Seasonal Variability and Dynamics of the Pacific North Equatorial Subsurface Current

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**ABSTRACT**

The North Equatorial Subsurface Current (NESC) is a subthermocline ocean current uncovered recently in the tropical Pacific Ocean, flowing westward below the North Equatorial Countercurrent. In this study, the dynamics of the seasonal cycle of this current are studied using historical shipboard acoustic Doppler current profiler measurements and Argo absolute geostrophic currents. Both data show a westward current at the depths of 200–1000 m between 4° and 6°N, with a typical core speed of about 5 and 2 cm s⁻¹, respectively. The subsurface current originates in the eastern Pacific, with its core descending to deeper isopycnal surfaces and moving to the equator as it flows westward. The zonal velocity of the NESC shows pronounced seasonal variability, with the annual-cycle harmonics of vertical isothermal displacement and zonal velocity presenting characters of vertically propagating baroclinic Rossby waves. A simple analytical Rossby wave model is employed to simulate the propagation of the seasonal variations of the westward zonal currents successfully, which is the basis for exploring the wind forcing dynamics. The results suggest that the wind curl forcing in the central-eastern basin between 170°W and 140°W associated with the meridional movement of the intertropical convergence zone dominates the NESC seasonal variability in the western Pacific, with the winds west of 170°W and east of 140°W playing a minor role in the forcing.

1. Introduction

The advent of satellite remote sensing in the past few decades has facilitated the study of upper-ocean circulation greatly. In comparison, information about the subthermocline ocean circulation has been lacking until the construction of the Argo float arrays in the world oceans, which have provided unprecedented data coverage over the subsurface global oceans. Using the P-vector absolute geostrophic currents based on Argo profiles, Yuan et al. (2014) studied the mean North Pacific Ocean circulation and identified a westward mean undercurrent below the eastward flowing North Equatorial Countercurrent (NECC), which they named the North Equatorial Subsurface Current (NESC). Its maximum mean geostrophic velocity exceeds 2 cm s⁻¹ in the western Pacific Ocean, with a mean transport as large as 4.0 Sv (1 Sv = 10⁶ m³ s⁻¹) toward the entrance of the Indonesian Throughflow.

This subsurface current has been indicated by sporadic historical measurements. Hayes et al. (1983) calculated zonal geostrophic velocity relative to a motionless depth...
at 500 m in the 110°W meridional vertical section in the eastern Pacific, showing a subsurface narrow westward current near 7°N. The repeated hydrographic surveys along the Hawaii–Tahiti section showed a subsurface geostrophic current flowing westward beneath the NECC (Wyrtki and Kilonsky 1984). The observation of Kessler (2006) using shipboard acoustic Doppler current profiler (SADCP) indicated a westward subthermocline flow in the eastern Pacific. The subsurface westward flow also appeared in the geostrophic currents relative to a motionless depth at 1000 m in the 137°E vertical section in Qiu and Chen (2010) and Zhai et al. (2013). However, none of these previous studies have focused on the westward undercurrent. Nor were they able to distinguish this undercurrent from the perturbations of mesoscale eddies and others due to the poor spatial–temporal coverage of the historical data, until Yuan et al. (2014), who were able to estimate the mean velocity of the NESC based on the Argo data and identify the mean current. The existence of the NESC was later confirmed by Wang et al. (2016) using moored current meters at (4.7°N, 140°E). Recently, moored acoustic Doppler current profiler at the latitudes of 4°, 6°, and 8°N in the 143°E meridional section for more than a year have shown the existence of the mean westward undercurrent with a core speed greater than 5 cm s⁻¹ (Zhang et al. 2018).

Temporal variations of the NESC have not been adequately explored so far. The seasonal variability of the Pacific equatorial subsurface currents has been studied by Lukas and Firing (1985), Kessler and McCreary (1993), and Marin et al. (2010), showing that the annual reversal of the Equatorial Intermediate Current is controlled by vertical propagation of baroclinic Rossby waves. Using Argo parking depth trajectory measurements, Cravatte et al. (2012) showed that the largely one cycle per year variations of the 1000-m-depth velocity in the equatorial Pacific Ocean are associated with the downward-propagating annual Rossby waves. Due to lack of observations, studies of the off-equatorial subsurface zonal currents above the Argo parking depth of 1000 m have been lacking.

