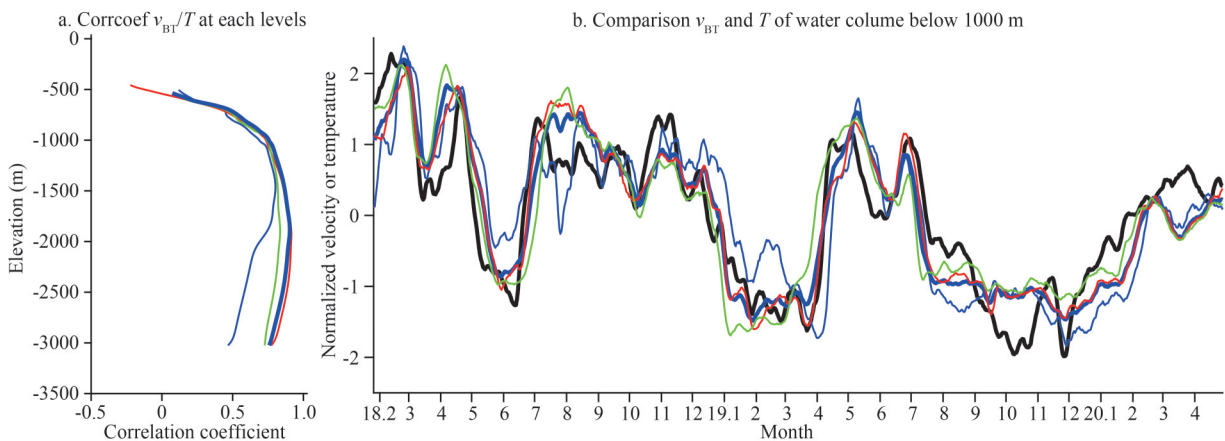


Fig.7 Profile of correlation coefficients between barotropic velocity component and the temperature at M1 (blue line), M2 (red line), M3 (green line) stations and the averaged temperature in the three stations (thick blue line) calculated at their corresponding levels (a); comparison of the normalized barotropic velocity (thick black line) against the normalized mean temperature of the water volume from 1 000 to 3 024 m at M1 (blue line), M2 (red line), M3 (green line), and the averaged temperature in the three stations (thick blue line) (b)

