Variability of the Mindanao Current Induced by El Niño Events

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ABSTRACT

Historical observations have documented prominent changes of the Mindanao Current (MC) during El Niño events, yet a systematic understanding of how El Niño modulates the MC is still lacking. Mooring observations during December 2010–August 2014 revealed evident year-to-year variations of the MC in the upper 400 m that were well reproduced by the Hybrid Coordinate Ocean Model (HYCOM). Composite analysis was conducted for 10 El Niño events during 1980–2015 using five model-based datasets (HYCOM, OFES, GEOS-ODA, SODA2.2.4, and SODA3.3.1). A consensus is reached in suggesting that a developing (decaying) El Niño strengthens (weakens) the MC, albeit with quantitative differences among events and datasets. HYCOM experiments demonstrate that the MC variability is mainly a first baroclinic mode response to surface wind forcing of the tropical Pacific, but the specific mechanism varies with latitude. The upstream part of the MC north of 7.5°N is controlled by wind forcing between 6° and 9°N through Ekman pumping, whereas its downstream part south of 7.5°N is greatly affected by equatorial winds. Prevailing westerly winds and Ekman upwelling in the developing stage cause cyclonic anomalous circulation in the northwest tropical Pacific that strengthens the MC, and the opposite surface wind forcing effect in the decaying stage weakens the MC. Although ocean models show difficulties in realistically representing the northward-flowing Mindanao Undercurrent (MUC) beneath the MC and its seasonal and interannual variations, all five products suggest an enhancement of the MUC during the decaying stage of El Niño.

1. Introduction

The Mindanao Current (MC), as the western boundary current of the North Pacific tropical gyre, is a strong southward coastal jet east of Mindanao Island. It is approximately 200 km wide and exists in the upper 600 m (e.g., Masuzawa 1969; Lukas 1988; Wijffels et al. 1995; Kashino et al. 2005; Zhang et al. 2014). The MC shows a subsurface velocity core with the maximal southward velocity of $\sim 1.3 \text{ m s}^{-1}$ at $\sim 100\text{-m}$ depth (Kashino et al. 2005), and its volume transport ranges widely from 13 to 39 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) according to estimates of various data sources and methods (Stommel and Yoshida

1972; Wijffels et al. 1995; Qu et al. 1998; Kashino et al. 2009; Schönau et al. 2015). Serving as the western boundary route of the shallow meridional overturning cell (McCreary and Lu 1994), the MC transports thermocline and intermediate water masses of the North Pacific to the equator (e.g., Bingham and Lukas 1994; Fine et al. 1994; Qu and Lindstrom 2004; Li and Wang 2012; Wang et al. 2015, 2016a) and plays a potentially important role in the heat budget of the warm pool and decadal climate variability of the tropical Pacific (e.g., Hu and Cui 1989, 1991; Hu et al. 1991; Lukas et al. 1996; Mantua et al. 1997; Gu and Philander 1997; Hu et al. 2015). In addition, the MC is also the major water source for the Indonesian throughflow (ITF) and thereby involved in the global ocean conveyor belt (Wyrki 1961; Broecker 1991; Gordon 1986; Gordon and Fine 1996; Sprintall et al. 2014). The flow beneath the MC is northward in climatology with multiple velocity cores,

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which was named the Mindanao Undercurrent (MUC) by [Hu and Cui \(1989\)](#). The MUC has a maximum velocity of $\sim 20 \text{ cm s}^{-1}$ and a total volume transport of 8–22 Sv ([Hacker et al. 1989](#); [Hu et al. 1991](#); [Lukas et al. 1991](#); [Wang and Hu 1998, 1999](#); [Qu et al. 1998](#); [Schönau and Rudnick 2017](#); [Qiu et al. 2015](#)). Recent mooring observations revealed prominent intraseasonal and semiannual variations of the MUC ([Wang et al. 2014](#); [Zhang et al. 2014](#); [Wang et al. 2016a](#)), manifesting as alternating northward and southward subthermocline flows along the Mindanao coast, providing a pathway for the intermediate water mass exchange between the South and North Pacific Oceans ([Qu and Lindstrom 2004](#); [Wang et al. 2015, 2016a](#)). Investigating the variability of the MC/MUC system on various time scales is helpful for understanding regional ocean dynamics and climate change.

Dynamics of the MC/MUC variability are intriguing owing to the complicated relationship between the two currents. Historical observations and numerical models have been utilized to understand the MC's variabilities on time scales ranging from intraseasonal to decadal (e.g., [Lukas 1988](#); [Qiu and Lukas 1996](#); [Tozuka et al. 2002](#); [Kashino et al. 2005, 2009, 2011](#); [Qu et al. 2012](#); [Zhao et al. 2012](#); [Zhang et al. 2014](#); [Wang et al. 2016b](#); [Hu et al. 2016](#); [Ren et al. 2018](#); [Duan et al. 2019a,b](#)). On interannual time scale, it has been well established that El Niño–Southern Oscillation (ENSO) is the dominant climate mode of the tropical Pacific and plays the major role in driving variability of the western Pacific circulation (e.g., [Qiu and Lukas 1996](#); [Kim et al. 2004](#); [Kashino et al. 2005, 2009, 2011](#)). Previous studies have reported the covariance between ENSO and the MC. Early time studies proposed that the MC might be involved in regulating the heat content of the warm pool and possibly creating the potential for El Niño development ([Wyrtki 1985, 1987](#)). [Lukas \(1988\)](#) found that the MC is relatively weak in the year prior to an ENSO event and stronger than average during an ENSO year, but these differences have no apparent relationship with ENSO strength. [Qiu and Lukas \(1996\)](#) demonstrated that the interannual MC is affected by both ENSO winds with a ~ 3 –7-yr period and the quasi-biennial winds confined to the tropical North Pacific region. [Kim et al. \(2004\)](#) suggested an enhanced MC transport and a northward shift of the North Equatorial Current (NEC) bifurcation latitude under El Niño condition. There were also several studies that examined individual ENSO events. They reported the MC acceleration after the onset of the 2002/03 El Niño ([Kashino et al. 2005](#)) and the stronger MC in late 2006 under El Niño conditions than in early 2008 under La Niña conditions ([Kashino et al. 2009](#)). [Hu et al. \(2016\)](#)

found that the observed interannual variability of MC is driven by wind forcing in the western Pacific Ocean through Rossby waves. Interannual variations of other currents in the northwestern tropical Pacific Ocean, such as the NEC and the North Equatorial Countercurrent (NECC), also show a close relationship to ENSO (e.g., [Wyrtki 1979](#); [Qiu and Joyce 1992](#); [Johnson et al. 2002](#); [Tozuka et al. 2002](#); [Qiu and Chen 2010](#); [Li et al. 2012](#); [Hsin and Qiu 2012](#); [Zhao et al. 2013](#); [Hu et al. 2015](#)).