Using a linear continuously stratified ocean circulation model, Kessler and McCreary (1993) demonstrated that westward and downward propagation of annual Rossby waves are forced by annual zonal winds in the equatorial Pacific. Marin et al. (2010) found consistency of the vertical propagation of the equatorial Rossby waves in an ocean general circulation model (OGCM) with the linear theory. Kessler and McPhaden (1995) pointed out the importance of multiple baroclinic modes of the Kelvin waves and the first meridional mode equatorial Rossby waves in the intraseasonal-to-interannual variability of the equatorial Pacific Ocean circulation. The dynamics of the off-equatorial subsurface zonal currents, which are influenced by both the equatorial and off-equatorial Rossby waves, have not been studied well.

Existing studies (e.g., Meyers 1979; Qiu and Lukas 1996; Kessler 2006) used a 1.5-layer reduced gravity model to investigate the dynamics of the wind-forced low-frequency variability of the tropical Pacific Ocean circulation. In particular, Qiu et al. (2013) used a 1.5-layer ocean circulation model forced with annual cycle winds to produce alternating zonal jets in the Pacific basin. The dynamics are suggested to be forced by the triangular instability of the Rossby waves near the eastern boundary. However, the subsurface character of the off-equatorial undercurrents, with vertical phase propagation, was not simulated successfully by the 1.5-layer model.

The purpose of this study is to investigate the dynamics of seasonal variations of the NESC in the equatorial North Pacific Ocean. The rest of this paper is organized as follows. Section 2 describes the datasets and the processing methods used in this study. Section 3 presents the seasonal variations of the NESC, the dynamics of which are investigated in section 4 using a simple analytical Rossby wave model. Conclusions and discussions are contained in section 5.

2. Data and method

In this study, historical SADCP measurements, P-vector geostrophic currents based on Argo profiles, and results from ocean models were used to investigate the seasonal variability and dynamics of the NESC.

a. SADCP data

Johnson et al. (2002) processed and gridded the historical SADCP and conductivity–temperature–depth (CTD) data collected from a total of 172 mapping of the meridional sections in the tropical Pacific from 143°E to 95°W during the 1990s. The gridding was 5 m vertically and 0.2° meridionally in 10 vertical sections along 143°E, 156°E, 165°E, 180°, 170°W, 155°W, 140°W, 125°W, 110°W, and 95°W. We used three sections in the western (165°E), central (170°W), and eastern (125°W) equatorial Pacific to describe the mean structure of the NESC.

In the 170°W section, historical SADCP data are the most abundant, with data from 33 cruises at 1-h intervals and on a 10-m vertical grid provided by University of Hawaii. We gridded the SADCP data onto a 0.2° meridional grid and averaged them into four seasons, which were used to validate the Argo geostrophic currents averaged into seasonal cycles.

b. The Argo absolute geostrophic currents

The Argo monthly gridded data from January 2004 to December 2016, provided by the Scrippps Institution of
Oceanography, include salinity and temperature profiles on a 1° longitude × 1° latitude horizontal grid and in 58 vertical levels from 2.5 to 1975 m (Roemmich and Gilson 2009). Absolute geostrophic currents (AGCs) were calculated using the P-vector method (Chu 1995) based on the gridded Argo profiles. The P-vector method assumes conservation of potential density and potential vorticity and is equivalent to the β-spiral method under the Boussinesq and geostrophic approximation (Yuan et al. 2014). The AGCs were calculated between 800 and 1975 m to avoid the surface mixed layer. The geostrophic currents above 800 m were calculated based on dynamic height calculation using the AGCs at 800 m as the reference velocity.

c. Wind product

The Advanced Scatterometer (ASCAT) gridded surface wind data are monthly wind speeds on a 0.25° longitude × 0.25° latitude grid from 2007 to 2018. The wind stress was calculated using the bulk formula, with the drag coefficient \( C_d = 1.3 \times 10^{-3} \). The monthly climatological wind stress was averaged over the entire time period of 2007–18.

d. The LICOM ocean model

The LASG/IAP Climate Ocean Model (LICOM) version 1.0 developed by the Institute of Atmosphere Physics of the Chinese Academy of Sciences was used to evaluate the influence of equatorial waves on the NESC. The model has a horizontal resolution of 0.5° latitude × 0.5° longitude and 30 vertical levels of varying thickness from 12.5 to 5319.1 m. A 900-yr spinup was forced with the climatological forcing of European Centre for Medium-Range Weather Forecasts (ECMWF). A hindcast run was forced with the ECMWF wind stress and heat flux between 1990 and 2001. More detailed description of model configuration can be found in Wang and Yuan (2015).