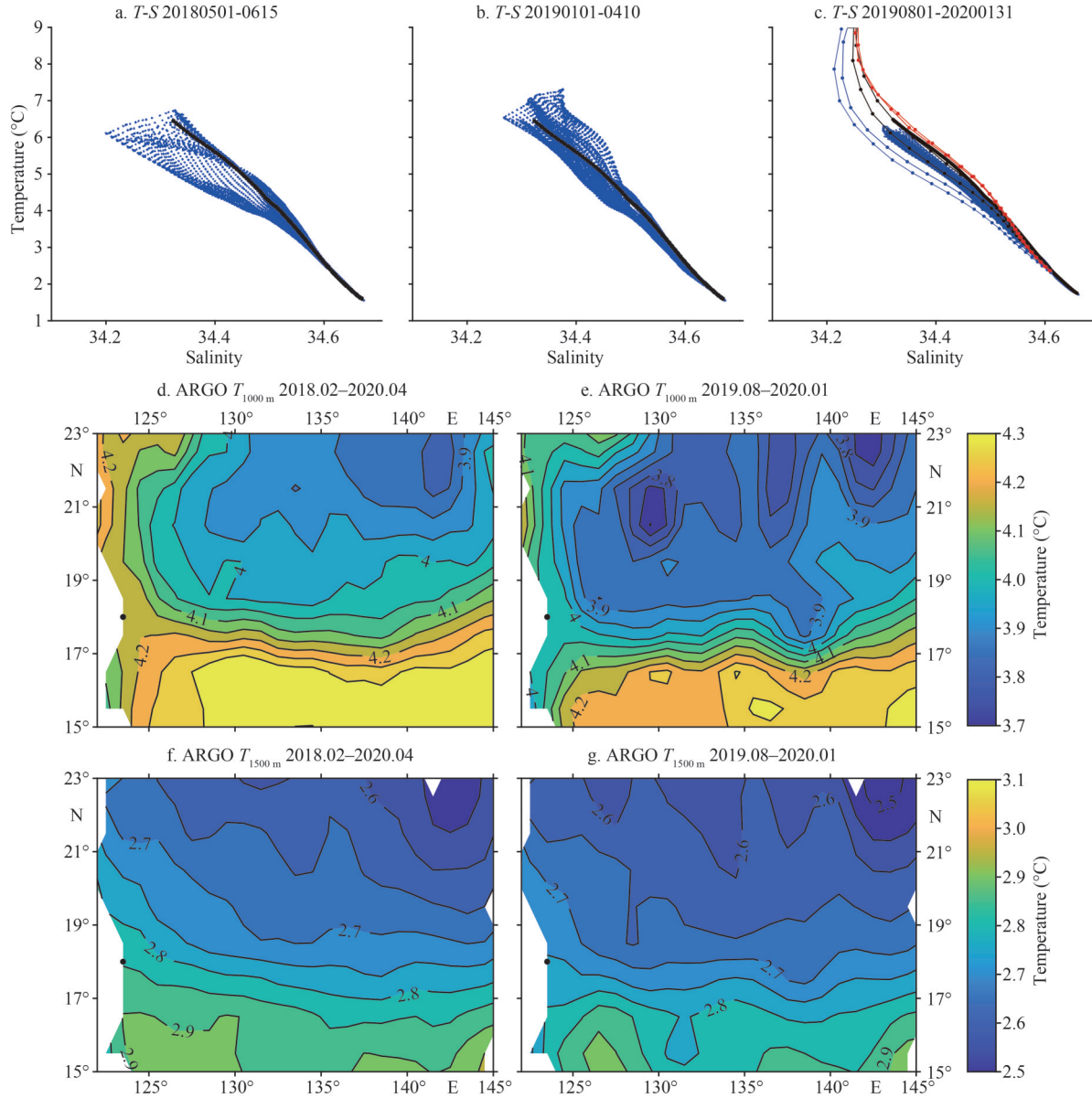


Fig.8 Mooring observed T - S diagram at M3 from 1 May 2018 to 15 June 2018 with the mean temperature/salinity curve (thick black line) during the observation period (a); same as (a) but from 1 January 2019 to 10 April 2019 (b); (c) same as (a) but from 1 August 2019 to 31 January 2020, superimposed by the mean temperature/salinity curves from ARGO data from August 2019 to January 2020 at the two stations (123.5°E, 15.5°N) and (123.5°E, 16.5°N) (red line), at the two stations (123.5°E, 17.5°N) and (123.5°E, 18.5°N) (black line), and at the two stations (123.5°E, 19.5°N) and (123.5°E, 20.5°N) (blue line); the mean temperature from ARGO data at 1 000 m from February 2018 to April 2020 (d) and from August 2019 to January 2020 (e); same as (d) and (e) but at 1 500 m (f & g). The black points at (d-g) indicate the position of (123.5°E, 18°N).

Compared to the mean T - S profile during the mooring period (black line), water particles during the prolonged third abnormal event exhibits colder and fresher properties and are more tightly clustered in T - S space than those in the first two events (Fig.8a-b). This distinct T - S signature provides independent evidence supporting a different

dynamic mechanisms for the third event (Fig.8c). For context, the mean T - S properties from Argo data at six stations (15.5°N–20.5°N) along the 123.5°E during the mooring period are presented (Fig.8c). At 19.5°N and 20.5°N, there is a significant low salinity characteristic (blue lines in Fig.8c), while at 15.5°N–16.5°N, the salinity is slightly higher (red

lines in Fig.8c). During the mooring observation period, the mean T - S curve at M3 (thick black line in Fig.8c) closely aligns with the Argo-derived mean at 17.5°N (black line in Fig.8c), whereas during the third abnormal event, the T - S properties at M3 (blue points in Fig.8c) shift toward the Argo-derived mean at 18.5°N (black line in Fig.8c). This implies that the deep waters during the third abnormal event were likely advected from the north of the observation site. Accordingly, the strong LUC during the third abnormal event transported colder and fresher water masses southward to the 18°N section. This inference is further supported by the horizontal distribution of Argo-derived temperature at 1 000 and 1 500 m (Fig.8d–g), which shows that water north of 18°N was generally cooler during the third event than the mean state over the entire mooring period. In summary, the third abnormal was associated with basin-wide oceanic adjustments driven by large-scale wind fields, which facilitated the southward transport of colder and fresher water masses to the 18°N section via an intensified LUC.