In comparison, our knowledge of the MUC's interannual variability is much more fragmental due to a lack of continuous subthermocline observation. While existing research has revealed evident intraseasonal and seasonal variabilities of the MUC ([Kashino et al. 2011](#); [Zhang et al. 2014](#); [Wang et al. 2014, 2016a](#); [Ren et al. 2018](#)), few studies address interannual variability of the MUC. By analyzing mooring data, [Hu et al. \(2016\)](#) documented weak interannual fluctuations of the MUC with a typical period shorter than that of the MC. [Song et al. \(2017\)](#) suggested that the interannual variability of the MUC is closely associated with that of the subthermocline anticyclonic gyre east of Mindanao Island.

In spite of the studies reviewed above, a systematic understanding of how El Niño events modulate the MC/MUC is still lacking. Short-term measurements at a single mooring site cannot fully resolve the robust signatures of ENSO on the MC/MUC system, whereas model-based datasets were not sufficiently validated against observational data in representing the MC variability. There are several scientifically important issues to be addressed. First, although some studies have documented strong anomalies of the MC during individual ENSO events ([Lukas 1988](#); [Kashino et al. 2005, 2009](#)), the general characteristics of the MC's evolution during ENSO cycle are still unclear. Second, the response of the MUC to El Niño is unknown. Third, the dynamical processes through which ENSO modulates the MC system require in-depth understanding.

The present study attempts to provide a comprehensive investigation of the interannual variations of the MC/MUC system induced by El Niño events through addressing the scientific issues mentioned above. We first use 4-yr mooring data to quantify the year-to-year variations of the MC and MUC and verify the performance of ocean model and assimilation datasets. Then we describe the general characteristics of the MC/MUC variability using five model-based datasets and examine its robustness. At last, model experiments are performed to gain insights into the dynamical processes of the MC. The remainder of this

FIG. 1. (a) Time–depth plot of original daily mean meridional velocity V (cm s^{-1}) measured by the moored ADCPs at 8°N , 127.05°E during December 2010–August 2014. (b) The 13-month low-passed and gap-filled V anomalies (cm s^{-1} ; monthly climatology removed). (c)–(e) As in (b), but for V anomalies (cm s^{-1}) from HYCOM, SODA3.3.1, and OFES, respectively. Black curves indicate the zero contour.

paper is organized as follows. Section 2 introduces the datasets and models utilized in our analysis. Section 3 describes the interannual variations of the MC system in observation and model-based datasets. Section 4 elucidates the mechanism of the MC variability induced by El Niño events. Section 5 presents the summary and discussion.

2. Data and methods

a. Mooring data

A subsurface mooring system was deployed at 8°N , 127.05°E east of Mindanao Island to monitor the MC and MUC, providing continuous records of ~ 45 months from December 2010 through August 2014. Previous studies have described the vertical structure and intra-seasonal to interannual variations of the MC/MUC system observed by this mooring (e.g., Zhang et al. 2014; Wang et al. 2016a,b; Hu et al. 2016; Ren et al. 2018). One upward-looking and one downward-looking 75-kHz Teledyne RD Instruments (TRDI) acoustic Doppler current profiler (ADCP) were equipped on the main float of the mooring at the designed depth of ~ 400 m, aiming to simultaneously measure the currents above

900 m. More details of this subsurface mooring system can be found in Zhang et al. (2014). There were data gaps in the original daily meridional current V data (Fig. 1a) generated by the vertical fluctuations of the main float, which were filled with a linear regression method (Ren et al. 2018) and then re-sampled into monthly data. To quantify interannual variations, we removed the monthly climatology and obtained monthly V anomaly through a 13-month low-pass filter (Fig. 1b).

b. HYCOM

The Hybrid Coordinate Ocean Model (HYCOM) combines the isopycnal, sigma (terrain following), and z -level coordinates to optimize the fidelity of the ocean circulation (Bleck 2002). In this study HYCOM version 2.2.18 is configured to the tropical-to-subtropical Pacific Ocean basin from 48°S to 48°N and from 110°E to 70°W , with a horizontal resolution of $1/3^{\circ} \times 1/3^{\circ}$ and 26 hybrid vertical layers (Li et al. 2015; Li and Han 2016). Three sponge layers are applied to the western, southern and northern open-ocean boundaries, where model temperature and salinity are related to the *World Ocean Atlas 2009* (WOA09) climatology (Antonov et al. 2010;

Locarnini et al. 2010). The diffusion and mixing parameters are specified in Li et al. (2013). We use 10-m winds, 2-m air temperature, humidity, surface net shortwave and longwave radiation, and precipitation from the 0.75° European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim products (Dee et al. 2011) as the surface atmospheric forcing fields. Zonal and meridional surface wind stress, τ_x and τ_y , are calculated with 10-m wind speed $|V_{10}|$ using the standard bulk formula

$$\tau_x = \rho_a c_d |V_{10}| u_{10}, \quad \tau_y = \rho_a c_d |V_{10}| v_{10}, \quad (1)$$

where $\rho_a = 1.175 \text{ kg m}^{-3}$ is the air density, $c_d = 0.0015$ is the drag coefficient, and u_{10} and v_{10} are the zonal and meridional components of 10-m winds.