We extracted equatorial waves by decomposing the results of the hindcast experiment. The wave decomposition method has been explained in Yuan et al. (2004). The three-dimensional pressure and zonal velocity of the OGCM were projected onto the baroclinic vertical modes, which were further projected onto the equatorial Kelvin and Rossby wave meridional modes to extract the wave coefficients. The decomposed coefficients of the equatorial Kelvin and Rossby waves of the first four baroclinic modes were used to reconstruct the zonal velocity at 5°N, which gave the impact of the equatorial waves on the NESC variability.

e. Historical hydrographic data

The temperature and salinity data at 1° longitude × 1° latitude resolution from World Ocean Atlas 2018 (WOA18; Locarnini et al. 2019; Zweng et al. 2019) were used to calculated the vertical density profiles down to the abyssal ocean of 5000 m. The ocean baroclinic modes are calculated based on the density profiles. Annual mean temperature and salinity of the WOA18 data at standard depth levels averaged between 3° and 7°N are used in this study.

f. Satellite observations

The sea surface geostrophic currents calculated from satellite altimeter sea level measurements are distributed by Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVSIO) on a 0.25° horizontal grid at daily intervals. The data from 2004 to 2016 are used to estimate surface eddy kinetic energy (EKE) along 5°N. The EKE is calculated as \( EKE = \frac{(u'^2 + v'^2)}{2} \), with the primes indicating deviations from their 33-day running means (Lyman et al. 2005).

3. Results

a. Mean currents

The mean SADCP zonal velocity in vertical sections at three longitudes cross the equatorial North Pacific have shown the South Equatorial Current (SEC) flowing to the west in the latitudes south of 4°N, and the NEC9 flowing to the east between 4° and 8°N in the upper ocean above roughly 180 m (Fig. 1). The NE9 is strong across the equatorial Pacific whereas the SEC is weaker in the western equatorial Pacific than in the east. Below the NE9 and the SEC are the NESC flowing to the west and the Northern Subsurface Countercurrent (NSCC) flowing to the east, respectively, in opposite directions to the surface currents. The core speed of the NESC is larger than 5 cm s\(^{-1}\) in the subsurface western Pacific at about 5°N. In comparison, the core speed of the NSCC is larger than 30 cm s\(^{-1}\) in the western basin. Since westward flows in the western equatorial Pacific are rare and transient, the NESC is suggested to be an important conduit of westward transportation of equatorial water masses, possibly to the entrance of the Indonesian Throughflow assuming zonal movement dictated by the planetary vorticity contours before encountering the western boundary (Fig. 1a).

The AGCs of the Argo profiles have shown similar structure of mean currents as the SADCP measurements (Figs. 1d–f), showing the eastward NECC and westward SEC in the surface layer and the westward NESC and the eastward NSCC in the subsurface. The core of the NESC flows westward between roughly \( \sigma_\theta = 26.5 \text{ kg m}^{-3} \) and \( \sigma_\theta = 26.8 \text{ kg m}^{-3} \) isopycnal surfaces, with the maximum speeds in deeper isopycnal surfaces and closer to the Equator in the west than in the east. The differences between SADCP and AGC velocity are

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FIG. 1. (a)–(c) SADCP and (d)–(f) AGC mean zonal velocity distributions (contours) in the meridional sections of (a), (d) 165°E; (b), (e) 170°W; and (c), (f) 125°W, respectively. Zonal mean velocity averaged between isopycnal surfaces 26.5σθ and 26.8σθ based on (g) SADCP and (h) Argo AGCs data. (a)–(c) The SADCP observations are made in the 1990s. (d)–(f) The AGCs are calculated using the P-vector method based on Argo salinity/temperature data of 2004–16. Westward velocities are shaded. In (a)–(f), the contour interval is 5 cm s⁻¹. A 2 cm s⁻¹ contour is added for westward currents. Three isopycnal surfaces of 26.5, 26.7, and 26.8 kg m⁻³ are drawn in thick contours. In (g) and (h), the contour interval is 2 cm s⁻¹.
primarily in the amplitudes, probably due to the limited number of SADCP observations and the different spatial resolutions of the two velocity products (0.2° vs 1°). The typical core zonal velocity of the NESC is 2 cm s\(^{-1}\) in the AGC data in the western Pacific Ocean, much smaller than the SADCP core velocity of 6 cm s\(^{-1}\). Both the SADCP measurements and the Argo AGCs indicate westward acceleration of the subsurface currents, the agreement of which on a mean undercurrent beneath the NECC across the equatorial North Pacific suggests the existence of the NESC (Figs. 1g,h).

### b. Seasonal variations

The SADCP measurements in the 170°W section have the best coverage in different seasons among all sectional data, which allows for the investigation of the seasonal variability of the equatorial currents (Fig. 2). In addition, the Argo profiles have a good coverage in the 170°W section, which can be used to study the seasonal variability of the equatorial currents. Both datasets display similar seasonal cycles: the core of the westward NESC is the strongest in February–April and disappears in August–November. In May–July, the NESC merges with the North Equatorial Current so that the two westward currents are not separated from each other.

