Seasonal-to-Interannual Time-Scale Dynamics of the Equatorial Undercurrent in the Indian Ocean*

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ABSTRACT

This paper investigates the structure and dynamics of the Equatorial Undercurrent (EUC) of the Indian Ocean by analyzing in situ observations and reanalysis data and performing ocean model experiments using an ocean general circulation model and a linear continuously stratified ocean model. The results show that the EUC regularly occurs in each boreal winter and spring, particularly during February and April, consistent with existing studies. The EUC generally has a core depth near the 20°C isotherm and can be present across the equatorial basin. The EUC reappears during summer–fall of most years, with core depth located at different longitudes and depths. In the western basin, the EUC results primarily from equatorial Kelvin and Rossby waves directly forced by equatorial easterly winds. In the central and eastern basin, however, reflected Rossby waves from the eastern boundary play a crucial role. While the first two baroclinic modes make the largest contribution, intermediate modes 3–8 are also important. The summer–fall EUC tends to occur in the western basin but exhibits obvious interannual variability in the eastern basin. During positive Indian Ocean dipole (IOD) years, the eastern basin EUC results largely from Rossby waves reflected from the eastern boundary, with directly forced Kelvin and Rossby waves also having significant contributions. However, the eastern basin EUC disappears during negative IOD and normal years because westerly wind anomalies force a westward pressure gradient force and thus westward subsurface flow, which cancels the eastward subsurface flow induced by eastern boundary–reflected Rossby waves. Interannual variability of zonal equatorial wind that drives the EUC variability is dominated by the zonal sea surface temperature (SST) gradients associated with IOD and is much less influenced by equatorial wind associated with Indian monsoon rainfall strength.

1. Introduction

In the Pacific and Atlantic Oceans, the Equatorial Undercurrent (EUC) is quasi permanent in nature (e.g., McPhaden 1986; Metcalf and Stalcup 1967; Izumo 2005); it flows eastward near the top of the thermocline under a typically westward-flowing surface current, driven by the prevailing easterly trade winds. By contrast, in the
Indian Ocean the EUC is a transient feature. It is associated with equatorial waves driven by the strong, seasonally varying component of surface wind (Fig. 1; Schott and McCreary 2001). Early observations suggested that the EUC was consistently detected from year to year only during boreal winter and boreal spring, typically from February to June. Based on current meter measurements, Knauss and Taft (1964) observed an EUC, with maximum speeds increasing from 0.27 to 0.81 m s\(^{-1}\) from the west to the east during March and April 1963, but they did not detect an eastward EUC in August 1962. Swallow (1964, 1967) reported a strong EUC from March to June 1964 near 58\(^\circ\) and 67.3\(^\circ\)E, with a maximum speed exceeding 1.20 m s\(^{-1}\), a magnitude comparable to the EUC of the Pacific Ocean. Indeed, the existence of an eastward EUC during February–June has been further confirmed by later observations at various locations along the equator between 53\(^\circ\) and 91\(^\circ\)E (e.g., Taft 1967; Knox 1974, 1976; Rao and Jayaraman 1968; Schott et al. 1997; Reppin et al. 1999; Sengupta et al. 2007). Meanders of the EUC with its core being displaced from 3\(^\circ\)S to 2\(^\circ\)N along section 55.5\(^\circ\)E was also reported for February–June of 1975 and 1976 (Leetmaa and Stommel 1980). In this paper, seasons refer to those of the Northern Hemisphere. The EUC is defined as an eastward zonal flow in the equatorial Indian Ocean with its core located in the thermocline above 300-m depth and beneath a westward- or weaker eastward-flowing surface current, and it lasts for at least 1 month.

