Features of the Equatorial Intermediate Current Associated with Basin Resonance in the Indian Ocean[✉]

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(Manuscript received 12 November 2017, in final form 11 April 2018)

ABSTRACT

This paper investigates the features of the Equatorial Intermediate Current (EIC) in the Indian Ocean and its relationship with basin resonance at the semiannual time scale by using in situ observations, reanalysis output, and a continuously stratified linear ocean model (LOM). The observational results show that the EIC is characterized by prominent semiannual variations with velocity reversals and westward phase propagation and that it is strongly influenced by the pronounced second baroclinic mode structure but with identifiable vertical phase propagation. Similar behavior is found in the reanalysis data and LOM results. The simulation of wind-driven equatorial wave dynamics in the LOM reveals that the observed variability of the EIC can be largely explained by the equatorial basin resonance at the semiannual period, when the second baroclinic Rossby wave reflected from the eastern boundary intensifies the directly forced equatorial Kelvin and Rossby waves in the basin interior. The sum of the first 10 modes can reproduce the main features of the EIC. Among these modes, the resonant second baroclinic mode makes the largest contribution, which dominates the vertical structure, semiannual cycle, and westward phase propagation of the EIC. The other 9 modes, however, are also important, and the superposition of the first 10 modes produces downward energy propagation in the equatorial Indian Ocean.

1. Introduction

The Equatorial Intermediate Current (EIC), defined as the zonal flow beneath the thermocline generally down to 1200-m depth, is well observed in the equatorial Pacific, Atlantic, and Indian Oceans (Firing 1987; Delcroix and Henin 1988; Jensen 1993; Kessler and McCreary 1993; Fischer and Schott 1997; Gouriou et al. 2006; Wang et al. 2016). The EIC in the Pacific and Atlantic Oceans shows a dominant annual cycle and appears to play an important role in modulating the equatorial and extra-equatorial exchanges of mass and heat (Brandt and Eden 2005; Marin et al. 2010). Differently, the EIC in the Indian Ocean shows a prominent semiannual cycle with seasonal

velocity reversal (Luyten and Roemmich 1982; Jensen 1993). It is well established that such a semiannual cycle in the equatorial Indian Ocean is mainly generated by the variable surface wind fields (Wyrtki 1973; Knox 1974; Wunsch 1977; Yamagata et al. 1996; Han et al. 1999; Yu et al. 2005; Ogata and Xie 2011; Duan et al. 2016) and imprinted deeply into the near-surface zonal currents (e.g., Han et al. 1999), Equatorial Undercurrent (EUC; e.g., Chen et al. 2015, 2016) and EIC. Investigating the mechanisms of the wind-driven response of the EIC in the Indian Ocean is both scientifically and climatically important, since the causes for the Indian Ocean EIC are not well known, and this investigation will help us to understand how and why the equatorial intermediate Indian Ocean responds to the global climate variability and change (Song and Colberg 2011; Balmaseda et al. 2013a).

Wind-driven Kelvin and Rossby waves and Rossby waves reflected from the eastern ocean boundary are

[✉] Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JPO-D-17-0238.s1>.

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observed to be important in causing the semiannual cycle of the surface and subsurface currents in the equatorial Indian Ocean (Wyrtki 1973; Anderson and Carrington 1993; Schott et al. 1997; Reppin et al. 1999; Iskandar et al. 2009; Chen et al. 2015; Nagura and McPhaden 2016). In comparison, our current understanding of the variability of the EIC derives mainly from sporadic observations and theoretical work. Based on the wavelength obtained from current-meter measurements in the western Indian Ocean, Luyten and Roemmich (1982) adopted a linear wave theory to explain the semiannual variability of the zonal velocity at 750 m and concluded that the mechanism is the propagation of a first meridional mode Rossby wave. O'Neill and Luyten (1984) analyzed the acoustic dropsonde measurements in the western Indian Ocean and found that the structure of the zonal current in the intermediate layer can be identified by the linear equatorial waves. Jensen (1993) applied a numerical isopycnal ocean model to study the equatorial variability of the Indian Ocean and demonstrated that the reversals of the EIC can be reproduced in a 3.5-layer reduced-gravity model. Despite these theoretical studies, the generation of the EIC remains unclear and has not yet been dynamically investigated.

An idea of equatorial energy rays has been suggested to explain the variability of the EIC in the Pacific and Atlantic Oceans (Kessler and McCreary 1993; Dewitte and Reverdin 2000; Thierry et al. 2004; Brandt and Eden 2005; Marin et al. 2010; Ishizaki et al. 2014). These equatorial rays can conceptually describe the westward and downward energy propagation and are calculated analytically as the sum of the low-order baroclinic mode equatorial Kelvin and Rossby waves (Gent 1981; McCreary 1984). In the equatorial Indian Ocean, however, such rays may be disturbed by the enhanced second baroclinic mode response because of basin resonance at the semiannual period. Using idealized and realistic simulations of the Indian Ocean, Jensen (1993), Han et al. (1999), and Ogata and Xie (2011) found that the propagation of the second baroclinic mode equatorial Kelvin and Rossby waves is in resonance with the semiannual wind forcing. Han et al. (2011) used a hierarchy of ocean models and found that the existence of basin resonance at the semiannual period is a robust feature in all models of different complexity, and the time that it takes to establish such basin resonance is approximately 180 days in the equatorial Indian Ocean. Subsequent studies of this basin resonance in the Indian Ocean have highlighted its influence on the seasonal variability of the EUC and sea surface height (Fu 2007; Chen et al. 2015; Nagura and McPhaden 2016; Cheng et al. 2017). In this paper, we report the variability of the EIC in the Indian Ocean and explore its causes.

An enhanced equatorial observing system, consisting of subthermocline acoustic Doppler current profiler (ADCP) moorings deployed in an equatorial array at different longitudes (80°, 85°, and 93°E), has to date been maintained for two years in the Indian Ocean. Here, we use these moored measurements to analyze the spatial and temporal structure of the Indian Ocean EIC. The remainder of the paper is organized as follows: The data, a continuously stratified linear ocean model (LOM), and equatorial resonance theory are introduced in section 2. In section 3, a comparison of the modeled and observed data is first presented, to demonstrate the suitability of the oceanic reanalysis and LOM used in this study. Second, the observed horizontal and vertical structures of the subthermocline zonal current are described, to demonstrate the spatial and temporal variability of the EIC. And third, the dynamics of the wind-driven wave response of the EIC are investigated. In particular, the association of the EIC with the equatorial basin resonance of the second baroclinic mode and contributions from other modes are presented. Finally, the main results are summarized and discussed, along with conclusions drawn, in section 4.

2. Data, linear ocean model, and equatorial resonance theory

a. Data

Three moorings (Q2, Q4, and Q5), each containing an upward-looking 75-kHz ADCP, were deployed by the South China Sea Institute of Oceanology (SCSIO), Chinese Academy of Sciences, to observe the EIC in the Indian Ocean (Fig. 1a). Q2 and Q4 were deployed at (0°, 93°E) and (0°, 85°E), respectively, and provided data from March 2015 to March 2017. Q5 was deployed at (0°, 80°E) in April 2015 and retrieved in March 2017. The ADCPs recorded the velocity profiles over a depth range of 50–500 m with a vertical bin size of 8 m. The velocity data were interpolated vertically to a standard 5-m depth interval and then averaged to derive the daily mean. To remove the tidal effect, a central three-day running mean was applied to the daily data. The current measurements obtained from the mooring of the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA; McPhaden et al. 2009) were analyzed to verify our model output. The mooring was deployed at 0°, 90°E from November 2000 to June 2012.