The vertically averaged zonal AGCs between 200 and 500 m in the 165°E and 170°W sections show pronounced seasonal variability (Figs. 3a,b). The westward subsurface currents in the 165°E section peak in August and change directions to flow eastward in November through February. In comparison, the westward subsurface currents in the 170°W section peak in March and change into eastward currents in July through November. The different seasonal variability in different meridional
sections suggests wave propagation, since the currents in the 170°W section evidently lead those in the 165°E section by 3–4 months.

The westward propagation of the seasonal variations of the NESC is further evidenced by the longitude–time evolution of the NESC transport (Fig. 3d), which explains the time lag of variations between 165° and 170°E (Fig. 3c). Considering the equatorward shifting of the NESC core, we divided the equatorial Pacific basin into three sections and integrated the NESC transports in
different depth–latitude domains in the three sections: 5°–7°N, 220–500 m in the western section west of 170°W; 6°–8°N, 190–470 m in the central section between 170° and 140°W; and 6°–8°N, 150–430 m in the eastern section east of 140°W (separated by the dash vertical lines in Fig. 3). The Hovmöller diagram of the transport clearly shows that the seasonal variations of the NESC propagate to the west. Figure 3d also shows that the seasonal reversal of directions is a basin character of the NESC transport, although the annual mean transport of the NESC is still westward (Fig. 1).

Figure 3d shows that the transport variability of the NESC gets larger while propagating westward. The effects of the eastern boundary reflection are confined east of 115°W and do not propagate far away from the eastern boundary evidently, probably due to the strong mixing in the eastern basin associated with the active tropical instability waves (TIWs) and strong vertical shear. The variability is the largest in the central-western equatorial Pacific, the dynamics of which are investigated in the next section using a vertically propagating Rossby wave model.

c. Harmonics analysis

The annual harmonics are extracted using the least squares fitting of the monthly zonal AGCs along 5°N to a cosine function oscillating at the annual frequency (Fig. 4). The phase of the annual harmonics is in agreement with the energy propagation of the meridional mode 1 annual Rossby waves, with the WKB ray path extending from 200 m at 145°W to 1000 m at 170°E in the western basin (Fig. 4b). The WKB ray path is a constant phase line of the Rossby wave propagation in the linear continuously stratified ocean circulation model, providing a conduit for wind-induced energy to propagate into the deep ocean (Kessler and McCreary 1993; Ramos et al. 2008). The phase of the annual harmonics shows a clear propagating pattern, with
energy propagating downward to the west in the central and western equatorial Pacific. East of 115°W, the propagation is less obvious, probably due to the strong TIWs propagating eastward and the processes near the eastern boundary interfering with the Rossby wave propagation.

The amplitude of the annual harmonics is concentrated in the upper ocean above the thermocline, especially in the central-eastern basin (Fig. 4a). In contrast to the phase propagation, the amplitudes of the annual harmonics decay quickly in the deep ocean. Kessler and McCreary (1993) have also noted that the vertical propagation of the off-equatorial waves loses magnitudes faster than that of the waves on the equator as they leave the upper layer and propagate into deeper depths in the continuously stratified ocean model, the dynamics of which are not clear.

We speculate that these amplitude differences are primarily due to the existence of TIWs and mesoscale and submesoscale eddies that are active in the frontal areas between the warm pool and the cold tongue in the central and eastern off-equatorial Pacific Ocean. The mesoscale eddies are produced by local nonlinear dynamics and they stay in the surface layer while propagating to the west. The eddies can produce perturbations at the depths as deep as 200–300 m through thermocline heaving. TIWs may generate Rossby waves through nonlinear rectification (Yuan 2005). All these nonlinear processes are concentrated in the upper layer and do not propagate down due to the weak nonlinearity in the deep ocean. The ensemble forcing on vertical propagation is weak due to the orthogonal relation between the first and the higher baroclinic mode functions. In this study, we focus on the wind-driven dynamics of the NESC seasonal cycle, leaving the TIWs and eddy-forced variability within and immediately below the thermocline to a future study.

The annual harmonics of the vertical isothermal displacement have also been extracted based on the Argo profiles (Figs. 4c,d). The vertical isothermal displacement is defined as $\xi = T_A/(\partial T/\partial z)$, where $T_A$ is the amplitude of potential temperature annual harmonics and $\partial T/\partial z$ is the mean vertical gradient of the potential temperature (Kessler and McCreary 1993; Marin et al. 2010). This scaling variable is representative of ocean vertical motion because annual temperature variation below the thermocline is dominated by the vertical displacement of the isopycnal surfaces (Kessler and McCreary 1993).