While the Indian Ocean EUC regularly occurs only during winter and spring based on earlier observations at different locations, reappearance of the EUC in late summer and early fall has also been detected for some years. Bruce (1973) observed an EUC in the western Indian Ocean during late August of 1964, with a core depth of near 75 m. Reppin et al. (1999) reported a reappearance of EUC in August 1994 at 80.5\(^\circ\)E with a core near 150 m, following the winter–spring EUC observed from February to May. The magnitude of the reappeared EUC exceeds 0.40 m s\(^{-1}\), and it flows under a westward surface current (Reppin et al. 1999). Note that 1994 is a year of positive Indian Ocean dipole (IOD) event (e.g., Saji et al. 1999), as is 2006 when the presence of EUC in August is also observed (Iskandar et al. 2009). Iskandar et al. (2009) analyzed 6 yr (December 2000–November 2006) of mooring data of the eastern equatorial Indian Ocean at 0\(^\circ\), 90\(^\circ\)E and showed that the subsurface zonal current between 90- and 170-m depths generally exhibits apparent semiannual variability, flowing eastward during winter and summer except for 2003, when the summer-time eastward current vanishes. They suggested that upwelling Kelvin waves driven by wintertime easterly winds is the primary cause for the eastward subsurface flow in boreal winter–spring. During summer, they suggested that both wind-driven Kelvin waves and eastern boundary–reflected Rossby waves contribute.

Numerical ocean models are able to simulate the observed EUC over the Indian Ocean. Using an ocean general circulation model (OGCM), Anderson and...
Cartrington (1993) reproduced the winter–spring EUC and the strong semiannual variability of surface and subsurface currents. Han et al. (2004) successfully simulated the observed winter–spring EUC and late summer reappearance in 1994. They pointed out that easterly wind anomalies associated with the positive IOD events (referred to as Indian Ocean zonal mode in Han et al. 2004) during 1994 and 1997 drive a westward surface current and an eastward EUC from August to December, a situation that resembles the EUC in the Pacific and Atlantic. They further argued that during positive IOD years, the EUC forms in August and lasts through December. The EUC shoals in the eastern basin because of a shallower thermocline, which is associated with upwelling in the eastern equatorial Indian Ocean during IOD events. Swapna and Krishnan (2008) and Krishna and Swapna (2009) examined the effects of IOD and Indian monsoon winds on setting up the summertime (June–September) EUC. They concluded that intensification of summer monsoon winds during positive IOD events produces nonlinear amplification of easterly wind stress anomalies to the south of the equator and enhances upwelling in the eastern Indian Ocean off Sumatra–Java, so that the thermocline shoaling provides a zonal pressure gradient and drives anomalous EUC in the subsurface. Recently, Nyadjro and McPhaden (2014) demonstrated how surface wind stress, zonal pressure gradient, and zonal flow in the thermocline vary consistently with the phase of the IOD, using 52 yr (1960–2011) of the ECMWF Ocean Reanalysis System, version 4 (ORAS4), ocean reanalysis product.

Even though progress has been made in observing and understanding the Indian Ocean EUC, its spatial structure, seasonal-to-interannual variability, and the fundamental dynamics that govern the EUC and its variability remain unclear and have not yet been systematically investigated. Equatorial waves have been suggested to remain unclear and have not yet been systematically investigated. The current measurements used in this study are obtained from two moorings of the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA; see McPhaden et al. 2009) located at the Indian Ocean equator. One mooring is deployed at 0°, 90°E and provides data from 14 November 2000 to 7 June 2012. The other is deployed at 0°, 80.5°E and provides data from 27 October 2004 to 17 August 2012. The two moorings cover depth ranges from 40 to 410 m and 25 to 350 m with every 10- and 5-m interval, respectively. These resolutions are reasonable for resolving the vertical structure of the EUC. Daily data from these moorings are used for our analysis.

In addition, monthly currents from ORAS4 (Balmaseda et al. 2013) from 1958 to 2011 are used to examine temporal variability and spatial structure of the EUC. ORAS4 assimilates temperature and salinity data, spanning the period 1958 to the present with a horizontal resolution of 1° × 1° and 42 vertical levels. The National Oceanic and Atmospheric Administration (NOAA) weekly sea surface temperature (SST) data (Reynolds et al. 2002) from 1982 are used to calculate the dipole mode index (DMI; http://stateoftheocean.osmc.noaa.gov/sur/ind/dmi.php), which is defined as the SST anomaly difference between the western node (10°S–10°N, 70°–90°E) and eastern node (10°S–0°, 90°–110°E) regions (Saji et al. 1999). Homogeneous Indian monthly rainfall datasets (1871–2012) from the Indian Institute of Tropical Meteorology (http://www.tropmet.res.in/Data%20Archival-51-Page) are used to produce the monsoon index. The data diagnostics also include sea surface height averaged for the 2001–11 period from the Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVISO) project (http://www.aviso.altimetry.fr/en/data/products/sea-surface-height-products.html), surface winds for the period 1948–2012 from the National Centers for Environmental Prediction (NCEP) reanalysis (Kistler et al. 2001), and the cross-calibrated multiplatform (CCMP) satellite ocean surface wind vectors available during July 1987–December 2011 (Atlas et al. 2008).