In contrast, the more scattered T - S distributions during the first two events (Fig.8a–b) suggest more complex water mass sources. Moreover, their shorter duration and coarse temporal resolution of monthly Argo data preclude an analysis of possible water mass origins.

5 CONCLUSION AND DISCUSSION

The dominant variability mode of the western boundary currents east of Luzon Island within the full water depth range and related physical processes were investigated by using more than two-year mooring observations at 18°N/122.7°E. It is revealed that barotropicity is the nature of the dominant variability in the western boundary current on intraseasonal to seasonal timescales. Barotropic velocity component serves as an effective proxy for this variability, which is closely related to temperature in the intermediate and deep ocean, typically below 1 000 m.

During the observation period, the temporal evolutions of the western boundary current was characterized by three abnormal events featuring an intensified LUC. The third abnormal event differed markedly from the first two both in their durations and temperature/velocity anomalies in the vertical direction. The duration of the third abnormal event is about 6 months, mainly related to the wind stress curl on the basin scale, while the first two abnormal

events were of much shorter duration and may be related to local subsurface mesoscale eddy (or submesoscale eddy) activity. Both large-scale wind fields and mesoscale processes may lead to a decrease in ocean temperature in the intermediate and deep layers, resulting in a strengthened LUC, which dominates the variability of the western boundary current within the entire water column.

In order to further clarify the nature of barotropic dynamics and understand why barotropic velocity is closely related to both the LUC and temperature in the intermediate and deep ocean, a simplified two-layer system (only including barotropic mode and the first baroclinic mode) is used to examine the relative contributions of barotropic and baroclinic components in the two layers to the variability of western boundary currents.

The western boundary currents east of Luzon can be conceptually represented as a two-layer system, which includes a northward Kuroshio in the upper layer (approximately 0–300 m) and a southward LUC in the intermediate and deep layers (below 300 m). The Kuroshio in the upper layer (subscript 1 in Eq.4) and the LUC in the intermediate and deep layers (subscript 2 in Eq.5) can be expressed by barotropic velocity and baroclinic velocities in different layers (Eqs.4 & 5). The mean thickness of the first layer is much smaller than that of the second layer, with typical values of $H_1=300$ m and $H_2=3\ 000$ m (Fig.1c).

$$v_1(t)=v_{BT}(t)+v_{1BC}(t), \quad (4)$$

$$v_2(t)=v_{BT}(t)+v_{2BC}(t). \quad (5)$$

In Section 3, it has been proven through mooring observations that the variance of barotropic current ($1.941\ 7\ \text{m}^2/\text{s}^2$) is greater than that of baroclinic currents ($1.642\ 9\ \text{m}^2/\text{s}^2$) in the full water column (Table 1). That is to say,

$$\text{VAR}(v_{BT}(t)) \times (H_1+H_2) > \text{VAR}(v_{1BC}(t)) \times H_1 + \text{VAR}(v_{2BC}(t)) \times H_2, \quad (6)$$

where VAR represents variance.

In the upper ocean, the Kuroshio velocity and its variations are primarily dominated by its baroclinic component. The barotropic velocity (Figs.1b & 3c, black line in Fig.9a) is much smaller than the baroclinic velocity of Kuroshio (Figs.1c & 3c, red line in Fig.9a). Moreover, the variance of the barotropic velocity in the upper ocean (black line in Fig.9b) is also much smaller than that of the baroclinic velocity (red lines in Figs.9b & 3c).