Comparing the annual harmonics of the two variables on the same ray path, the phase of vertical isothermal displacement lags that of zonal velocity by about 3 months. This is in contrast to the harmonics on the equator, with the isotherm displacement leading the zonal velocity (Kessler and McCreary 1993). The phase differences can be explained by the off-equatorial geostrophy. The isotherm displacements immediately north and south of the NESC have different westward
propagation speeds of the Rossby waves, with the speeds off the equator smaller than on the equator, hence the equatorial phases leading the off-equatorial phases. The finite difference of the isothermal displacement in cosine functions in the meridional direction generates the zonal velocity in a sine function multiplied by a sine function of the finite difference of the southern and northern phases. The NESC zonal velocity in the sine function leads the local isotherm displacement in the cosine function by 90°. In contrast, the equatorial geostrophy has different dynamic balances, which generate different phase lags from the off-equatorial geostrophy. The wavelengths estimated by the phase variations are about 14,500 km in zonal direction and 4.1 km in vertical direction, which are comparable with previous studies (Kessler and McCreary 1993).

The amplitude of the isotherm displacement is concentrated in the central-eastern equatorial Pacific, probably due to the shoaling of the thermocline, which is subject to significant displacement associated with the warm pool zonal movement. The wavelength and phase relation of the zonal velocity and of the vertical isotherm displacement suggest the existence of vertically propagating annual Rossby waves along 5°N.

d. CEOF analysis

To extract the dominant modes of variability propagating in space and time, the complex empirical orthogonal function (CEOF; Barnett 1983) analysis has been conducted on the monthly climatological AGC zonal velocity along 5°N. We focus on the currents between 200 and 1500 m.

The principal components (PCs) and spatial patterns of the first CEOF mode accounts for 70% of total variances (Fig. 5). The real and imaginary parts of the spatial pattern (Figs. 5b,c), showing the shift after 90° of propagation, suggest westward and downward propagation of the seasonal variability (cf. Fig. 4). The westward NESC in the western equatorial Pacific peaks in winter and disappears in June through October, as indicated by the real PC time series (red line in Fig. 5a). The quadrature function in the imaginary PC shows the phase shift in time after 90° of propagation.

4. Dynamics

The dynamics of the NESC seasonal variations are investigated using a linear continuously stratified model (LCSM; see the appendix) of ocean circulation, in which the ocean stratification is based on the WOA18 climatology (Locarnini et al. 2019; Zweng et al. 2019) between 3° and 7°N and west of 110°W in the Pacific Ocean. The model is forced by the monthly climatological wind stress of ASCAT and is integrated along the Rossby wave characteristic lines from the eastern equatorial Pacific Ocean.

a. The Rossby wave model

Assuming that the vertical stratification is horizontally uniform and constant in time, the vertical eigenfunction \( \psi_n \) and the associated gravity wave phase speed \( c_n \) are
calculated from a Sturm–Liouville system (McCreary 1981), based on which the LCSM can be separated into the superposition of a number of vertical modes governed by associated shallow water equations. The vertical mode decomposition is sensitive to the spatial–temporal variability of stratification, especially the zonal thermocline slope in the tropical Pacific Ocean. We limit our calculations to the interior ocean by starting the characteristic wave line integration westward from 110°W, where the zonal velocity boundary condition is specified based on the Argo AGCs.

Within one internal deformation radius from 5°N, the Marshall Islands and the islands of Kiribati are situated roughly between 170° and 175°E and the Micronesian islands are distributed between 150° and 165°E, which form a partial barrier for the long Rossby waves to propagate westward into the western Pacific. The mixing induced by these islands is represented by a larger damping coefficient of $0.8 \times 10^{-6} \text{ m}^2 \text{ s}^{-3}$ west of 170°E, ramped up from $1.0 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$ east of 175°E, in the linear model. Without the enhanced mixing, the simulated zonal velocity in the western basin would be much larger than the observed.

The Rossby wave equation is integrated along the wave characteristic lines using phase speeds of the respective baroclinic modes. The summation of the zonal velocity of each mode gives the total response. Figure 6 shows the annual harmonics of simulated zonal velocity along 5°N using the first three or the first seven baroclinic modes, respectively. The apparent upward and westward phase propagation (Figs. 6b,d) is analogous to the observed harmonics propagation (Fig. 4b), suggesting the success of the simple model in reproducing the propagation of the seasonal variability of the subsurface currents along 5°N. The principal features have been captured by the first three baroclinic modes (Figs. 6a,b), with the responses of higher baroclinic modes adding little more (Figs. 6c,d). This is because the higher modes have smaller wave speeds and dissipate more strongly during the westward propagation.