b. The OGCM and experiments

The OGCM used in this study is a recent version of the Hybrid Coordinate Ocean Model (HYCOM; e.g., Bleck...
configured to the Indian Ocean basin (50°S–30°N, 30°–122.5°E) with a horizontal resolution of 0.25° × 0.25° and 26 vertical layers. The surface forcing fields include the 0.25° × 0.25° CCMP ocean surface winds available from July 1987 to December 2011; precipitation from the 0.25° × 0.25° Tropical Rainfall Measuring Mission (TRMM) Multi-satellite Precipitation Analysis (TMPA) level 3B42 product (Kummerow et al. 1998) available for 1998–2011; 2-m air temperature and humidity from the ECMWF interim reanalysis (ERA-Interim) products (Dee et al. 2011) available since 1979; and surface shortwave and longwave radiation from Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996; Loeb et al. 2001) available since 2000. Surface latent and sensible heat fluxes are calculated from wind speed, air temperature, specific humidity and model sea surface temperature using the Coupled Ocean–Atmosphere Response Experiment, version 3 (COARE 3.0), algorithm (Kara et al. 2005). Further details about the model and forcing fields can be found in Li et al. (2014). Earlier versions of HYCOM have been successfully used to understand wave dynamics in the equatorial Indian (e.g., Han et al. 2011) and Atlantic Oceans (e.g., Han et al. 2008).

Taking the World Ocean Atlas 2009 (WOA09) annual climatology of temperature and salinity as initial conditions, the model is spun up from a state of rest for 30 yr. Restarting from the spun-up solution, HYCOM is integrated forward in time from 1 March 2000 to 30 November 2011 with the daily forcing fields described above. To exclude the transient effect of the model simulation in 2000, we analyze the 3-day-averaged model outputs from 2001 to 2011.

c. The linear ocean model and experiments

To help elucidate the wind-driven equatorial wave dynamics, a continuously stratified linear ocean model (LOM) is also used. The model is described in detail in McCreary (1980, 1981) and is applied to several Indian Ocean studies (e.g., Shankar et al. 1996; McCreary et al. 1996; Yuan and Han 2006). Although nonlinearity is essential for understanding the mean equatorial zonal flow near the surface, it is relatively weak in the thermocline on seasonal time scales (Nagura and McPhaden 2014). Consequently, linear dynamics can be a good approximation for seasonal and interannual variability of the transient EUC in the Indian Ocean. Here, we only introduce aspects that are essential for our discussion. The equations of motion are linearized about a background state of rest with stratification represented by Brünt–Väisälä frequency, and the ocean bottom is assumed flat at 4000 m. Under these restrictions, solutions for the zonal velocity $u$, meridional velocity $v$, and pressure $p$ can be expanded in the vertical normal modes with eigenfunctions $\psi_n(z)$:

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

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

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

where $N$ is the total mode number. The terms $u_n$, $v_n$, and $p_n$ are expansion coefficients that satisfy the following equations:

$$(\partial_t + A \frac{A}{c_n}) u_n - f v_n + \frac{1}{\rho} p_{nx} = \tau^x Z_n(pH_n) + v_2 \nabla^2 u_n, \quad (2a)$$

$$(\partial_t + A \frac{A}{c_n}) v_n + f u_n + \frac{1}{\rho} p_{ny} = \tau^x Z_n(pH_n) + v_2 \nabla^2 v_n, \quad \text{and} \quad (2b)$$