The model ocean spins up from a state of rest for 30 years under monthly climatologic atmospheric forcing. Subsequent to the spinup run, three parallel experiments are performed for the period of 1979–2016. The control run (CTL) is forced with the original daily ERA-Interim atmospheric fields and contains the complete processes including variabilities on various time scales arising from both atmospheric forcing and ocean internal origin. CTL is used as the reference solution for the validation of HYCOM performance against observational data. The other two experiments (EXP1 and EXP2) are used to understand the causes of interannual variability. EXP1 uses daily wind stress forcing as CTL, but all the other forcing fields (wind speed, radiation, precipitation, and air temperature and humidity) are fixed to monthly climatology. As such, the variability in EXP1 is predominantly induced by wind stress-driven ocean dynamical processes. According to existing studies, low-frequency variations of the MC are mainly caused by wind forcing in its latitudinal range (e.g., Qiu and Lukas 1996; Kashino et al. 2009, 2011; Qiu and Chen 2010; Ren et al. 2018), and EXP2 is designed to quantify this effect. In EXP2, daily wind stress is used between 6° and 9°N (same as CTL), and wind stress at other latitudes is fixed to monthly climatology. There are transition zones between 4°–6°N and 9°–11°N where original daily wind forcing gradually alters to monthly climatology. Same as EXP1, other forcing fields are fixed to monthly climatology. In addition, the difference between EXP1 and EXP2 (EXP1 – EXP2) can roughly measure the effect of the wind forcing beyond 6°–9°N.

c. Other datasets

Besides HYCOM, there are other model-based datasets used to estimate the MC/MUC variability. The first

is the Simple Ocean Data Assimilation version 2.2.4 (SODA2.2.4) during 1979–2010, based on the simulation of the Parallel Ocean Program (POP) version 2.0.1 (Smith et al. 1992), using the National Oceanic and Atmospheric Administration (NOAA) Twentieth Century Reanalysis (20CR) V2 fields (Compo et al. 2011) as surface atmospheric forcing and covering the global ocean with horizontal resolutions of $0.25^\circ \times 0.4^\circ$ and 40 vertical layers (Carton and Giese 2008; Giese and Ray 2011). SODA2.2.4 assimilates temperature and salinity data of the WOA09 (Locarnini et al. 2010) and sea surface temperature (SST) data from the International Comprehensive Ocean–Atmosphere Dataset (COADS) release 2.5 (Woodruff et al. 2011). The second dataset is SODA version 3.3.1 (SODA3.3.1) during 1980–2015, which is built on the Modular Ocean Model (MOM) version 5, forced by the Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2; Lee et al. 2011). SODA3.3.1 data have improved $0.25^\circ \times 0.25^\circ$ horizontal and 50-level vertical resolutions (Carton et al. 2018) and assimilate World Ocean Database 2013 (Locarnini et al. 2013), upgraded SST datasets (from COADS2.1, satellite and in situ SST, and Pathfinder SST), and satellite sea level and near-surface currents (Carton et al. 2018). The third is the eddy-resolving simulations of the Oceanic General Circulation Model for the Earth Simulator (OFES) during 1979–2015 using daily surface forcing from the National Centers for Environmental Prediction–National Centre for Atmospheric Research (NCEP–NCAR) reanalysis product (Kalnay et al. 1996). OFES has a horizontal resolution of $0.1^\circ \times 0.1^\circ$ and 54 vertical layers (Masumoto et al. 2004; Sasaki et al. 2004). The fourth is Goddard Earth Observing System Model integrated Ocean Data Assimilation System version 4 (GEOS-ODA) during 1979–2010 forced by MERRA-2 from the National Aeronautics and Space Administration Global Modeling and Assimilation Office, with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ and 40 vertical layers (Vernieres et al. 2012). Comparing with other available ocean model or reanalysis datasets, the datasets mentioned above (including HYCOM) have relatively high resolutions (at least 0.5°) and long time coverage (>30 years) so that more El Niño events can be included in the analysis. Note that these datasets are derived from different models and forced by different wind products. Therefore, a consensus of them is probably suggestive of a robust feature of the MC/MUC.

In addition to these model-based datasets, the $0.25^\circ \times 0.25^\circ$, monthly satellite sea surface height (SSH) data of 1993–2016 provided by the Archiving Validation, and Interpretation of the Satellite Oceanographic (AVISO) (Le Traon et al. 1998; Ducet et al. 2000) are also used. The 10 m, monthly zonal velocities U during 1999–2016

at 5°N, 147°E, 5°N, 156°E, and 2°N, 137°E from Tropical Atmosphere Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) are also used to validate the applicability of five model-based datasets (McPhaden et al. 1998). To examine the relationship with ENSO, the Niño-3.4 index is downloaded from the NOAA Climate Prediction Center (<http://www.cpc.ncep.noaa.gov/data/indices/>). For all the observational and model data, the monthly climatology was removed and a 13-month Hanning low-pass filter was applied to obtain the interannual anomaly.

3. Interannual variability of the MC

Figure 1a shows the daily mean meridional currents V observed by the mooring ADCPs from December 2010 to August 2014. It captures the southward MC existing in the upper 400 m (the MC layer) and the southward/northward alternating flows between 400 and 760 m (the MUC layer). We need to state that the MUC in this study refers broadly to the currents between 400 and 760 m regardless of their specific directions. The 400–760-m layer actually covers merely the very shallow portion of the MUC. Historical observations suggest that the northward flow of the MUC can reach down to ~2000 m (e.g., Qu et al. 1998; Firing et al. 2005; Qiu et al. 2015). Within the duration of measurements, year-to-year variabilities of the MC and MUC are discernible as reported by Hu et al. (2016) (Fig. 1b). The mooring observations overall suggest that year-to-year variations of the MC are likely quite different in both magnitude and timing from those of the MUC. Albeit with detailed differences, the 13-month low-passed V anomalies at the mooring location of HYCOM (Fig. 1c) are broadly consistent with mooring observation (Fig. 1b), particularly in the MC layer. Although missed some MUC anomalies, HYCOM is still able to capture the overall characteristics of the year-to-year variability of the MC/MUC system, including the major differences between the MC and MUC anomalies. By contrast, SODA3.3.1 (Fig. 1d) and OFES (Fig. 1e) show more evident discrepancies with observations. The two can reproduce most variations of MC (SODA3.3.1 and OFES perform well in October 2011–August 2014 and December 2010–August 2013, respectively) but generally fail to capture the MUC's variations. Both of them show large discrepancies from the mooring observation in the MUC layer, and OFES tends to produce in-phase variations between the MC and MUC. Therefore, all these three models have difficulties in simulating the interannual anomalies of the MUC.

For further comparison, Fig. 2 shows the V time series averaged over the MC and MUC layers at the mooring site from HYCOM CTL, SODA3.3.1, and OFES.