Some differences between the observed and simulated zonal velocity are identified. The amplitudes of the simulated subsurface currents are the largest in the western-central basin whereas the observed are concentrated in the upper layer in the eastern basin. The difference is due to the analytical model unable to simulate the active instability waves and mesoscale and submesoscale eddies in the off-equatorial frontal areas near the eastern boundary of the warm pool and the western boundary of the cold tongue. These nonlinear features are generally locally generated and are not represented in the LCSM. We argue that these features...
are concentrated in the surface layer and do not propagate vertically to drive the undercurrents in deeper oceans. A piece of evidence of this argument is the successful simulation of the phase propagation of the annual harmonics in depths larger than 300 m. In fact, the differences between the simulated (using the first three baroclinic modes) and observed amplitudes are primarily in the surface layer in the eastern Pacific Ocean, with their phase differences essentially not propagating below \( \sim 200 \)-m depths (Fig. 7). The nonlinear processes (TIWs, eddies, etc.), which are active between 145° and 125°W as indicated by the high-level EKE, cannot be simulated by the linear model.

The linear model has captured the seasonal variations of the observed NESC in the subthermocline ocean below about 200 m well (Figs. 4, 6), suggesting that the seasonal variability of the NESC is primarily forced by the wind-driven dynamics. The comparisons of the annual cycle variations of the simulated and geostrophic zonal currents at the 300- and 500-m depths at 5°N, 160°E, 5°N, 180°, and 5°N, 160°W, respectively, have shown the amplitude differences less than 1 cm and phase differences less than 1 month in the western, central, and eastern Pacific (Fig. 8). These comparisons can also be identified by a visual inspection of Fig. 7. An area of slightly larger phase difference is identified between 160° and 170°E (blue color in Fig. 4b, which is absent in the isopycnal phase propagation of Fig. 4d). Further investigations suggest that this phase difference is not present if the baroclinic geostrophic zonal velocity relative to a 2000-m level of no motion should be used to calculate the observed annual harmonics. It is, therefore, suggested that this difference of phases is due to barotropic circulation induced by the island perturbations. The phase difference is only less than 1 month and can be neglected in the simulation of the Rossby wave.
propagation. The comparisons give us confidence of exploring the wind-driven dynamics of the subsurface seasonal variability using the simple model.

b. Wind curl forcing

The role of local and remote wind forcing in modulating the seasonal variations of the NESC are investigated using the linear Rossby wave model. Four model experiments are carried out to test the sensitivity of the model simulation to wind forcing over different regions. The main run (MR, Figs. 9a,e) is the solution to Eq. (A4) obtained by applying the wind stress curl forcing west of 110°W. Three sensitivity experiments are conducted to investigate the controlling dynamics of the NESC. First, the wind curl forcing west of 170°W is set to zero, which represent the absence of western Pacific winds. To compare with the MR simulation, the differences between the sensitivity experiment and the MR are displayed (Figs. 9b,f), which show small differences from the MR when the forcing west of 170°W is absent. The maximum amplitude difference is less than 0.2 cm s\(^{-1}\) in the western Pacific below 200 m and the phase differences less than 1 month.

In comparison, another sensitivity experiment with the wind curl in the region between 170° and 140°W set to zero shows glaringly large differences from the MR, especially in the west basin (Figs. 9c,g). The amplitude differences west of 160°W are larger than 1.0 cm s\(^{-1}\), with the maximum reaching 3.4 cm s\(^{-1}\), nearly the same amplitude as the MR simulation. The phase differences are larger than 1 month between 200 and 600 m. Both suggesting that the wind curl forcing in the central-eastern basin is the dominant forcing of the NESC seasonality in the western and central equatorial Pacific.
The wind curl forcing east of 140°W is also found to be weaker than the central-eastern Pacific wind curl forcing (Figs. 9d,h), with amplitude differences less than 1.0 cm s$^{-1}$, in spite of the sizable phase differences due to wave propagation.