Monthly current data from the European Centre for Medium-Range Weather Forecasts Ocean Reanalysis System 4 (ORAS4; Mogensen et al. 2012; Balmaseda et al. 2013b) were used to analyze the 3D structures of

FIG. 1. (a) Locations of the ADCP mooring sites along the equator. The mooring at the yellow square was deployed by RAMA and those at the red stars were deployed by the SCSIO, Chinese Academy of Sciences. The background coloring indicates the bathymetry. (b) Temporal history of ADCP sampling, LOM, and ORAS4 along the equator. The time series of the SCSIO moorings are from April 2015 to March 2017, for two years.

the subthermocline zonal current in the Indian Ocean. ORAS4 uses version 3.0 of the Nucleus for European Modelling of the Ocean (NEMO V3.0) ocean model. The model relaxes weakly (20-yr time scale) to climatological temperature and salinity from the *World Ocean Atlas 2005* (WOA05) (Locarnini et al. 2006; Antonov et al. 2006). The ORAS4 is constructed by assimilating various historical observations into the ocean model with a horizontal resolution of $1^\circ \times 1^\circ$ with equatorial refinement (0.3°) and 42 vertical levels. The data assimilations include temperature, salinity, and along-track satellite-derived sea surface height anomaly and no assimilation of velocity observations. The temperature and salinity profiles are obtained from expendable bathythermographs (XBTs), conductivity–temperature–depth (CTD) sensors, RAMA moorings, and autonomous pinniped bathythermographs (APBs). The sea surface height anomaly data are obtained from the Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVISO). The timeline of ORAS4 assimilation datasets has been summarized in Nyadjro and McPhaden (2014). In this paper, the time series in the date range from 2001 to 2017 were selected to match the time period of moored observations. We also compared the selected time series to the whole dataset (1958–2017), and they gave similar results (see the supplemental material).

b. Model and experiments

A continuously stratified LOM was used to assess the relative importance of the equatorial baroclinic modes for the structure and propagation of the EIC. This model is described in detail in McCreary (1981) and has been applied to explain the dynamics of the East Indian Coastal Current, the EUC, western boundary reflection, and basin resonances (McCreary et al. 1996;

Shankar et al. 1996; Yuan and Han 2006; Han et al. 2011; Chen et al. 2015). It has been demonstrated that the LOM is able to reasonably simulate the observed zonal current variability at intraseasonal, seasonal, and interannual time scales in the equatorial Indian Ocean (e.g., Han 2005; Han et al. 2011; Chen et al. 2015), suggesting that the linear ocean dynamics dominate equatorial current variability. Consequently, the LOM can be used to investigate the dynamics of the EIC in the Indian Ocean.

In the LOM, the equations of motion are linearized about a background state of rest with a realistic stratification represented by Brünt–Väisälä frequency, and the ocean bottom is assumed flat at 4000 m. With these restrictions, the solutions can be represented as expansions in the vertical normal modes n of the system. The zonal velocity u , meridional velocity v , and pressure p in the solutions can be represented as expansions in the vertical normal modes of the system with eigenfunctions $\psi_n(z)$:

$$u = \sum_{n=0}^N u_n \psi_n, \quad (1a)$$

$$v = \sum_{n=0}^N v_n \psi_n, \quad \text{and} \quad (1b)$$

$$p = \sum_{n=0}^N p_n \psi_n, \quad (1c)$$

where the expansion coefficients u_n , v_n , and p_n are functions only of x , y , and t . Strictly speaking, the total mode number should extend to infinity, but the solutions converge rapidly enough with n (McCreary et al. 1996; Shankar et al. 1996). Herein $n = 25$ is selected to represent the total baroclinic mode number (as in Han et al. 2011; Chen et al. 2015). The total solution is the sum of

all the selected modes. The terms u_n , v_n , and p_n are governed by the following equations:

$$\left(\partial_t + \frac{A}{c_n^2}\right)u_n - fv_n + \frac{1}{\bar{\rho}}p_{nx} = \tau^x Z_n / (\bar{\rho}H_n) + v_h \nabla^2 u_n - \delta u_n, \quad (2a)$$

$$\left(\partial_t + \frac{A}{c_n^2}\right)v_n + fu_n + \frac{1}{\bar{\rho}}p_{ny} = \tau^y Z_n / (\bar{\rho}H_n) + v_h \nabla^2 v_n, \quad \text{and} \quad (2b)$$

$$\left(\partial_t + \frac{A}{c_n^2}\right)\frac{p_n}{\bar{\rho}c_n^2} + u_{nx} + v_{ny} = 0, \quad (2c)$$

where c_n is the characteristic speed of equatorial Kelvin wave for vertical mode number n . The c_n values for the first 10 baroclinic modes ($n = 1, 2, \dots, 10$), estimated from a mean background stratification based on density observations in the Indian Ocean (see [Moore and McCreary 1990](#)), are 264.2, 166.9, 104.5, 74.7, 59.7, 49.3, 41.5, 36.8, 32.8, and 29.7 cm s^{-1} . The Coriolis parameter is $f = \beta y$ under equatorial β -plane approximation, and v_h is the coefficient of the horizontal eddy viscosity. The coupling intensity of each mode to the wind field is determined by $Z_n = \int_{-D}^0 Z(z) \psi_n dz / H_n$, where $H_n = \int_{-D}^0 \psi_n^2 dz$; $Z(z)$ is the vertical profile of wind that is introduced as a body force, where $Z(z)$ is constant in the upper 50 m and linearly decreases to zero from 50- to 100-m depth. Density $\rho = 1 \text{ g cm}^{-3}$ is a typical density value of seawater; τ^x and τ^y are the surface zonal and meridional wind stress, respectively. The terms associated with A/c_n^2 represent vertical friction with $A = 0.00013 \text{ cm}^2 \text{ s}^{-3}$, and they provide damping for the equatorial Kelvin and Rossby waves. Since the damping is inversely proportional to c_n^2 , the low-order baroclinic modes Kelvin and Rossby waves experience weak damping effects because of their faster speeds (e.g., $c_1 = 264.2 \text{ cm s}^{-1}$ and $c_2 = 166.9 \text{ cm s}^{-1}$), and thus they can propagate far away from the forcing region. By contrast, the higher-order modes experience strong damping effects because of their slower speeds, and thus their response is local and mainly restricted to the forcing region [see [Han \(2005\)](#) for detailed discussion]. All solutions of the control run have a damper with coefficient $\delta(x, y)$ near the eastern boundary of the basin [in the last term of Eq. (2a)], as discussed next, to absorb the energy of forced equatorial Kelvin wave in this region.