HYCOM produces a stronger MC ($\sim 60 \text{ cm s}^{-1}$) than mooring observation (Fig. 2a), and the mean current in the MUC layer between 400 and 760 m is southward which is opposite in direction to the weak northward MUC in mooring observations (Fig. 2d). However, it is interesting to see that the year-to-year variations of the MC observed by the mooring are faithfully reproduced by HYCOM. The southward anomalies in 2011 and 2014 and the northward anomalies in 2012 of the MC, along with the MUC's northward anomaly in 2014, are captured by HYCOM. In SODA3.3.1 and OFES, the year-to-year variations of the MC and MUC show obvious discrepancies from the observations in both the MC and MUC layers. Comparing with OFES and SODA3.3.1, HYCOM is likely more suitable for investigating the interannual variability of the MC/MUC system. Considering the effects of topography and eddy-current interaction on the subthermocline western boundary currents (Qiu et al. 2015), here we are also aware that none of the models is able to successfully simulate the mean strength and temporal variability of the MUC. Therefore, our analysis presented below is mainly focused on the MC, and results for the MUC are treated with caution.

SODA2.2.4 and GEOS-ODA are not available for the mooring observation period, and it is unknown whether the two datasets are consistent with mooring observations. To verify these model simulations of ocean currents, we compare the five model-based datasets with 10-m U anomalies from 12 TAO/TRITON buoys in the western tropical Pacific and provide their correlation coefficients in Fig. 3. The buoy sites of 8°N are located at the edge of the NEC, the 5°N buoys are at the NECC, and the 2°N, 2°S, and 5°S buoys are at the SEC. Previous studies have pointed out the NEC, NECC, and SEC show significant interannual variability and closely associated with that of the MC (Qiu and Lukas 1996; Qiu et al. 2015). The measurements of surface U anomalies are broadly consistent with the five model-based datasets, especially with HYCOM. HYCOM can better simulate the observed variability with most buoys with correlations of 0.59–0.80, except for the 5°N, 137°E and 8°N, 137°E buoys where correlations are lower than 0.50. By comparing HYCOM with four other datasets, we find that GEOS-ODA, OFES, and SODA3.3.1 are almost similar to HYCOM at the buoys near the equator with correlations larger than 0.60. The discrepancies also increase at 5° and 8°N. To a large degree, the internal ocean variability, such as the jet meanders and mesoscale eddies in the NECC region, result in current variations that are unpredictable for OGCMs (Wang et al. 2016c). The U variations during ENSO events, such as the 1998/99 La Niña and 2009/10 El Niño, are clearly

FIG. 2. The 50–400-m average V (cm s^{-1}) of (a) HYCOM, (b) SODA3.3.1, and (c) OFES at 8°N , 127.05°E compared with that from moored ADCPs. (d)–(f) As in (a)–(c), but for the 400–760-m average V . The straight lines indicate annual mean V values.

discernible in both observations and model-based datasets. In addition, since the large-scale currents in the ocean interior are primarily zonal, variability of the V component basically represents internal oceanic instabilities. This is why none of the model-based datasets can reproduce the variations of V observed by buoys (figure not shown). The correlation coefficients are generally lower than 0.40. Since signal-point current records by buoys are largely contaminated by unpredictable signals of eddies, the large area-averaged SSH is more suitable to represent the large-scale current variability during ENSO. SSH of the Mindanao Dome (MD) region (7° – 9°N , 127° – 140°E) is good indicator for the MC strength, since it largely determines the zonal SSH gradient near the Mindanao coast and thus the geostrophic flow of the MC (e.g., Lukas 1988; Wang et al. 2016b; Ren et al. 2018). SSHs of model data are closely consistent with AVISO for interannual variability with correlation coefficient of 0.95. The SSH negative peaks corresponding to the El Niño events are well reproduced by the models (figure not shown). These comparisons indicate that models are of good fidelity in representing the interannual variabilities of

the upper ocean circulation and the MC associated with ENSO.

We are not able to evaluate the effect of El Niño on the MC using mooring data owing to the absence of strong events during the observation period. Alternatively, the simulation of HYCOM agrees broadly well with mooring observation and is usable to examine the relationship with ENSO. The 13-month low-pass-filtered MC (50–400 m) is compared with the low-passed Niño-3.4 index in Fig. 4a. The MC shows significant interannual variability but differ in phase, especially during the El Niño events. The peaks and troughs of MC generally appear during strong El Niño events. The MC shows the minimal and maximal V anomalies (strong MC and weak MC) during the developing and decaying stages of the El Niño, respectively. The Niño-3.4 index and the MC show a maximum correlation of -0.51 (negative correlation indicates southward velocity anomaly occurring with positive Niño-3.4) when the MC leads by 4 months (Fig. 4b) and positive correlations when Niño-3.4 leads the MC by more than 9 months. These results indicate that the MC is strengthened during the developing stage of the El Niño (Kashino et al. 2005,

FIG. 3. The 10-m U anomalies (cm s^{-1} ; annual mean removed) at 12 buoy sites from TAO/TRITON, HYCOM, GEOS-ODA, OFES, SODA3.3.1, and SODA2.2.4. The correlations between TAO/TRITON buoys and HYCOM, HYCOM, and other four datasets are computed and shown.

2009) and is weakened during the decaying stage. Previous studies also documented consistent interannual variations in the NEC and NECC (e.g., Tozuka et al. 2002; Kashino et al. 2005, 2009; Zhao et al. 2013; Hu et al. 2015).

To examine the robustness of the interannual variability of the MC during El Niño events, we employed the other four datasets (OFES, GEOS-ODA, SODA3.3.1, and SODA2.2.4). Composite V anomalies for 10 El Niño events during 1980–2015 in the MC (0–400 m) layer and MUC (400–800 m) layer are shown in Fig. 5. Following Trenberth (1997), we select El Niño events based on the criterion of the Niño-3.4 index (SST anomaly of 5°N – 5°S , 170° – 120°W) exceeding 0.4°C for 6 months. Except for the 1982/83 and 1992/93 events, all the other events reach the peak stage in the 3 months of November–January (NDJ). There is no discernible difference between the composites organized by peak month and calendar month (figures not shown). Given that calendar month is more widely used in existing studies, here we present the results based on the calendar month in our following content. For the MC (Fig. 5a), all the five datasets show the southward V anomalies during the developing stage and northward anomalies during the decaying stage, and the ensemble-mean anomalies are -1.38 and $+1.75 \text{ cm s}^{-1}$, respectively, confirming the results of HYCOM in Fig. 4. The anomaly of GEOS-ODA is evidently weaker than the others. Although

there are evident differences among the five datasets, they reach consensus for the MC during El Niño events. It is possible that the signatures of El Niño events on the MC are rather robust and therefore captured by all the five models. Most of these composite anomalies are insignificant based on a Student's t test, owing to the small sample number (10) and large event-to-event difference. Compared with the strong seasonal and intraseasonal variabilities on an order of 10 cm s^{-1} (e.g., Zhang et al. 2014; Wang et al. 2016a; Ren et al. 2018), these interannual variations are evidently weaker.