The above analyses suggest that the wind-induced vertically propagating Rossby waves determine the seasonal variations of the NESC in the western Pacific. The simple model simulations have shown that the NESC in the western basin is mainly controlled by the wind curl in the central-eastern basin between 170° and 140°W remotely. The wind curl forcing in the western and eastern basins is less important than that the central-eastern Pacific wind curl forcing. Since we know that the wave characteristics are nearly horizontal in the thermocline and then turn downward as they get into the weaker stratification below, it is not surprising that winds west of 170°W do not produce much signal below 600 m. Those waves take at least several thousand kilometers to propagate down to that level.

c. Roles of ITCZ and equatorial waves

The seasonal variability of the wind stress curl in the central-eastern equatorial North Pacific is associated with multiple processes, including the Walker cell variations and the meridional movement of the intertropical convergence zone (ITCZ), etc. (Fig. 10). The ITCZ is a band of surface wind convergence and high precipitation, with a dipole of maximum positive and negative wind stress curl straddling the convergence. The ITCZ moves to the southernmost position in spring and back to the northernmost position in autumn. Seasonal variations of the ITCZ is believed to be forced by the solar radiation and modulated by the tropical air–sea coupling, which change the surface wind pattern and induce the wind stress curl variability in the central and eastern Pacific. The Walker cell circulation is associated with the warm pool–cold tongue structure of the tropical sea surface temperature and is included in the ITCZ wind system in the eastern and central-eastern Pacific. In addition, other processes like the regional ocean–atmosphere coupled variabilities associated with the TIWs may also produce wind curl variations (Chelton et al. 2001). The NESC seasonality in the western Pacific is found to be dominated by the seasonal variations of the ITCZ through vertical Rossby wave propagation.

To prove the above hypothesis, an idealized wind patch, including only zonal wind stress component, with a Gaussian meridional distribution and an annual variation as denoted by Kessler and McCreary (1993), is used to represent the annual variations of the ITCZ in the central and eastern equatorial Pacific Ocean. The control and three sensitivity experiments are repeated, using the idealized wind forcing. The seasonal variations of the NESC are similar to those forced with realistic winds and are most sensitive to the seasonal cycle of the ITCZ wind stress curl in the central-eastern Pacific (Fig. 11). Compared with the small amplitudes using realistic wind curl forcing over the western Pacific, larger NESC annual harmonics are generated by the idealized wind forcing over the western Pacific (cf. Figs. 9b,f with Figs. 11b,f). This is because the idealized winds are based

FIG. 10. The spatial distribution of monthly climatological wind speed (vectors; m s$^{-1}$) and wind stress curl (colors; 10$^{-6}$ N m$^{-2}$) based on the ASCAT wind product. The wind vector scale is at the top-right corner of (a).
on zonally averaged amplitudes in the eastern Pacific, which overestimate the wind amplitudes in the western Pacific (Yuan 2005). The errors basically propagate westward and downward following the Rossby wave rays and are concentrated in the far western Pacific west of 150°E.

In spite of the above errors, the idealized zonal wind is actually similar to the first EOF (90% of the total variance) and first CEOF (93% of the total variance) modes of the realistic zonal wind in the central-eastern Pacific (Fig. 12). The first EOF and the first CEOF modes of the meridional winds takes 90% and 91% of the total meridional wind variances, respectively (not shown). These dominant modes produce the ITCZ meridional movement when superposed onto the annual mean winds, and are, therefore, suggested to be associated with the ITCZ seasonality. Further experiments suggest that the NESC seasonal variations in the western Pacific are dominated by the zonal wind forcing (not shown). The dominance of the first EOF and first CEOF modes of the zonal winds suggests that the higher-mode wind forcing is secondary compared to the ITCZ seasonal zonal wind forcing on the NESC seasonal variations in the subsurface western Pacific.

The zonal velocity variations along 5°N are also influenced by the equatorial waves. To estimate the NESC variations induced by the equatorial Kelvin and Rossby wave, equatorial wave decomposition is first conducted on a LICOM simulation of the equatorial Pacific Ocean circulation using the method of Yuan et al. (2004). The Kelvin waves and the first meridional-mode equatorial Rossby waves of the first four baroclinic modes are extracted and are used to reconstruct the zonal velocity along 5°N. Annual harmonics of the reconstructed zonal velocity of the Rossby waves also exhibit the vertically propagating pattern (Figs. 13a,b). However, the magnitudes are much smaller than that of the MR forced by the wind curl (Figs. 13c,d; cf. Figs. 6a,b). The Kelvin wave zonal velocity along 5°N is even smaller, with the maximum around one-half of the Rossby wave zonal velocity (not shown). The results suggest that the off-equatorial Rossby waves forced by the wind curl variations associated with the meridional

![Fig. 11. As in Fig. 9, but for the idealized wind forcing.](image-url)
movement of the ITCZ dominate the NESC seasonal variations in the western Pacific Ocean.

5. Discussions and conclusions

The mean state and seasonal variations of the NESC are studied based on historical SADCP and Argo AGC data, showing a westward mean current beneath the wind-driven, eastward-flowing NECC. The mean NESC originates in the eastern Pacific Ocean east of 90°W with a typical core velocity larger 2 cm s\(^{-1}\) and flows westward and equatorward while descending to deeper isopycnal surfaces. Significant seasonal variations and their propagation are identified. The annual harmonics and CEOF analyses suggest that the seasonal variability of the NESC is associated with vertically propagating Rossby waves, forced by the equatorial Pacific wind.