The LOM with a realistic Indian Ocean basin without the Maldives was first spun up for 20 years forced with monthly mean climatology of cross-calibrated multiplatform (CCMP) satellite winds ([Atlas et al. 2008](#)) for the 1988–2011 period. Restarting from the spinup run, the LOM was integrated forward in time using monthly CCMP winds from 1988 to 2011. This solution is referred to as the LOM main run (LOM-MR), and the total

solution of LOM-MR is the sum of the first 25 modes. A second run was performed with a damper in the eastern equatorial ocean to isolate the effects of eastern boundary reflected Rossby waves. This damper is nonzero only in the eastern equatorial ocean [damping term with coefficient δ in Eq. (2a)] and it efficiently absorbs the energy of incoming equatorial Kelvin waves, and thus no Rossby waves are reflected back into the ocean interior from the eastern boundary (e.g., [McCreary 1981](#); [McCreary et al. 1996](#); [Han 2005](#); [Han et al. 2011](#)). We refer to this run as LOM-DAMP. The difference between the two experiments [LOM-MR minus LOM-DAMP (LOM-REFLECT)] isolates the reflected Rossby wave effects. The detailed configurations of the LOM can be further found in [Han \(2005\)](#), [Han et al. \(2011\)](#), and [Chen et al. \(2015\)](#).

c. Equatorial semiannual resonance

Previous studies have investigated how Kelvin and Rossby waves in the equatorial Indian Ocean interact to form basin resonances at the semiannual and 90-day periods (e.g., [Jensen 1993](#); [Han 2005](#); [Han et al. 2011](#)). Here, we only introduce aspects that are essential for our discussion. Harmonic analysis of the equatorial wind stress used to force the LOM (not shown here) reveals that the amplitudes of the semiannual and annual components are comparable in the central basin (3°S – 3°N , 65° – 80°E), and the ratio of semiannual component to annual component is nearly 1.0, similar to [Jensen \(1993\)](#) and [Ogata and Xie \(2011\)](#). However, observations show particularly stronger semiannual variability in near-surface zonal currents, EUC, and EIC (e.g., [Wyrtki 1973](#); [O'Brien and Hurlburt 1974](#); [Knox 1974](#); [McPhaden 1982](#); [Luyten and Roemmich 1982](#); [Gent et al. 1983](#); [Molinari et al. 1990](#); [Anderson and Carrington 1993](#); [Donguy and Meyers 1995](#); [Han et al. 1999](#); [Reppin et al. 1999](#); [Chen et al. 2015](#)). While the strong semiannual variability of zonal surface current and EUC are attributed to the resonant forcing by the semiannual winds (e.g., [Jensen 1993](#); [Han et al. 1999](#); [Chen et al. 2015](#)), causes for the EIC variability remain elusive.

Variations of the zonal equatorial winds can excite equatorial Kelvin and Rossby waves. An incident Kelvin wave can transfer part of its energy to a westward-traveling Rossby wave upon reflection at an eastern boundary. For a single baroclinic mode, resonance can occur ([Cane and Sarachik 1981](#)) and a basin mode ([Cane and Moore 1981](#)) can be set up when

$$T = 4L/mc_n, \quad m = 1, 2, 3, \dots, \quad (3)$$

where T is the forcing period, L is the width of the equatorial basin, c_n is the Kelvin wave speed for a given vertical mode n , and m is a positive integer. For $m = 1$,

FIG. 2. Comparison of the whole time series and mean seasonal cycle of zonal currents (cm s^{-1}) obtained from RAMA (red), ORAS4 (blue), and LOM-MR (green) at $0^\circ, 90^\circ\text{E}$ at a depth of (a), (b) 250, (c), (d) 350, (e), (f) 500, and (g), (h) 1000 m. The zero-lag correlation coefficients r between RAMA and ORAS4 and between RAMA and LOM-MR are shown in blue and green, respectively, in (a) and (c). The zero-lag correlation coefficients between ORAS4 and LOM-MR are shown in green in (e) and (g).

T equals the time for a Kelvin wave to cross the basin (L/c_n) plus the time for the first meridional mode Rossby wave to return ($3L/c_n$). For the equatorial Indian Ocean, the approximate width of this basin is $L = 6600\text{ km}$, and the second baroclinic mode ($n = 2$) speed is $c_2 = 166.9\text{ cm s}^{-1}$, Eq. (3) is satisfied by the c_2 and L for $T = 180$ days when $m = 1$. This is the approximate time for the second baroclinic mode equatorial Kelvin wave to cross the basin (~ 1.5 months) and the first meridional mode Rossby wave to return (~ 4.5 months). Hence, the basin geometry of the equatorial Indian Ocean allows resonance to occur for the second baroclinic mode at the semiannual period. As we shall see below, this resonance can explain the enhanced oceanic semiannual response of EIC in the observations and the LOM.

3. Results

a. Model/data comparison

Existing studies have assessed the ability of ORAS4 and LOMs to reproduce the near-surface zonal currents in the eastern equatorial Indian Ocean by comparing them with RAMA and satellite data (Nyadjro and McPhaden 2014; Chen et al. 2015; Nagura and McPhaden 2016). Herein, we further examine the ability

of the ORAS4 product and LOM results to reproduce the EIC in the Indian Ocean. Figure 2 compares the zonal velocity beneath the pycnocline for RAMA, ORAS4, and LOM-MR at the RAMA location ($0^\circ, 90^\circ\text{E}$) from 2000 to 2011. ORAS4 matches well with RAMA, with a correlation coefficient of 0.81 at 250 m (Fig. 2a). The standard deviations (STDs) of RAMA and ORAS4 are 12.2 and 10.2 cm s^{-1} , respectively, with a root-mean-square error (RMSE) of the reanalysis relative to the observation of 7.1 cm s^{-1} . At 350-m depth, the ORAS4 zonal current still shows high correlation with RAMA, and the correlation coefficient reaches 0.72 (Fig. 2c). The STDs of RAMA and ORAS4 are 10.4 and 7.88 cm s^{-1} , respectively, and the RMSE is 7.17 cm s^{-1} . LOM-MR can also reasonably reproduce the variability of the observed EIC, with correlation coefficients of 0.70 at 250 m and 0.57 at 350 m. The correlation coefficient between LOM-MR and ORAS4 is 0.75 at 500-m depth and 0.60 at 1000 m (Figs. 2e,g). Seasonal cycles of zonal velocity from RAMA, ORAS4, and LOM-MR exhibit semiannual cycles with comparable amplitudes at different depths (Figs. 2b,d,f,h). These good agreements show that LOM-MR can be used to help demonstrate the role played by wind-driven equatorial waves in causing the EIC's variability, similar

FIG. 3. Comparison of the time–depth variations of the monthly zonal currents (cm s^{-1}) from April 2015 to March 2017 derived from (a),(b) the moored ADCPs and (c),(d) ORAS4, along the equator, at (a),(c) (0° , 93°E) and (b),(d) (0° , 80°E).

to the studies of [Chen et al. \(2015\)](#) on the EUC's variability.

Depth–time plots of the monthly mean equatorial zonal currents at 93° (Q2) and 80°E (Q5) show similar properties as the mooring and ORAS4 data ([Fig. 3](#)), with strong semiannual variability dominating the total current over the entire water column—in particular, in the upper 500 m, where current amplitudes are large. As can be seen from the moored and reanalysis data, between 50 and 200 m, the EUC flows eastward with velocities greater than 50 cm s^{-1} from March to May and September to November. Below $\sim 200 \text{ m}$, the westward velocity prevails in the subthermocline, with magnitudes greater than 10 or 20 cm s^{-1} , respectively, during the same periods. The structures of the baroclinic modes can be clearly seen at 80° and 93°E , from both the mooring observations and ORAS4 data, and the second baroclinic mode structure can be visually identified at 80°E in ORAS4, with two zero crossings occurring near 170 and 1300 m ([Fig. 3d](#); derived from the first mode of the empirical orthogonal function decomposition of ORAS4's zonal velocity—not shown).

b. Observed horizontal structure of the EIC

[Figures 4a–c](#) show the time–depth variations of the daily zonal currents measured by the ADCPs of Q2, Q4, and Q5. At Q2 (0° , 93°E ; [Fig. 4a](#)), the subsurface currents at about 120 m are eastward in April 2015 with a velocity greater than 50 cm s^{-1} , while the subthermocline at 350 m has a westward current with a maximum velocity of $\sim 20 \text{ cm s}^{-1}$. The subthermocline currents reverse their directions in mid-May 2015, with

strong seasonal variations in the following months. At the equatorial 85°E mooring, Q4 ([Fig. 4b](#)), the strong westward subthermocline currents during April–June 2015 that exceed 30 cm s^{-1} are the dominant signal. At the westernmost mooring, Q5 (80°E ; [Fig. 4c](#)), although the mooring shows strong vertical movement, there is still a strong seasonal variation in the subthermocline layer with the westward current in April–July and October–December for 2015 and 2016. The World Ocean Circulation Experiment (WOCE) observations also showed similar seasonal variations in the subthermocline zonal current (see plate 1 in [Reppin et al. 1999](#)).