Figure 6 shows the maxima and minima of V anomalies of the MC and MUC as functions of the lead-lag time relative to the El Niño's peak month ($T = 0$). Standard deviations of velocity anomalies and their occurring months are shown to quantify the spread of different datasets and events. The standard deviations of the MC (Fig. 6a) are as large as $\sim 3.5 \text{ cm s}^{-1}$ for V and ~ 7 months for the occurring time. Despite the large spreading, one can see that almost all of the V anomaly maxima were positive and occurred after $T = 0$. A similar distribution is seen for the minima, with most of them negative and occurring prior to $T = 0$. Therefore, the “signal-to-noise” ratio of the MC anomalies is generally close to 1. Despite the large spread of different El Niño events and different datasets, the stronger MC in the developing stage of

FIG. 4. (a) The 50–400-m V anomaly and 400–760-m V anomaly from HYCOM at 8°N , 127.05°E , compared with the Niño-3.4 index. All the variables are 13-month low-pass filtered and normalized by the standard deviation. The 11 El Niño events with Niño-3.4 $\geq 0.4^{\circ}\text{C}$ are marked by cyan bars. Lead-lag correlations of the (b) 50–400-m V anomaly and Niño-3.4 index and (c) 400–760-m V anomaly and Niño-3.4 index. Correlation values exceeding 95% significance test are plotted as blue bars.

El Niño and the weaker MC during the decaying stage are likely robust.

Next we investigate the spatial structure of the MC variations by examining the SSH and horizontal current fields in the developing stage [June–August, JJA(0)], the mature stage [NDJ(0)] and decaying stage [JJA(+1)] (Fig. 7). As noted in parentheses after the season abbreviation, 0 indicates an El Niño developing year, and +1 indicates the subsequent decaying year. By comparing the composite SSH anomalies between AVISO and HYCOM, we find that HYCOM simulations can realistically represent the SSH variations (Figs. 7a,b). In the MC layer, the anomalous low SSH and cyclonic circulation emerge over the northwestern tropical Pacific Ocean in the developing–mature stage of the El Niño. In the decaying stage, positive SSH anomalies occur at 0° – 10°N , consistent with the anomalous anticyclonic circulation (Figs. 7b,c). The variation of the MC is a component of these anomalous circulation gyres, and therefore the MC is generally strengthened in the developing–mature stage and weakened in the decaying stage, since the cyclonic circulation and anticyclonic circulation enhances and attenuates the MD, respectively (Masumoto and Yamagata 1991; Tozuka

et al. 2002). The strengthening and weakening of the MC are associated with those the NEC and the NECC, which can be discerned in Fig. 7c and is consistent with existing observations studies (e.g., Kashino et al. 2005, 2009; Zhao et al. 2013; Hu et al. 2015). The results shown in Fig. 7 overall suggest that the interannual variations of the MC during the El Niño are associated with large-scale anomalous circulation gyres. Next, we elucidate how the wind forcing of the El Niño causes anomalies of the northwest Pacific circulation and the MC variability in the following section.

4. Mechanisms

The mechanisms of the MC variability require further investigation from the perspective of wind forcing and ocean response. Previous studies have suggested that the MC is affected by local wind forcing and Rossby waves forced by remote winds (e.g., Qiu and Lukas 1996; Hu et al. 2015, 2016; Duan et al. 2019a). To examine these relationships, regression maps of Ekman pumping velocity (EPV) anomalies and zonal wind anomalies onto the normalized MC anomaly are shown in Fig. 8. EPV is computed as

FIG. 5. (a) Composite V anomalies (cm s^{-1}) of 10 El Niño events during 1979–2016 averaged over 7° – 9°N , 126.5° – 129.5°E for the 0–400-m layer from SODA3.3.1, SODA2.2.4, GEOS-ODA, OFES, and HYCOM. The thick dashed curve denotes the ensemble mean value. Cycles indicate exceeding the 95% significance Student's t test. (b) As in (a), but for 400–800-m layer. As noted in parentheses after the season abbreviation, 0 indicates an El Niño developing year, whereas 1 indicates the subsequent decaying year.

$$w_E = \frac{1}{\rho} \text{curl} \left(\frac{\tau}{f} \right), \quad (2)$$

where f is the Coriolis parameter, and $\rho = 1024 \text{ kg m}^{-3}$ is the mean density of seawater within the Ekman layer. The regression maps in Fig. 8 suggest that prior to a positive anomaly of the MC (weakening), there are easterly surface winds (negative) and Ekman downwelling (negative) in the western tropical Pacific that are likely the cause for the weakened MC. There are also westerly winds and Ekman upwelling in the central and eastern Pacific, which, however, cannot explain the weakened MC, since these wind forcing signatures act to drive upwelling Rossby wave at the MC latitudes and enhance the MC. The largest regression coefficients

FIG. 6. (a) The maxima (red) and minima (black) of the 0–400-m V anomaly (cm s^{-1}) averaged over 7° – 9°N , 126.5° – 129.5°E with respect to the lead-lag time of El Niño events; $T = 0$ denotes the peak month of an El Niño event. Results of HYCOM, SODA3.3.1, SODA2.2.4, GEOS-ODA, and OFES are plotted with different symbols. The “+” indicates the one standard deviation range. The thin dashed line denotes the standard deviation of the V anomaly. (b) As in (a), but for the maxima of the 400–800-m V anomaly.

appear in the equator, indicating possible impacts of equatorial winds to the MC variability.