A simple linear Rossby characteristic wave model is used to study the dynamics of the seasonal variability of the NESC. Numerical experiments suggest that wind stress curl over the central-eastern equatorial Pacific Ocean is the main driving force of the NESC seasonal variability in the western Pacific. In particular, the wind stress curl variations induced by the meridional movement of the ITCZ force strong seasonal variations in the central equatorial Pacific Ocean, which propagate westward and downward into the deep western Pacific Ocean. The dynamics are through the vertical propagation of off-equatorial Rossby waves associated with low baroclinic modes of the stratified ocean. The equatorial Rossby and Kelvin waves are found to make much smaller contribution to the seasonal variations of the NESC than the off-equatorial Rossby waves.
Kessler (2002) found that half of the eastward NSCC turns northward and westward near Costa Rica Dome. This westward flow could be the source of the NESC. At that time, the mean NESC was not identified. The connection of the two subsurface currents need further study. Due to sparse Argo profile data near the Pacific western boundary, the destination of the NESC was not determined at present. The NESC could also exist in other oceans, like the North Atlantic Ocean (Schott et al. 2003). The variability in zonal velocity and temperature below the equatorial thermocline influenced by vertically propagating Rossby waves has been studied before (e.g., Thierry et al. 2006; Johnson 2011; Ishizaki et al. 2014; Nagura 2018). But the reconstruction of the off-equatorial zonal currents at 5°N has been conducted for the first time, which shows success in simulating the main characters of the seasonal variability. Internal nonlinear oceanic processes such as eddies and TIWs are likely to affect subsurface variations in the western Pacific Ocean. Their dynamics have not been considered here. Future studies should seek answers to these issues.

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APPENDIX

The Rossby Wave Model

Following McCreary (1981), the momentum and continuity equations for vertical baroclinic modes are

\[
-f v_n + p_{nx} = F_n, \quad (A1)
\]

\[
f u_n + p_{ny} = G_n, \quad (A2)
\]

\[
p_m + c_n^2 (u_{nx} + v_{ny}) = 0, \quad (A3)
\]

where \( F_n \) and \( G_n \) measure how well each mode couples to the forcing wind. Where \( (u_n, v_n) \) are \((x, y)\) components of horizontal velocity, \((\tau^x, \tau^y)\) are \((x, y)\) components of wind stress vector \(\tau\). Other terms are as follows: \(f\) is the Coriolis parameter; \(\psi_n(z)\) is the vertical structure function of mode.
where long Rossby wave speed \( c = -\beta c_n^2/f^2 \). \( \beta \) is the meridional derivative of the Coriolis parameter \( f \). The damping coefficient is formulated as \( \varepsilon = A/c_n^2 \), where \( A = 1 \times 10^{-7} \text{m}^2\text{s}^{-3} \) east of 175\(^\circ\)E and \( A = 8 \times 10^{-7} \text{m}^2\text{s}^{-3} \) west of 170\(^\circ\)E to simulate the island effects, with exponential ramping between 175\(^\circ\) and 170\(^\circ\)E for a smooth transition. The meridional derivative of Eq. (A4) multiplied by \(-1/(\rho f)\) gives the equation about the zonal velocity \( u_n(x, t) \):

\[
 u_m + c u_n + \varepsilon u_n = \left[ \frac{c_n^2}{\rho f} \right] \left[ \text{Curl} \left( \frac{1}{f} \int_{-D}^{0} \psi_n^2 \, dz \right) \right]_y .
\]

The solution \( u_n(x, t) \) can be written in integration along wave characteristics that start from the eastern boundary \( x_E \):

\[
 u_n(x, t) = u_{nE} \left( x_E, t - \frac{x - x_E}{c} \right) e^{-\left( \varepsilon/c \right)(x-x_E)}
 + \frac{c_n^2}{\rho f} \int_{x_E}^{x} e^{-\left( \varepsilon/c \right)(x-x')} \left\{ \text{Curl} \left[ \tau' / f \int_{-D}^{0} \psi_n^2 \, dz \right] \right\} \, dx'.
\]

The first term on the right-hand side \( u_{nE} \) is the value of mode-\( n \) zonal velocity at the eastern boundary. The zonal velocity field \( u \) equals to the summation of the first \( N \) baroclinic modes,

\[
 u = \sum_{n=1}^{N} u_n \psi_n(z).
\]

REFERENCES


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