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

In the above, $c_n$ represents the equatorial Kelvin wave speed for vertical mode number $n$. The $c_n$ values for the first eight baroclinic modes ($n = 1, 2, \ldots, 8$) are 264, 167, 105, 75, 60, 49, 42, and 37 m s$^{-1}$ with observed stratification within the Indian Ocean (e.g., McCreary et al. 1996; Han et al. 2004). The factors $Z_n = \int_{-Y}^{0} Z(z) \psi_n dz$ and $H_n = \int_{-Y}^{0} \psi_n dz$ determine how strongly the driving wind couples to each mode; $f = \beta y$ is the Coriolis parameter under equatorial beta-plane approximation; and $Z(z)$ 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. The terms associated with $A^2/c_n^2$ represent vertical friction, with $A = 0.00013$ cm$^2$ s$^{-3}$. Density $\rho = 1$ g cm$^{-3}$ is a typical density value of seawater. More details can be found in McCreary et al. (1996).

The LOM is configured for the tropical Indian Ocean north of 29°S with a horizontal resolution of 25 km × 25 km, which is close to the 0.25° × 0.25° HYCOM grids. The total solution is the sum of the first 25 modes [$N = 25$ in Eqs. (1a)–(1c)], and the solutions are well converged with this choice. The linear model is first spun up for 20 yr and then integrated forward in time for the period of 1988–2011 using monthly CCMP wind stress forcing. This solution is referred to as LOM main run (LOM_MR). To isolate the effects of eastern boundary–reflected Rossby waves, a damper in the eastern equatorial ocean is applied [see McCreary et al. (1996) for.
detailed descriptions of the damper]. The damper 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. We refer to this solution as LOM_DAMP. The difference (LOM_MR − LOM_DAMP) isolates the reflected Rossby wave effects.

Note that the damper is not effective at the western boundary even for a straight north–south meridional wall. Therefore, we will not assess the effect of western boundary reflection, which is more complex than the eastern boundary because eastward-propagating short Rossby waves are also involved in addition to the Kelvin wave, and strong mixing near the western boundary can quickly damp the short Rossby waves. This is consistent with Le Blanc and Boulanger (2001), who suggested that at the western boundary of the equatorial Indian Ocean, reflection of the equatorial Rossby wave into the Kelvin wave is subject to large seasonal and interannual variability. The reflection efficiency at the western boundary is much lower than that of the eastern boundary, where reflection efficiency is 85% in terms of amplitude for equatorial Kelvin waves reflecting into first meridional mode Rossby waves. Nagura and McPhaden (2010a,b) studied the equatorial wave effect on seasonal to interannual variability of the Wyrtki jet by assuming reflection efficiencies of 85% at both eastern and western boundaries. However, as Nagura and McPhaden (2014b) showed, velocity variability in the interior of the Indian Ocean is relatively insensitive to western boundary reflections. Thus, LOM_DAMP primarily measures the directly forced Kelvin and Rossby waves, with western boundary reflection playing a minor role.

3. Observed EUC: Temporal variability and spatial structure

To emphasize the seasonal and interannual variability of EUC, we first apply a 31-day running mean to the daily RAMA data and calculate monthly mean values from the 3-day HYCOM output. Monthly climatology of zonal currents is obtained for the 2001–11 period, using the values of each year (rather than just the strong EUC years) to calculate the climatological mean. Then the seasonal variability of EUC is identified from the monthly climatology, and interannual variability is identified using the monthly data of each year.

Consistent with previous studies, the 31-day running-mean zonal currents from the mooring deployed at 0°, 90°E for the 2001–12 period show that the EUC in the Indian Ocean is a transient phenomenon; it consistently appears for each year during boreal winter–spring, particularly from February to April (Fig. 2), based on its definition given in section 1. The depth of the winter–spring EUC core is near the depth of 20°C isotherm (D20) within the thermocline, with the exception of 2007 when the EUC core is located near D23, which is in the upper thermocline (Fig. 2), where the vertical temperature gradient often reaches the maximum amplitude (not shown). The maximum speed of the EUC during February–April can reach 0.75 m s\(^{-1}\). From April to May, the eastward spring Wyrtki jet (Wyrtki 1973) appears, and the EUC weakens and cannot be identified in May. Consistent with previous observational studies (e.g., McPhaden 1982; Reppin et al. 1999), there is a clear upward phase propagation, which suggests downward propagation of equatorial wave energy.