An important phenomenon is that the EIC presents a significant semiannual cycle. As with the monthly mean data shown in [Fig. 4](#), the dominance of the semiannual variability in the subthermocline can be clearly seen at all three mooring locations, with currents flowing westward during April–June and November–January in 2015 and eastward overall for the rest of the year. The situation for 2016 is similar, except that the westward current during boreal fall begins in October—approximately one month earlier than that in 2015. To isolate the subthermocline signal, we obtain the daily time series of the 200–500-m averaged zonal currents ([Fig. 4d](#)) at each mooring location and calculated their lagged autocorrelations ([Fig. 4e](#)). The autocorrelation shows strong semiannual fluctuation during the observational period, with westward (eastward) currents occurring twice a year—consistent with the observations of [Luyten and Roemmich \(1982\)](#) in the western equatorial Indian Ocean.

FIG. 4. Time–depth variations of daily zonal currents observed by moored ADCPs [(a) Q2, (b) Q4, (c) Q5]. The blue dotted lines represent the 200- and 500-m depths. (d) Time series of 200–500-m-depth-averaged daily zonal currents obtained from Q2 (blue), Q4 (red), and Q5 (green). (e) Lagged autocorrelation function of the daily zonal currents derived from Q2 (blue), Q4 (red), and Q5 (green). (f) Lagged correlation between Q2 and Q4 (red) and Q2 and Q5 (blue).

The spectrum of the auto-correlation series of monthly equatorial zonal currents obtained from ORAS4 and LOM-MR in the Indian Ocean further verify the dominant semiannual period of the EIC.

Because of the similar features of the spectrum result of the EIC at different longitudes, we present the results averaged between 60° and 95°E to show the semiannual variation of the EIC using data in the upper 2000 m.

FIG. 5. (a) Vertical distribution of the power spectra of monthly zonal currents averaged between 1°S and 1°N, 60° and 95°E in ORAS4 from 2001–17. A cross represents significance at the 95% level. (b) Depth–time evolution of the climatological seasonal cycle of the zonal velocity averaged between 60° and 95°E obtained from ORAS4 (coloring). The black contour represents zero velocity. (c) Time–longitude evolution of the climatological seasonal cycle between 200–1200-m-depth-averaged zonal velocity obtained from ORAS4 (coloring). The black contour represents zero velocity. (d)–(f) As in (a)–(c), but for LOM-MR from 2001 to 2011.

Figures 5a and 5d show that a large spectral peak occurs at the semiannual period in the upper 2000 m. In agreement with the power spectra, the spatiotemporal evolution of the vertical structure of equatorial zonal velocity averaged between 60° and 95°E and the horizontal structure of equatorial zonal velocity averaged from 200 to 1200 m show strong semiannual patterns (Figs. 5b,e,c,f). These results suggest that the semiannual variation is a basic feature of the EIC in the equatorial basin.

Mooring observations also demonstrate that the EIC has clear zonal phase propagation. The lagged correlations of the 200–500-m averaged zonal currents between Q2 and Q4, as well as between Q4 and Q5, clearly demonstrate this point (Fig. 4f). The correlation coefficient between Q2 and Q4 (Fig. 4f; red line) reaches its highest value of 0.50 when Q2 is leading by 22 days, suggesting a westward-propagating phase from Q2 to Q4. Similarly, Q4 leads Q5 by ~10 days. The zonal propagating phase speed is 46.3 cm s^{-1} from Q2 to Q4 and 63.7 cm s^{-1} from Q4 to Q5, which are roughly

consistent with the first meridional mode Rossby wave speed associated with the second baroclinic mode estimated from realistic simulations of the equatorial Indian Ocean (Han et al. 1999). The subthermocline currents averaged at 200–1200 m from both ORAS4 and LOM-MR verify the westward phase propagation of the observed EIC (Figs. 5c,f). Their results show that a westward current forms in March–May in the eastern basin and then propagates to the western basin in June–August. Subsequently, an eastward flow occurs in June–September in the eastern basin and reaches the western basin in August–October. A similar process occurs in another six months. The westward propagation of the subthermocline velocities has a similar phase speed of $\sim 64.0 \text{ cm s}^{-1}$.

c. Observed vertical structure of the EIC

One of the most obvious features in the moored observations is the dominance of the vertical mode in the zonal current, leading to a distinct current structure that alternates signs with depth (Fig. 4). Based on Q2, the

observed zonal velocities between 60 and 190 m in the thermocline, and between 200 and 500 m in the sub-thermocline, tend to be out of phase. The correlation coefficient of the averaged zonal currents between 60–190 and 200–500 m reaches -0.50 , with a significance level of 99%. The reversed flow direction between the thermocline and subthermocline is verified by the observations from Q4 and Q5 (Figs. 4b,c). Based on the vertical structure from the moored observations, parts of the observed signals show a clear dominant baroclinic mode structure with a visible phase jump, while other sections of the signals show upward phase propagation, such as during September–December 2015 at Q2. The observed vertical structure suggests the importance of the vertically standing wave structure (e.g., a single baroclinic mode structure), and the apparent vertical propagation suggests the significant contributions of other baroclinic modes to the EIC, in addition to the strong vertically standing wave.

The vertical structures of the equatorial zonal velocity obtained from ORAS4 and LOM-MR reproduce most of the features of the moored observations. The zonal velocities show noticeable visible phase jumps (Figs. 5b,e). A maximum is found at ~ 120 m, which corresponds to the core of the EUC. Below the EUC, a signature of the standing mode is evident. The results of LOM-MR show a more visible standing equatorial mode structure, with two zero crossings in the upper 2000 m (and over the entire water column; not shown), which represents the structure of the second baroclinic mode that resonates with the semiannual wind forcing, as discussed in Han et al. (2011). By contrast, ORAS4 presents slightly pronounced vertical phase propagation in the depth range of 200–1200 m. The skill of the LOM-MR in replicating the dominant features of the EIC in the Indian Ocean demonstrates that a dominant part of the horizontal and vertical structure of the velocity field for the EIC can be explained by the linear wind-driven equatorial wave dynamics.

d. Dynamics

To address the role played by the wind-driven equatorial wave dynamics, we examine the solutions to LOM experiments. Clearly, reflected Rossby waves from the eastern boundary interfere constructively with the directly forced Kelvin and Rossby waves' response at the 200–1200-m depth of the EIC (cf. Figs. 6b and 6c), producing an intensified total EIC during most of the year (Fig. 6a). These results confirm that the semiannual basin resonance in the equatorial Indian Ocean is the primary cause for the strong semiannual variability of the EIC. To further quantify this argument, we obtain the time series of climatological EIC averaged between

60° and 95°E and between 200 and 1200 m from LOM-MR, LOM-DAMP, and their difference (Fig. 6d). Again, reflected Rossby waves (green) and the directly forced response (blue) are in phase, producing an intensified total EIC (red). Reflected Rossby waves, however, have apparently larger magnitudes than directly forced response for the 200–1200-m average. As shown in Fig. 6d, the STD of the zonal velocity in the depth range of EIC is 4.54 cm s^{-1} for LOM-MR, 1.06 cm s^{-1} for LOM-DAMP, and 3.58 cm s^{-1} for LOM-REFLECT for the climatological equatorial zonal velocities between 2001 and 2011, with a large percentage of LOM-REFLECT going to LOM-MR. The large westward velocities appear in April and October, with values of -7.41 and -4.58 cm s^{-1} , respectively, in LOM-MR, and values of -6.93 and -3.94 cm s^{-1} in the LOM-REFLECT. Conversely, the strong eastward velocities occur in January and July, with values of 5.03 and 5.99 cm s^{-1} , respectively, in LOM-MR and values of 3.14 and 4.35 cm s^{-1} in LOM-REFLECT, both showing a strong semiannual cycle.