We utilize the other two HYCOM experiments forced by different wind fields to explore the dynamical processes of the interannual variations of the MC system. Figure 9 compares the composite MC anomaly at the mooring site for 11 El Niño events during 1979–2016 from the three HYCOM experiments. The interannual variations of the MC from CTL and EXP1 are similar in amplitude and phase, with strong (weak) MC during the developing (decaying) stage of the El Niño, confirming the dominance of wind forcing. EXP2 can also well reproduce the CTL variations before June (+1), and after that time its anomaly is evidently weaker those of CTL and EXP1. This means that the MC variations at 8°N in the developing–mature stage were predominantly caused by wind forcing of 6° – 9°N . EXP1 – EXP2 roughly measures the wind forcing effect beyond 6° – 9°N , which has generally small effects on the MC

FIG. 7. (a) Composite SSH anomalies (m) of seven El Niño events during 1993–2016 from AVISO in (left) JJA (0), (center) NDJ (0), and (right) JJA (+1). (b) As in (a), but from HYCOM. Composite (c) 0–400-m current and (d) 400–800-m current of 11 El Niño events during 1979–2016 from HYCOM. The star marks the mooring location.

during developing and mature stages but plays a role in strengthening the MC at 8°N during the decaying stage. The correlation coefficients of CTL/EXP1, EXP1/EXP2, and EXP1/EXP1 – EXP2 are 0.87, 0.50, and 0.49, respectively.

To better understand the effects of the 6°–9°N wind forcing, we present the time–longitude plots of the SSH anomalies averaged over 6°–9°N from AVISO, EXP1, and EXP2 in Fig. 10. Negative SSH anomalies appear in the western Pacific Ocean during the El Niño events and then positive SSH anomalies emerge as westward-propagating signals from the eastern basin, as elucidated by the delayed oscillator theory of ENSO (Suarez and Schopf 1988). It is found that EXP1 can realistically simulate the observed SSH anomalies in AVISO data, albeit with detailed discrepancies in

amplitude (Figs. 10a,b). This comparison suggests that HYCOM is capable of well representing the wind-forced oceanic wave dynamics at these latitudes. Anomalies in EXP2, forced merely by the 6°–9°N winds, show similar spatial–temporal characteristics to AVISO and EXP1, but their magnitudes are weaker (Fig. 10c). Therefore, the interannual SSH variations are also contributed by wind changes outside 6°–9°N.

As shown in Fig. 7, the anomalies of the MC and MUC are associated with large-scale anomalous cyclonic/anticyclonic circulation gyres in the northwestern tropical Pacific Ocean. Here we use the vorticity of zonal current ΔU to quantify the large-scale gyre, computed as the zonal current U difference between 4°–7°N and 8°–12°N. Positive and negative ΔU anomalies indicate cyclonic

FIG. 8. Regression of surface wind fields onto the 0–400-m V anomaly: (a)–(f) EPV anomalies (color shading; m s^{-1}) and zonal wind anomalies (gray contours; m s^{-1}) leading the 0–400-m V anomaly by 10, 8, 6, 4, 2, and 0 months, respectively. Stippling indicates 95% significance for the regression coefficient of EPV. Here the V anomaly is averaged over (7° – 9°N , 126.5° – 129.5°E) and derived from HYCOM CTL.

and anticyclonic anomalous circulation over 6° – 9°N and thus southward and northward anomalies of the MC, respectively. The time–longitude plot of the ΔU anomaly in the MC layer from EXP2 is presented in Fig. 10a, representing the upper-ocean circulation variations caused by 6° – 9°N winds. Interannual variations of ΔU generally agree with those of SSH in Fig. 10c, and anomalies show evident westward-propagating features across the Pacific basin. In the western Pacific Ocean, particularly between 140°E and 180° , wave signals from the central-to-eastern Pacific Ocean are intensively modified by local wind forcing (Qiu and Lukas 1996; Qiu and Chen 2010; Duan et al. 2019a). Positive (negative) ΔU anomalies occur in the western Pacific Ocean and strengthen (weaken) the MC during the developing (decaying) stage of the El Niño. These ΔU variations can be largely explained by the basin-scale EPV at this latitude range (Fig. 11d). The positive ΔU anomaly corresponds to the Ekman upwelling over the Pacific basin associated with westerly wind anomalies (Wyrtki 1975), while the negative ΔU anomaly is produced under Ekman downwelling.

To gain further insight into the ocean responses to wind forcing, it is instructive to examine ΔU anomalies projected onto the first and second baroclinic modes (mode 1 and mode 2) based on a vertical normal mode decomposition using the climatological density profile in HYCOM (Shankar et al. 1996). The details of mode decomposition and projection are described in Ren et al. (2018). As shown in Figs. 11b and 11c, the ΔU anomaly of mode 1 (ΔU_{M1}) is close to the total anomaly. The westward propagation of ΔU_{M1} shows a phase speed of 0.21 m s^{-1} according to calculation through a Radon transformation (Kak et al. 2002) and is close to the observed phase speed of the first baroclinic Rossby waves at this latitude range (Chelton et al. 1998). In comparison, the contributions of mode 2 (ΔU_{M2}) (Fig. 11c) and higher-order modes (figures not shown) are very small. Therefore, the circulation variability in the MC layer is predominantly the ocean response to the wind forcing in the form of the first baroclinic mode Rossby waves, consistent with existing studies (e.g., Qiu and Lukas 1996; Qiu and Chen 2010; Li et al. 2012; Zhao et al. 2013).

It is discernible that SSH anomalies forced by 6° – 9°N winds (Fig. 10c) are weaker than the total wind forced

FIG. 9. Composite V anomaly (cm s^{-1}) of 11 El Niño events during 1979–2016 averaged over the 0–400-m layer derived from CTL, EXP1, EXP2, and the differences between EXP1 and EXP2 (EXP1 – EXP2). The V anomaly is at the mooring location (8°N , 127.05°E).

anomalies (Fig. 10b), and the former accounts for only 36% of the latter near the western boundary. In Fig. 12, we compare the standard deviation (STD) of SSH anomaly at different latitudes and evaluate the explained percentage of EXP1 by EXP2. The STD of EXP2 is evidently smaller than that of EXP1 at the latitudes south of 8°N (Fig. 12a). For this downstream MC region, EXP2 shows the STD around one-half of that of EXP1 and explains <40% of the EXP1 variance (Fig. 12b). The upstream MC region at 8° – 12°N is not the case, with most of the EXP1 variance explained by EXP2. Therefore, the western boundary SSH north of $\sim 8^\circ\text{N}$ is mainly controlled by 6° – 9°N winds, while that in the south is greatly affected by wind forcing of other latitudes such as equatorial winds.