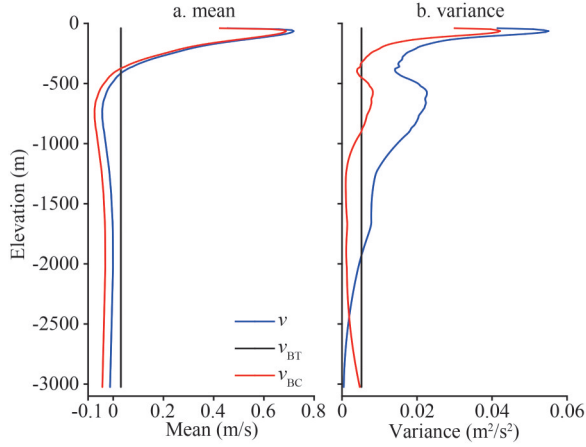


Fig.9 Comparison between mean meridional velocity (m/s, blue line) and its barotropic (m/s, black line) and baroclinic (m/s, red line) components (a); comparison between their variances (m²/s²) shown in (a) (b)

$$\text{VAR}(v_{BT}(t)) \times H_1 < \text{VAR}(v_{BC}(t)) \times H_1. \quad (7)$$

Combining Eq.6 and Eq.7, it can be inferred that

$$\text{VAR}(v_{BT}(t)) \times H_2 > \text{VAR}(v_{BC}(t)) \times H_2. \quad (8)$$

Equation 8 means that the variability of barotropic velocity in the intermediate and deep layers is much greater than that of baroclinic velocity, which is consistent with observational results (black and red lines in Fig.9b). It is worth noting that in the mean state, the magnitudes of the barotropic and baroclinic velocities in the intermediate and deep ocean are comparable (Fig.9a), but their variability amplitudes are completely different.

In this simplified two-layer system, the variability in the upper layer is almost due to the variance contribution of its baroclinic velocity, while the variability in the intermediate and deep layers is almost due to the contribution of barotropic velocity. When considering $v_{BC}(t) \times H_1 + v_{BC}(t) \times H_2 = 0$, it is not difficult to conclude from Eq.6 that the variance of barotropic velocity in the intermediate and deep layers (the left side of Eq.8) is greater than the variance of baroclinic velocity in the upper layer (the right side of Eq.7). Therefore, the nature of barotropic dynamics in this study refers to the fact that the variance contribution of barotropic component (in-phase variabilities of the Kuroshio and the LUC) exceeds that of baroclinic component (out-of-phase variabilities of the Kuroshio and the LUC) in the entire water column. Consequently, it is reasonable that the temporal evolution of barotropic

component is closely related to the evolution of the observed LUC in the intermediate and deep layers (Eq.8).

In the two-layer system where the thickness of the first layer is much smaller than that of the second layer, it can be deduced from the geostrophic balance equation that the pressure gradient caused by temperature variability in the upper layer mainly affects the upper layer current, with minimal influence on the intermediate and deep layer current. In contrast, temperature variability in the intermediate and deep layers can directly affect the current variability in the intermediate and deep layer, mainly contributing to barotropic velocity in the full-depth range. Therefore, according to Eq.8, it can be understood why there is a close relationship between barotropic velocity and temperature variability in the intermediate and deep layers (Fig.7).

As mentioned earlier, variability of barotropic component is more important than that of baroclinic component. Now consider an extreme case under a two-layer system, in which baroclinic component (the velocity difference between the Kuroshio and LUC) keeps unchanged over all time and only barotropic velocity varies in the entire water column. In the mean state, the sea surface slopes upward toward the offshore, while the thermocline slopes downward toward the offshore (black lines in Fig.10). When a southward barotropic velocity anomaly occurs in the entire water column (possibly due to the decrease of temperature in the intermediate and deep layers), the Kuroshio weakens slightly, the LUC intensifies, and the sea surface becomes gentler but the thermocline slope should be almost unchanged (blue lines in Fig.10). When a northward barotropic velocity anomaly occurs, the Kuroshio strengthens slightly, the LUC weakens or even reverses, and the sea surface becomes steeper but the thermocline slope should be almost unchanged (red lines in Fig.10). Obviously, the extreme case here did not take into account the variability of baroclinic component in the upper ocean associated with wind fields, mesoscale eddies, or wave-mean current interactions.