As discussed earlier, the annual and semiannual zonal surface wind forcing in the equatorial Indian Ocean has comparable strength, but the semiannual variability of the zonal current in the observations and simulations is much stronger than the annual variability. This is because the resonant effect is excited by the semiannual wind stress. When resonance occurs, Rossby waves reflected from the eastern ocean boundary intensify the directly forced equatorial Kelvin and Rossby waves in the basin interior, resulting in an enhanced semiannual response in the zonal velocity field. Such basin resonance in the equatorial Indian Ocean is bound up with the eastern boundary reflection, which is fairly robust for the surface and subsurface current systems.

Even though the second baroclinic mode dominates the EIC variability, other baroclinic modes can also be important because the observed EIC shows clear vertical phase propagation during some times of a year, rather than being always dominated by the standing mode structure of the second mode (Fig. 4). Results from the sum of the first 10 modes can reproduce the main features of the EIC well, such as the horizontal and vertical structure, velocity core, and strength (Figs. 7a,b). The higher-order baroclinic modes (modes 11–25) have very small c_n , and therefore very strong friction (because the coefficient of vertical viscosity A/c_n^2 is large, as discussed earlier), which are subject to strong damping and make little contribution to the variability of the EIC. Among these, the second baroclinic mode displays significant semiannual variability with strong velocity reversals, vertical structure, velocity core, and strength over the width of the EIC. This mode is clearly the

FIG. 6. (a) Time–depth plot of the climatological zonal velocities averaged between 1°S and 1°N, 60° and 95°E obtained from (a) LOM-MR, (b) LOM-DAMP, and (c) LOM-REFLECT. (d) Time series of the zonal velocities averaged between 200 and 1200 m, in which the red line is LOM-MR, the blue line is LOM-DAMP, and the green line is LOM-REFLECT.

dominant mode for the semiannual cycle of the EIC (Fig. 7d), which plays a similar role in the semiannual variability of the equatorial zonal surface current and EUC (e.g., Han et al. 1999; Yuan and Han 2006; Nagura and McPhaden 2010; Chen et al. 2015).

Figure 8a quantitatively shows the importance of the second baroclinic mode in generating the seasonal variability of the EIC. The monthly mean zonal velocities averaged in the depth range of 200–1200 m show that the seasonal cycle and velocity amplitude of the zonal currents obtained from the second baroclinic mode (Fig. 8a; blue line) are well matched with that of all 25 modes (Fig. 8a; red solid line) and the first 10 modes (Fig. 8a; red thin line). The STD of the EIC speed derived from the climatological velocities averaged for 2001–11 is 4.53 cm s⁻¹ for the sum of modes 1–25, 4.51 cm s⁻¹ for the sum of modes 1–10, 1.19 cm s⁻¹ for the sum of modes 1 and 3–10, 1.30 cm s⁻¹ for first mode, 5.18 cm s⁻¹ for second mode, and 0.13 cm s⁻¹ for third mode. The analysis also indicates weak amplitudes of the first and third modes in the intermediate layer. The first baroclinic mode has very strong velocities in the upper layer and thus plays an important role in the seasonal cycle of the surface and subsurface currents (Han et al. 1999; Chen et al. 2015). However, the flow field of the first baroclinic mode presents velocities that are relatively

weak and out of phase with the sum of modes 1–25 (low correlation coefficient of -0.19) and thus has little direct impact in the intermediate layer (Figs. 7c and 8a). The third baroclinic mode has the second zero-crossing line near 700 m and is close to zero for the 200–1200-m average because of the cancellation effect (Figs. 7e and 8a); it explains $\sim 3.0\%$ of the total variance of the sum of modes 1–25 and thus has little direct impact on the variability of the EIC.

Even though the second baroclinic mode plays the most important role in causing the observed EIC compared to other baroclinic modes, vertical phase propagations can be clearly seen from the observations during some seasons and some years (e.g., Fig. 4). Note that there is no vertical phase propagation if only one baroclinic mode contributes, because a single baroclinic mode can be viewed as a standing wave. The observed vertical phase propagations suggest the contributions from other baroclinic modes, and these contributions are mentioned above. To understand the downward energy propagation of the EIC, we applied semiannual harmonic analysis and Wentzel–Kramers–Brillouin (WKB) energy rays theory on the first 10 baroclinic modes of LOM-MR (Fig. 9). The details of WKB theory can be found in Philander (1978), Eriksen (1981), Luyten and Roemmich (1982), and Lukas and Firing (1985),

FIG. 7. Longitude–depth plots of the monthly averaged zonal velocities averaged between 1°S and 1°N for January, April, July, and October obtained from LOM-MR (2001–11): (a) sum of modes 1–25, (b) sum of modes 1–10, (c) mode 1, (d) mode 2, (e) mode 3, and (f) sum of modes 1 and 3–10. Black lines are the 0 cm s^{-1} contours.

where the slopes are given as $-\omega_0/N_b(z)$ for the Kelvin ray and $(2n + 1)\omega_0/N_b(z)$ for the n th meridional mode Rossby wave, where ω_0 denotes the semiannual frequency (180-day period) and $N_b(z)$ denotes the background Brünt–Väisälä frequency profile [as given by the N_b used in LOM by Moore and McCreary (1990)]. Note

that the energy rays for equatorial Kelvin and Rossby waves require the contributions from many (at least more than two) baroclinic modes (e.g., McCreary 1984). The results show that the EIC associated with the second baroclinic mode produces an amplitude maximum with a value greater than 10 cm s^{-1} in the

FIG. 8. Comparison of different baroclinic modes for the monthly mean of the intermediate averaged (1°S and 1°N ; 200–1200 m) zonal velocity: (a) central basin (60° – 95°E) and (b) eastern basin (90° – 95°E). The red lines are the sum of 25 modes; the dark green lines are the sum of modes 1–10; the green lines are mode 1; the blue lines are mode 2; and the yellow lines are mode 3.

FIG. 9. Semiannual harmonics of the monthly mean zonal velocities averaged between 1°S and 1°N for the baroclinic modes obtained from LOM-MR (2001–11): (a) sum of modes 1–10, (b) mode 2, (c) sum of modes 1 and 3–10. (top) The amplitude (cm s^{-1}), (middle) the phase (month), and (bottom) the percentage of explained variance (%). The phase is defined by the month with the strongest westward velocity. The WKB theoretical ray paths are shown: the wind-generated Kelvin (dashed green lines) and reflected Rossby rays of the first (solid red lines) and second (dashed red line) meridional modes.

central basin. This makes a significant contribution to the amplitude of the sum of the first 10 modes and broadens the “ray” structure significantly in the central basin [Figs. 9a(1) and 9b(1)]. These results suggest that the equatorial basin resonance associated with the second baroclinic mode and contributions from other baroclinic modes (e.g., modes 1 and 3–10) together explain the observed EIC variability in the equatorial Indian Ocean, which shows dominant modal structure at some times and apparent vertical energy propagation of the long Rossby wave rays during other times.