In agreement with Iskandar et al. (2009), subsurface zonal flow between 90° and 170°E in the equatorial Indian Ocean exhibits significant semiannual variability, flowing eastward again from late summer to early fall in most years (Fig. 2). However, not all eastward-flowing subsurface currents can be defined as EUC; only the ones with subsurface maxima flowing under westward or weakly eastward surface currents are defined as EUC. From August–October, the EUC reappears in several years during the 2001–11 period, but it is completely absent in 2003, 2008, and 2010 at this mooring location (Fig. 2). At the mooring location of the central equatorial Indian Ocean (0°, 80.5°E), the observed subsurface currents from 2004 to 2012 also show significant semiannual variability (Fig. 2). There are, however, apparent differences between the two locations. For example, an EUC with its core velocity of 0.58 m s\(^{-1}\) is observed in July (with missing data in August) 2008 at 80.5°E, but it is absent at 90°E. The obvious discrepancy between the two moorings suggests that the EUC—particularly its summer and fall reappearance—is a less robust feature of the circulation than the boreal winter–spring EUC.

To reveal the spatial structure of the EUC, we examine the time evolution of the zonal current from ORAS4 data, since the long current time series from RAMA moorings are only available at 90° and 80.5°E. ORAS4 data agree reasonably well with the RAMA observations, even though the EUC magnitudes in ORAS4 currents are systematically weaker than the mooring observations (Fig. 3) as also found by Nyadjro and McPhaden (2014), likely because of the 1° × 1° area mean for ORAS4 data compared to RAMA data for a specific location. The 2001–11 mean monthly climatology of the ORAS4 zonal current (Fig. 4) shows that the winter–spring EUC can exist across the equatorial basin and obtains its maximum amplitude (~0.38 m s\(^{-1}\)) in the central and western basins from March to April. The EUC reappears in August–September with a much weaker magnitude (~0.21 m s\(^{-1}\)). By analyzing vertical
sections of zonal velocity along the equator for individual years (figure not shown), we find that the EUC exists from summer to early fall of all years during 2001–11 except for 2010 (a negative IOD year), when the EUC disappears across the basin (see section 4c). While the winter–spring EUC generally has a basin-scale structure (section 4b), the summer–fall EUC has large interannual variability regarding its location and depth, which explains its weak amplitude in the 2001–11 11-yr average. Indeed, for all the 54-yr ORAS4 products from 1958 to 2011, the EUC essentially occurs for each spring, and it reappears in the summer–early fall of most
years at different regions of the equatorial Indian Ocean.

4. Model-simulated EUC and dynamics
   a. Model/data comparison

To verify the HYCOM model performance, we first compare the model solutions with RAMA data and ORAS4 products. The simulated monthly zonal currents at the RAMA location (0°, 90°E) from HYCOM generally show consistent patterns of the EUCs and Wyrtki jets with the RAMA observations, even though the model sometimes overestimates the current magnitudes (Fig. 5). To further quantify the model/data comparison and estimate the simulation errors, the 60–140-m depth-averaged zonal currents, which span the EUC core depths of both winter–spring and summer–fall, are shown in Fig. 6. The phase of the monthly zonal currents from the model agrees well with the observations, with a correlation coefficient of 0.83 (above the 95% confidence level). The observed and simulated zonal current standard deviations (STDs) are 0.21 and 0.23 m s⁻¹, with a root-mean-square error (RMSE) of the model relative to the observations of 0.12 m s⁻¹. At 0°, 80.5°E, HYCOM subsurface currents also agree well with the mooring data, with a correlation coefficient of 0.84 from October 2004 to December 2011, observed and simulated STDs of 0.21 and 0.25 m s⁻¹, and RMSE of 0.13 m s⁻¹. Note that the HYCOM results during periods when RAMA data are missing are not used to estimate model errors. Note that not all eastward currents during summer–fall shown in Fig. 6 can be defined as EUC. This point will be further discussed in sections 4b and 4c below.