Given the tight linkage between SSH and upper-ocean circulation, the mechanisms of the MC/MUC variability may also vary with latitude. In analog to Figs. 9 and 13 compares the composite V anomalies of three HYCOM experiments at different latitudes. For the MC at 9°N (Fig. 13a), EXP2 is much stronger than EXP1 in anomaly amplitude, and EXP1 – EXP2 shows opposite variations to EXP2 and EXP1. It means that the effect of winds at other latitudes is to greatly attenuate the wind forcing of 6° – 9°N on the MC variability. This effect is mainly exerted by the winds north of 9°N , as suggested by the simple model experiments of Duan et al. (2019a). Combining the results at 8°N (Fig. 9), our model experiments overall suggest the dominance of 6° – 9°N winds in the upstream MC. For the downstream MC at 7° and 6°N (Figs. 13c,d), EXP2 makes little contribution to CTL and EXP1, and EXP1 – EXP2 can

explain most of CTL variations throughout the composite event, indicating the major driving effect by winds beyond 6° – 9°N . These results extend our knowledge of the interannual variability of the MC, since in previous studies the MC anomalies are primarily attributed to the wind forcing of its latitudinal range. Here a latitude-dependent mechanism is proposed, with the more important role of equatorial winds at lower latitudes.

We further elucidate the effect of equatorial winds on the MC variability. To do so, we plot the composite 10-m wind fields, EPV anomalies, and zonal current anomalies from EXP1 – EXP2 in the MC layers in Fig. 14. A developing El Niño involves prevailing westerly equatorial winds and large-scale Ekman upwelling north of the equator. The two factors have strong effects near the equator. Under such conditions, a basin-scale anomalous cyclonic circulation is generated between 0° and 10°N in the MC layer (Fig. 14b), typical of Rossby wave response to equatorial westerly winds. This cyclonic circulation strengthens the NEC, MC, and NECC. It takes several months for the ocean adjustments in response to the wind forcing, which is the time for the Rossby waves reaching the western boundary. In addition, the short Rossby waves provoked by coastal Kelvin waves gather near the western boundary and strengthen the coastal geostrophic currents (Pedlosky 1987). Therefore, the ocean circulation anomalies reach the peak strength in the mature stage of the El Niño.

When the El Niño reaches its mature stage, the center of westerly winds shifts to the central Pacific Ocean with Ekman upwelling and easterly winds appear in the far western Pacific Ocean causing Ekman downwelling north of the equator. The easterly winds near the western boundary begin to damp the existing cyclonic circulation in the MC layer, after which an anticyclonic circulation emerges in the decaying stage. The easterly wind anomalies are reinforced in the decaying stage and dominate the entire western Pacific by JJA(+1), giving rise to the strengthened anomalous anticyclonic circulation in the low-latitude North Pacific Ocean in the MC layers, causing the weakening of the MC (Fig. 14b). We also projected the U anomalies of EXP1 – EXP2 onto the vertical normal modes. The MC anomalies are predominantly contributed by mode 1 (Fig. 15a), and mode 2 has a very limited effect (Fig. 15b). Influences of higher baroclinic modes are also weak, as suggested by the resemblance between the sum of mode 1 and mode 2 (Fig. 15c) and the total anomaly (Fig. 14b). These results reach the consensus with previous studies that low-frequency variations of the upper-ocean circulation in the western tropical Pacific are primarily the

FIG. 10. (a) Time–longitude plots of the SSH anomaly (m) averaged over 6° – 9° N from (a) AVISO, (b) EXP1, and (c) EXP2. The 11 El Niño events are marked with black lines ($\text{Niño-3.4} \geq 0.4^{\circ}\text{C}$).

1st baroclinic mode response to wind forcing (e.g., [Qiu and Lukas 1996](#)).

5. Summary and discussion

In this study, we investigate the interannual variability of the MC/MUC system during El Niño events, through analyzing the 4-yr mooring observations, five model-based datasets (HYCOM, OFES, SODA2.2.4, SODA3.3.1, and GEOS-ODA) and performing HYCOM experiments.

The mooring data during December 2010–August 2014 exhibited complicated year-to-year variations of the MC and MUC. Models can faithfully reproduce MC variations but show difficulties in representing the structure and variations of the MUC. HYCOM shows a better fidelity in simulating the observed MC variations than others. As revealed by a composite analysis, five model-based datasets reach a qualitative consensus in suggesting the stronger MC in the developing stage of El Niño and the weaker MC during the decaying stage,

FIG. 11. Time-longitude plots of the (a) 0–400-m ΔU (representing relative vorticity) anomaly (cm s^{-1}), (b) ΔU projected onto the first vertical normal mode (mode 1), (c) ΔU projected onto the second vertical normal mode (mode 2) averaged over 6° – 9°N from HYCOM EXP2, and (d) EPV anomaly (m s^{-1}) averaged over 6° – 9°N from ERA-Interim.

albeit with the large spread of different El Niño events and different datasets. HYCOM experiments demonstrate that the dominance of tropical Pacific wind forcing in driving these anomalies. Variability of the upstream MC north of 7.5°N is mainly controlled by wind forcing between 6° and 9°N through Ekman pumping, while the downstream MC south of 7.5°N is more affected by the equatorial winds. Prevailing equatorial westerly winds and off-equatorial Ekman upwelling in the developing stage cause cyclonic circulation in the northwest Pacific, which strengthens the MC. Easterly winds and Ekman downwelling emerge during the decaying stage, which damp the existing cyclonic circulation and generate anticyclonic circulation, causing the weakening of the MC. According to vertical mode decomposition, the MC variations are mainly the first baroclinic mode response to wind forcing.