This study argues that the barotropic mode is more important. However, based on the same mooring observations at 18°N/122.7°E, Ma et al. (2022) concluded that the first two baroclinic modes are more dominant. This discrepancy arises because Ma et al. (2022) did not consider the barotropic mode in vertical mode decomposition of current

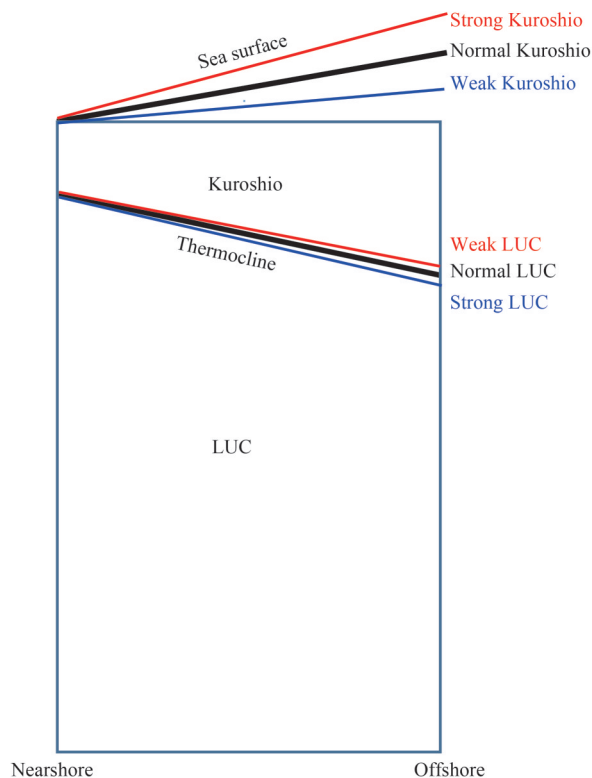


Fig.10 Schematic diagram of barotropic dynamics (without baroclinic contribution) governing variability of the western boundary currents east of Luzon

velocities (see their fig.9 and fig.10). The reconstructed meridional velocity using the first three baroclinic modes in Ma et al. (2022, see their fig.10b) closely resemble the baroclinic velocity component in this study (Fig.1c). Moreover, they also mentioned that the velocity below 2 000 m could not be adequately reproduced by the superposition of the first three or even more baroclinic modes, and speculated that this might be due to nonlinear dynamic processes or background stratification. Therefore, it is robustly demonstrated that barotropic mode is very important, especially in reproducing meridional velocity in the intermediate and deep ocean (Figs.1 & 2). The difference from Ma et al. (2022) further implies that the barotropic dynamics in this study, governing variabilities of western boundary current east of Luzon, are closely link to the variabilities of the LUC (Fig.1d), and are also strongly correlated with the variabilities of temperature in intermediate and deep ocean (Fig.7b).

Previous studies have demonstrated the importance of surface mesoscale eddies or atmospheric intraseasonal forcing in modulating the variability of

ocean currents (e.g., Hu et al., 2013; Lien et al., 2014; Yuan et al., 2021). These processes primarily originate from surface processes, which can cause intraseasonal variability east of Luzon. As shown in Fig.6, the temperature and velocity have significant intraseasonal variability in the upper 1 000 m. However, this study emphasizes the potential existence of another important signal source in the intermediate and deep ocean. The first two cooling events with large displacements of isotherms (Fig.5) serve as compelling evidence for additional signal source in the deep ocean east of Luzon. This signal source in the deep ocean may be associated with the topographic Rossby wave or upwelling that can be induced by the steep topography (large β_{TOPO}) or western boundary. More details about the signal source in the deep ocean are left for future studies.

6 DATA AVAILABILITY STATEMENT

The datasets generated and/or analyzed during the current study are available from the corresponding author on reasonable request.

7 ACKNOWLEDGMENT

The authors are grateful to all crews during the two cruises of mooring deployment and recovery. Monthly Argo data is available from the website http://sio-argo.ucsd.edu/RG_Climatology.html.

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