4. Discussion and conclusions

The EIC in the Indian Ocean occurs beneath the thermocline layer with a prominent semiannual cycle,

which is thought to be related to the wind-driven wave dynamics. Because of the lack of observational data, the spatial and temporal features and dynamics of the EIC were unclear and have not yet been systematically studied. Combining data from moorings and ocean reanalysis with modeling experiments using a LOM, the present study provides a systematic investigation of the features and dynamics of the EIC in the Indian Ocean.

Results from the multiple ADCP observations show that the zonal current in the depth range of 200–500 m reveals a pronounced semiannual cycle along the equatorial Indian Ocean with a maximum velocity of more than 20 cm s^{-1} . This observed zonal current is part of the EIC and presents a clear westward phase propagation. Vertically, the equatorial zonal flow often displays an evident structure of the second baroclinic mode but also shows evident vertical phase propagation

during some times of a year, with the EIC being the subthermocline current. The observed zonal wavelength in the eastern equatorial Indian Ocean, calculated from the westward phase propagation, measures 8000 km between moorings Q2 and Q4 and 11 000 km between moorings Q4 and Q5.

Fair agreement is found between the observations, reanalysis, and LOM in the current structures and amplitude. The results of ORAS4 show that it is capable of realistically characterizing the semiannual reversals of the EIC. LOM is capable of realistically simulating the main features of the Indian Ocean EIC, suggesting the importance of wind-driven linear wave dynamics. The results of a suite of LOM experiments show that Rossby waves reflected from the eastern ocean boundary associated with the second baroclinic mode interfere constructively with the directly forced Kelvin and Rossby waves in the ocean interior, resulting in the strong semiannual variability of the EIC. This is because of the resonant excitation of the equatorial Kelvin and Rossby waves associated with the second baroclinic mode by the semiannual wind in a basin the size of the equatorial Indian Ocean, as shown by Eq. (3).

Further analyses of LOM solutions show that the sum of the first 10 baroclinic modes can reproduce the main characteristics of the EIC well. Among the first 10 modes, the resonant second baroclinic mode plays the dominant role in forming the features of the EIC, including the vertical structure, seasonal cycle, zonal phase propagation, and the velocity core. Other baroclinic modes, however, can also have significant contributions. The superposition of all 10 modes produces the downward energy propagation during some times of a year, as shown by the observations and simulated by the models.

Note that the directly forced waves are weak compared to the reflected Rossby wave in the intermediate layer (Figs. 6b and 6c). For the solutions of directly forced Kelvin and Rossby waves (not shown), the first and second baroclinic modes have comparable amplitude but tend to be out of phase. Thus, they have offsetting contributions to the variability of directly forced EIC. For the solutions of eastern boundary reflected Rossby waves (not shown), the second baroclinic mode of reflected Rossby waves has the strongest contribution to the variability of reflected EIC. These results verify the reflected Rossby waves playing a more important role in the EIC than the directly forced Kelvin and Rossby waves.

Note that during positive Indian Ocean dipole (IOD; e.g., Saji et al. 1999; Webster et al. 1999) years, equatorial westerly wind is weak in boreal fall because of the strong easterly wind anomalies during the IOD peak.

The weak westerly wind in the fall considerably weakens the semiannual variability, and the zonal wind is dominated by the annual cycle (not shown). As a result of the weak semiannual wind, the basin resonance at the semiannual period is also weak. How the variability of the EIC responds to the strong annual wind forcing during IOD years should be further investigated in our future research [similar to Brandt et al. (2016) on the Equatorial Atlantic Circulation]. Note that there are also reflected Kelvin wave rays after directly forced Rossby waves impinge on the western boundary [Fig. 9a(1)]. Without the western boundary, resonance basin modes cannot be set up. However, the enhanced response between directly forced Kelvin and Rossby waves and Rossby waves reflected from the eastern boundary will still occur (Han et al. 2011). More observations and numerical simulations are necessary to examine the exact wave reflection and ray propagation trajectories at the western boundary of equatorial Indian Ocean.

Acknowledgments. We thank two anonymous reviewers for their helpful comments. Advice from Dr. Lei Yang and Dr. Yeqiang Shu are greatly appreciated. (ECMWF ocean reanalysis ORAS4 data were downloaded at <https://www.ecmwf.int/en/research/climate-reanalysis/ocean-reanalysis>. RAMA moored data were downloaded at <https://www.pmel.noaa.gov/tao/drupal/rama-display/>.) Anyone who wants to get access to these mooring data could contact the coauthor Dongxiao Wang (dxwang@scsio.ac.cn). This work is supported by the Major State Research Development Program of China Grant 2016YFC1402603 and 2017YFC1405100; NSFC Grants 41521005, 41706027, 41476011, 41676013, and 41776003; GNSF Grant 2016A030310015; Grant KLOCW1604; Grant LTOZZ1702; NSF AGS 1446480; and Guangzhou Science and Technology Foundation Grant 201804010133.

REFERENCES

- Anderson, D. L. T., and D. J. Carrington, 1993: Modeling interannual variability in the Indian Ocean using momentum fluxes from the operational weather analyses of the United Kingdom Meteorological Office and European Centre for Medium Range Weather Forecasts. *J. Geophys. Res.*, **98**, 12 483–12 499, <https://doi.org/10.1029/93JC00407>.
- Antonov, J. I., R. A. Locarnini, T. P. Boyer, A. V. Mishonov, and H. E. Garcia, 2006: *Salinity*. Vol. 2, *World Ocean Atlas 2005*, NOAA Atlas NESDIS 62, 182 pp.
- Atlas, R., J. Ardizzone, and R. N. Hoffman, 2008: Application of satellite surface wind data to ocean wind analysis. *Remote Sensing System Engineering*, P. E. Ardanuy and J. J. Puschell, Eds., Society of Photo-Optical Instrumentation Engineers (SPIE Proceedings, Vol. 7087), 70870B, <https://doi.org/10.1117/12.795371>.