Although HYCOM shows difficulties in reproducing the observed MUC structure and variability, here we present discussion of interannual variability of the MUC during ENSO. The MUC has the maximal V anomalies (strong MUC) commonly during the decaying stage of the El Niño (Fig. 4a). The Niño-3.4 leads the MUC by 3 months with a maximum correlation coefficient of 0.54, suggesting that the MUC tends to become stronger in the decaying stage (Fig. 4c). Although none of datasets can realistically simulate the observed MUC characteristics, all of them show northward V anomalies during the decaying stage of the El Niño with an ensemble-mean anomaly of $\pm 0.75 \text{ cm s}^{-1}$, indicating that the MUC gets stronger (enhanced by $\sim 10\%$ relative to its climatologic strength) than average (Fig. 5b). Four of the five datasets (except SODA3.3.1) suggest a slight weakening of the MUC (by $\sim 0.3 \text{ cm s}^{-1}$) during the developing stage. Alternatively, all the five models

FIG. 12. (a) Standard deviation (STD) of the SSH anomaly (m) averaged over 126.5°–129.5°N from HYCOM EXP1 (blue curve) and EXP2 (red curve) during 1979–2016 at different latitudes. (b) The explained percentage of EXP1 SSH anomaly by that of EXP2.

may have exaggerated the El Niño's effect on the MUC, and the composite is subjected to the common model bias. Without validation against direct observation, the results for the MUC are highly questionable. In Fig. 6b, the MUC anomaly shows a mean value of 1.9 cm s^{-1} , and a standard deviation of also $\sim 1.9 \text{ cm s}^{-1}$, occurring 5 months after $T = 0$ with a standard deviation of ~ 6 months, suggesting the robustness of the MUC's interannual variability during El Niño. The variations in the MUC layer are weaker than those in the MC

layer and characterized by an anomalous large-scale circulation gyre over the northwestern tropical Pacific Ocean (Fig. 7d). In JJA(0), the anomalous anticyclonic gyre is located near the equator, and a weaker and smaller cyclonic gyre occurs between 6° and 12°N, which induces generally southward anomalies in the vicinity of the MUC. During the mature and decaying stages, the circulation gyre gradually shifts to the north and strengthens the MUC. Same as the MC layer, the MUC during the El Niño are also associated with

FIG. 13. As in Fig. 9, but at different latitudes: (a) 9°, (b) 8°, (c) 7°, and (d) 6°N.

FIG. 14. (a) Composite 10-m wind anomalies (vectors; m s^{-1}) and EPV anomalies (color shading; m s^{-1}) of 10 El Niño events during 1979–2016 in (left) JJA (0), (center) NDJ (0), and (right) JJA (+1). (b) As in (a), but for U anomalies (cm s^{-1}) of the 0–400-m layer in EXP1 – EXP2.

large-scale anomalous circulation gyres. These analyses for the MUC are subjected to large uncertainties and described with caution.

In this study, we mainly focus on the interannual variability and mechanism of the MC and discuss the MUC variability during El Niño. Some important issues remain. The duration of the mooring observations was too short and did not capture strong El Niño events. Therefore, we are unable to justify the ENSO's effect on the MC/MUC system using mooring data. In fact, a single mooring is far from enough for sufficiently resolving the spatial structures of the MC/MUC system, considering that the MUC has multiple velocity cores (Hu and Cui 1989) and sometimes exhibits meanders (Firing et al. 2005). The present study further reveals that the mechanism controlling the MC variability varies with latitude (Fig. 13). Therefore, it appears that a mooring array covering the entire current system deployed for a long duration will fully resolve the MC/MUC responses to El Niño events and uncover more fascinating characteristics. The data derived from the array can also be used to constrain reanalysis systems to achieve more accurate dynamical diagnosis (e.g., Liu et al. 2018a,b).

The simulation of models for the subsurface ocean needs to be improved. For example, models with higher resolution are essential to better represent the subsurface eddy–current interaction which has been suggested to be important for the formation and variability

of the MUC (Qiu et al. 2015). The models are supposed to cover the Indian Ocean to simulate the effect of ITF on the western boundary flow. Moreover, the designed numerical experiments are relatively simple and may produce artificial structures of the variability due to unrealistic wind forcing distribution. More experiments should be performed to confirm our conclusions and achieve more in-depth understandings. EXP2 is designed to exclusively measure the effect of 6° – 9°N winds, by fixing winds in other areas to climatology. As pointed out by one of the reviewers, there exists large wind gradient between the daily winds between 6° and 9°N and the climatological winds outside induced by short-term weather disturbances such as typhoons, although two transition zones are applied at 4° – 6°N and 9° – 11°N . This rectification by synoptic weather disturbances and intraseasonal oscillations on oceanic interannual variability is assumed to be small in this study, which can be evaluated in the future in the particular study.

We only examined the effects of ENSO on the interannual variability in the MC/MUC. In fact, its interannual variability is complex, and ENSO is not the exclusive driver. Other processes may also affect the MC/MUC system. For example, local nonlinear processes, such as active mesoscale eddies near the Mindanao coast (e.g., Firing et al. 2005; Qu et al. 2012; Zhang et al. 2014), may rectify onto the mean flow and its low-frequency variability through the turbulent

FIG. 15. (a) Composite 0–400-m layer U anomaly (cm s^{-1}) in EXP1 – EXP2 projected onto mode 1 in (left) JJA (0), (center) NDJ (0), and (right) JJA (+1). (b),(c) As in (a), but for mode 2 and the first two modes (mode 1 plus mode 2), respectively.

Sverdrup balance (Qiu et al. 2015) and thereby contribute to the interannual variability of the MC/MUC system. Additionally, the Indian Ocean dipole can modulate the strength of the Walker circulation, and its quick demise induces a sudden collapse of anomalous zonal winds over the Pacific Ocean (Izumo et al. 2010). The Indian Ocean basin warming can also influence the change in wind forcing over the northwest Pacific Ocean (e.g., Yang et al. 2007; Xie et al. 2009). These signatures from the Indian Ocean climate might play a role in the interannual variability of the MC/MUC system.

Recent studies have indicated that El Niño events show evident diversity and can be broadly classified into eastern Pacific, central Pacific, and mixed type events (e.g., Kao and Yu 2009; Kug et al. 2009). Different types of El Niño events have been shown to exert different impacts on the western Pacific circulation (e.g., Hsin and Qiu 2012; Zhao et al. 2013; Lyu et al. 2018; Tan and Zhou 2018). They have different wind forcing characteristics and possibly affect the MC in different ways. The dynamical responses of the western Pacific circulation to ENSO diversity require in-depth investigations, which is an interesting topic for future study.

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