- Balmaseda, M. A., K. Mogensen, and A. Weaver, 2013a: Evaluation of the ECMWF Ocean Reanalysis ORAS4. *Quart. J. Roy. Meteor. Soc.*, **139**, 1132–1161, <https://doi.org/10.1002/qj.2063>.
- , K. E. Trenberth, and E. Kallen, 2013b: Distinctive climate signals in reanalysis of global ocean heat content. *Geophys. Res. Lett.*, **40**, 1754–1759, <https://doi.org/10.1002/grl.50382>.
- Brandt, P., and C. Eden, 2005: Annual cycle and interannual variability of the mid-depth tropical Atlantic Ocean. *Deep-Sea Res. I*, **52**, 199–219, <https://doi.org/10.1016/j.dsr.2004.03.011>.
- , M. Claus, R. J. Greatbatch, R. Kopte, J. M. Toole, W. E. Johns, and C. W. Böning, 2016: Annual and semiannual cycle of equatorial Atlantic circulation associated with basin-mode resonance. *J. Phys. Oceanogr.*, **46**, 3011–3029, <https://doi.org/10.1175/JPO-D-15-0248.1>.
- Cane, M. A., and D. W. Moore, 1981: A note on low-frequency equatorial basin modes. *J. Phys. Oceanogr.*, **11**, 1578–1584, [https://doi.org/10.1175/1520-0485\(1981\)011<1578:ANOLFE>2.0.CO;2](https://doi.org/10.1175/1520-0485(1981)011<1578:ANOLFE>2.0.CO;2).
- , and E. S. Sarachik, 1981: The response of a linear baroclinic equatorial ocean to periodic forcing. *J. Mar. Res.*, **39**, 651–693.
- Chen, G., W. Han, Y. Li, D. Wang, and M. J. McPhaden, 2015: Seasonal-to-interannual time-scale dynamics of the Equatorial Undercurrent in the Indian Ocean. *J. Phys. Oceanogr.*, **45**, 1532–1553, <https://doi.org/10.1175/JPO-D-14-0225.1>.
- , —, Y. Shu, Y. Li, D. Wang, and Q. Xie, 2016: The role of Equatorial Undercurrent in sustaining the eastern Indian Ocean upwelling. *Geophys. Res. Lett.*, **43**, 6444–6451, <https://doi.org/10.1002/2016GL069433>.
- Cheng, X. H., J. P. McCreary, B. Qiu, Y. Q. Qi, and Y. Du, 2017: Intraseasonal-to-semiannual variability of sea-surface height in the eastern equatorial Indian Ocean and southern Bay of Bengal. *J. Geophys. Res. Oceans*, **122**, 4051–4067, <https://doi.org/10.1002/2016JC012662>.
- Delcroix, T., and C. Henin, 1988: Observations of the Equatorial Intermediate Current in the western Pacific Ocean (165°E). *J. Phys. Oceanogr.*, **18**, 363–366, [https://doi.org/10.1175/1520-0485\(1988\)018<0363:OOTEIC>2.0.CO;2](https://doi.org/10.1175/1520-0485(1988)018<0363:OOTEIC>2.0.CO;2).
- Dewitte, B., and G. Reverdin, 2000: Vertically propagating annual and interannual variability in an OGCM simulation of the tropical Pacific Ocean in 1985–94. *J. Phys. Oceanogr.*, **30**, 1562–1581, [https://doi.org/10.1175/1520-0485\(2000\)030<1562:VPAAIV>2.0.CO;2](https://doi.org/10.1175/1520-0485(2000)030<1562:VPAAIV>2.0.CO;2).
- Donguy, J. R., and G. Meyers, 1995: Observations of geostrophic transport variability in the western tropical Indian Ocean. *Deep-Sea Res.*, **42**, 1007–1028, [https://doi.org/10.1016/0967-0637\(95\)00047-A](https://doi.org/10.1016/0967-0637(95)00047-A).
- Duan, Y., and Coauthors, 2016: Anomalous behaviors of Wyrtki jets in the equatorial Indian Ocean during 2013. *Sci. Rep.*, **6**, 29688, <https://doi.org/10.1038/srep29688>.
- Eriksen, C. C., 1981: Deep currents and their interpretation as equatorial waves in the western Pacific Ocean. *J. Phys. Oceanogr.*, **11**, 48–70, [https://doi.org/10.1175/1520-0485\(1981\)011<0048:Dcatia>2.0.CO;2](https://doi.org/10.1175/1520-0485(1981)011<0048:Dcatia>2.0.CO;2).
- Firing, E., 1987: Deep zonal currents in the central equatorial Pacific. *J. Mar. Res.*, **45**, 791–812, <https://doi.org/10.1357/002224087788327163>.
- Fischer, J., and F. A. Schott, 1997: Seasonal transport variability of the deep western boundary current in the equatorial Atlantic. *J. Geophys. Res.*, **102**, 27 751–27 769, <https://doi.org/10.1029/97JC02327>.
- Fu, L. L., 2007: Intraseasonal variability of the equatorial Indian Ocean observed from sea surface height, wind, and temperature data. *J. Phys. Oceanogr.*, **37**, 188–202, <https://doi.org/10.1175/JPO3006.1>.
- Gent, P. R., 1981: Forced standing equatorial ocean wave modes. *J. Mar. Res.*, **39**, 695–709.
- , K. O'Neill, and M. A. Cane, 1983: A model of the semiannual oscillation in the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **13**, 2148–2160, [https://doi.org/10.1175/1520-0485\(1983\)013<2148:AMOTSO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1983)013<2148:AMOTSO>2.0.CO;2).
- Gouriou, Y., T. Delcroix, and G. Eldin, 2006: Upper and intermediate circulation in the western equatorial Pacific Ocean in October 1999 and April 2000. *Geophys. Res. Lett.*, **33**, L10603, <https://doi.org/10.1029/2006GL025941>.
- Han, W., 2005: Origins and dynamics of the 90-day and 30–60-day variations in the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **35**, 708–728, <https://doi.org/10.1175/JPO2725.1>.
- , J. P. McCreary, D. L. T. Anderson, and A. J. Mariano, 1999: Dynamics of the eastern surface jets in the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **29**, 2191–2209, [https://doi.org/10.1175/1520-0485\(1999\)029<2191:DOTESJ>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<2191:DOTESJ>2.0.CO;2).
- , —, Y. Masumoto, J. Vialard, and B. Duncan, 2011: Basin resonances in the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **41**, 1252–1270, <https://doi.org/10.1175/2011JPO4591.1>.
- Ishizaki, H., H. Nakano, T. Nakano, and N. Shikama, 2014: Evidence of equatorial Rossby wave propagation obtained by deep mooring observations in the western Pacific Ocean. *J. Oceanogr.*, **70**, 463–488, <https://doi.org/10.1007/s10872-014-0247-3>.
- Iskandar, I., Y. Masumoto, and K. Mizuno, 2009: Subsurface equatorial zonal current in the eastern Indian Ocean. *J. Geophys. Res.*, **114**, C06005, <https://doi.org/10.1029/2008JC005188>.
- Jensen, T. G., 1993: Equatorial variability and resonance in a wind-driven Indian Ocean model. *J. Geophys. Res.*, **98**, 22 533–22 552, <https://doi.org/10.1029/93JC02565>.
- Kessler, W. S., and J. P. McCreary, 1993: The annual wind-driven Rossby wave in the subthermocline equatorial Pacific. *J. Phys. Oceanogr.*, **23**, 1192–1207, [https://doi.org/10.1175/1520-0485\(1993\)023<1192:TAWDRW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1993)023<1192:TAWDRW>2.0.CO;2).
- Knox, R. A., 1974: Reconnaissance of the Indian Ocean equatorial undercurrent near Addu Atoll. *Deep-Sea Res. Oceanogr. Abstr.*, **21**, 123–129, [https://doi.org/10.1016/0011-7471\(74\)90069-2](https://doi.org/10.1016/0011-7471(74)90069-2).
- Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, and H. E. Garcia, 2006: *Temperature*. Vol. 1, *World Ocean Atlas 2005*, NOAA Atlas NESDIS 61, 182 pp.
- Lukas, R., and E. Firing, 1985: The annual Rossby wave in the central equatorial Pacific Ocean. *J. Phys. Oceanogr.*, **15**, 55–67, [https://doi.org/10.1175/1520-0485\(1985\)015<0055:TARWIT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1985)015<0055:TARWIT>2.0.CO;2).
- Luyten, J. R., and D. H. Roemmich, 1982: Equatorial currents at semiannual period in the Indian Ocean. *J. Phys. Oceanogr.*, **12**, 406–413, [https://doi.org/10.1175/1520-0485\(1982\)012<0406:ECASAP>2.0.CO;2](https://doi.org/10.1175/1520-0485(1982)012<0406:ECASAP>2.0.CO;2).
- Marin, F., E. Kestenare, T. Delcroix, F. Durand, S. Cravatte, G. Eldin, and R. Bourdalle-Badie, 2010: Annual reversal of the Equatorial Intermediate Current in the Pacific: Observations and model diagnostics. *J. Phys. Oceanogr.*, **40**, 915–933, <https://doi.org/10.1175/2009JPO4318.1>.
- McCreary, J. P., 1981: A linear stratified ocean model of the coastal undercurrent. *Philos. Trans. Roy. Soc. London*, **302A**, 385–413, <https://doi.org/10.1098/rsta.1981.0176>.
- , 1984: Equatorial beams. *J. Mar. Res.*, **42**, 395–430, <https://doi.org/10.1357/002224084788502792>.
- , W. Han, D. Shankar, and S. R. Shetye, 1996: Dynamics of the east India coastal current. 2. Numerical solutions. *J. Geophys. Res.*, **101**, 13 993–14 010, <https://doi.org/10.1029/96JC00560>.

- McPhaden, M. J., 1982: Variability in the central equatorial Indian Ocean. Part 1: Ocean dynamics. *J. Mar. Res.*, **40**, 157–176.
- , and Coauthors, 2009: RAMA: The Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction. *Bull. Amer. Meteor. Soc.*, **90**, 459–480, <https://doi.org/10.1175/2008BAMS2608.1>.
- Mogensen, K., M. Alonso Balmaseda, and A. Weaver, 2012: The NEMOVAR ocean data assimilation system as implemented in the ECMWF ocean analysis for system 4. ECMWF Tech. Memo. 668, 59 pp., <https://www.ecmwf.int/en/library/11174-nemovar-ocean-data-assimilation-system-implemented-ecmwf-ocean-analysis-system-4>.
- Molinari, R. L., D. Olson, and G. Reverdin, 1990: Surface current distributions in the tropical Indian Ocean derived from compilations of surface buoy trajectories. *J. Geophys. Res.*, **95**, 7217–7238, <https://doi.org/10.1029/JC095iC05p07217>.
- Moore, D. W., and J. P. McCreary, 1990: Excitation of intermediate-frequency equatorial waves at a western ocean boundary: With application to observations from the Indian Ocean. *J. Geophys. Res.*, **95**, 5219–5231, <https://doi.org/10.1029/JC095iC04p05219>.
- Nagura, M., and M. J. McPhaden, 2010: Wyrтки jet dynamics: Seasonal variability. *J. Geophys. Res.*, **115**, C07009, <https://doi.org/10.1029/2009JC005922>.
- , and —, 2016: Zonal propagation of near-surface zonal currents in relation to surface wind forcing in the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **46**, 3623–3638, <https://doi.org/10.1175/JPO-D-16-0157.1>.
- Nyadjro, E. S., and M. J. McPhaden, 2014: Variability of zonal currents in the eastern equatorial Indian Ocean on seasonal to interannual time scales. *J. Geophys. Res. Oceans*, **119**, 7969–7986, <https://doi.org/10.1002/2014JC010380>.
- O'Brien, J. J., and H. E. Hurlburt, 1974: An equatorial jet in the Indian Ocean: Theory. *Science*, **184**, 1075–1077, <https://doi.org/10.1126/science.184.4141.1075>.
- Ogata, T., and S. P. Xie, 2011: Semiannual cycle in zonal wind over the equatorial Indian Ocean. *J. Climate*, **24**, 6471–6485, <https://doi.org/10.1175/2011JCLI4243.1>.
- O'Neill, K., and J. R. Luyten, 1984: Equatorial velocity profiles. Part II: Zonal component. *J. Phys. Oceanogr.*, **14**, 1842–1852, [https://doi.org/10.1175/1520-0485\(1984\)014<1842:EVPPIZ>2.0.CO;2](https://doi.org/10.1175/1520-0485(1984)014<1842:EVPPIZ>2.0.CO;2).
- Philander, S. G. H., 1978: Forced oceanic waves. *Rev. Geophys.*, **16**, 15–46, <https://doi.org/10.1029/RG016i001p00015>.
- Reppin, J., F. A. Schott, J. Fischer, and D. Quadfasel, 1999: Equatorial currents and transports in the upper central Indian Ocean: Annual cycle and interannual variability. *J. Geophys. Res.*, **104**, 15 495–15 514, <https://doi.org/10.1029/1999JC900093>.
- Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole in the tropical Indian Ocean. *Nature*, **401**, 360–363, <https://doi.org/10.1038/43854>.
- Schott, F., J. Fischer, U. Gartnericht, and D. Quadfasel, 1997: Summer monsoon response of the northern Somali Current, 1995. *Geophys. Res. Lett.*, **24**, 2565–2568, <https://doi.org/10.1029/97GL00888>.
- Shankar, D., J. P. McCreary, W. Han, and S. R. Shetye, 1996: Dynamics of the east India Coastal Current: 1. Analytic solutions forced by interior Ekman pumping and local alongshore winds. *J. Geophys. Res.*, **101**, 13 975–13 991, <https://doi.org/10.1029/96JC00559>.
- Song, Y. T., and F. Colberg, 2011: Deep ocean warming assessed from altimeters, Gravity Recovery and Climate Experiment, in situ measurements, and a non-Boussinesq ocean general circulation model. *J. Geophys. Res.*, **116**, C02020, <https://doi.org/10.1029/2010JC006601>.
- Thierry, V., A. M. Treguier, and H. Mercier, 2004: Numerical study of the annual and semi-annual fluctuations in the deep equatorial Atlantic Ocean. *Ocean Modell.*, **6**, 1–30, [https://doi.org/10.1016/S1463-5003\(02\)00054-9](https://doi.org/10.1016/S1463-5003(02)00054-9).
- Wang, F., J. N. Wang, C. Guan, Q. Ma, and D. X. Zhang, 2016: Mooring observations of equatorial currents in the upper 1000 m of the western Pacific Ocean during 2014. *J. Geophys. Res. Oceans*, **121**, 3730–3740, <https://doi.org/10.1002/2015JC011510>.
- Webster, P. J., A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled oceanic–atmospheric dynamics in the Indian Ocean during 1997–98. *Nature*, **401**, 356–360, <https://doi.org/10.1038/43848>.
- Wunsch, C., 1977: Response of an equatorial ocean to a periodic monsoon. *J. Phys. Oceanogr.*, **7**, 497–511, [https://doi.org/10.1175/1520-0485\(1977\)007<0497:ROAEOT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1977)007<0497:ROAEOT>2.0.CO;2).
- Wyrtki, K., 1973: An equatorial jet in the Indian Ocean. *Science*, **181**, 262–264, <https://doi.org/10.1126/science.181.4096.262>.
- Yamagata, T., K. Mizuno, and Y. Masumoto, 1996: Seasonal variations in the equatorial Indian Ocean and their impact on the Lombok Throughflow. *J. Geophys. Res.*, **101**, 12 465–12 473, <https://doi.org/10.1029/95JC03623>.
- Yu, W. D., B. Q. Xiang, L. Liu, and N. Liu, 2005: Understanding the origins of interannual thermocline variations in the tropical Indian Ocean. *Geophys. Res. Lett.*, **32**, L24706, <https://doi.org/10.1029/2005GL024327>.
- Yuan, D. L., and W. Q. Han, 2006: Roles of equatorial waves and western boundary reflection in the seasonal circulation of the equatorial Indian Ocean. *J. Phys. Oceanogr.*, **36**, 930–944, <https://doi.org/10.1175/JPO2905.1>.