HYCOM also reasonably simulates the climatological features of the EUC, including its semiannual variability in the central-western basin (cf. Figs. 4 and 7). The correlation coefficient of the area-averaged zonal velocities for 0.5°S–0.5°N, 50°–95°E and 60–140 m between HYCOM and ORAS4 reaches 0.93 (Fig. 8).

The LOM also reasonably captures the subsurface current amplitude and variability (Figs. 6, 8), but it often systematically overestimates the current variability amplitude. This is because nonlinearity in the ocean (both observations and HYCOM) tends to limit the linear growth that occurs in the LOM. Time series for 60–140-m-averaged zonal currents from LOM and RAMA have reasonable agreement, with correlation coefficients of 0.51 and 0.54 at 90° and 80.5°E, respectively (Fig. 6). The STDs are 0.25 and 0.28 m s⁻¹ at 90° and 80.5°E for LOM, which are larger than the 0.21 m s⁻¹ from RAMA data. The area-mean subsurface zonal currents across the Indian Ocean equator for 0.5°S–0.5°N, 50°–95°E and 60–140 m from LOM agree well with those from ORAS4 and HYCOM (Fig. 8). The correlation coefficients between LOM and HYCOM (ORAS4) are 0.75 (0.74). These good agreements demonstrate that the LOM can be used to shed light on the roles played by the wind-driven equatorial waves in determining the Indian Ocean EUC and variability.

To examine the sensitivity of model/data agreement to the depth selection, we also analyzed the zonal currents averaged over the 100–140-m depths and obtained similar agreement with those shown in Figs. 6 and 8 (not shown). Because the 100–140 m depth is below the summer–fall EUC core for the positive IOD years (e.g., 2006, 2007, 2008, and 2011 of Fig. 5; see section 4c), we choose to show currents for 60–140 m depth, which
cover the EUC core for both the winter–spring and summer–fall seasons.

b. Dynamics: Winter–spring EUC

Since the basinwide EUC regularly occurs each year during winter–spring when equatorial easterly wind prevails, we first examine the time evolution of wind and EUC from January to May. Given that HYCOM solutions agree with RAMA moorings as well as ORAS4 data, we will primarily present HYCOM results hereafter, together with LOM solutions to elaborate the winter–spring EUC dynamics. The spatial patterns of subsurface currents from LOM_MR have overall agreement with those of the HYCOM and ORAS4 data, even though significant differences exist regarding detailed structures and magnitudes (cf. Figs. 9b–d). These differences reflect the effects of nonlinearity in the oceanic system, which are included in ORAS4 data and HYCOM solution but excluded from LOM_MR. Note that from February to April, subsurface zonal currents averaged over 60–140 m of each year (Fig. 9) can be defined as EUC in most regions of the Indian Ocean as discussed above (also see red solid and red dashed lines of Fig. 10). Therefore, we generally refer to the 60–140-m-averaged current shown in Fig. 9 as the EUC.

From January to February, easterly wind occupies most regions of the equatorial Indian Ocean, particularly the western and central basin (Fig. 9a). From March to May, the easterlies are gradually replaced by westerlies, which drive the spring Wyrtki jet at the surface (Wyrtki 1973; Han et al. 1999; Nagura and McPhaden 2010a). Corresponding to the easterly winds, the winter–spring eastward EUC from HYCOM is set up along the equator after ~1 month (Fig. 9b; see also Nyadjro and McPhaden 2014). The correlation coefficient between zonal wind stress averaged over 5°S–5°N, 50°–90°E and zonal current from HYCOM averaged over 0.5°S–0.5°N, 50°–90°E, and from 60–140 m reaches −0.44 (−0.47) above 95% significance when the wind leads by 2 (1) months for the entire (January–May) 2001–11 period. During 2001, 2008, and 2011, boreal wintertime easterlies are confined to the western basin west of 70°E (Fig. 9a). Corresponding to
FIG. 5. Monthly zonal current at 0°, 90°E during (a) February–April from the mooring, (b) February–April from HYCOM, (c) August–October from the mooring, and (d) August–October from HYCOM. White and black lines represent the 0 and 0.2 m s$^{-1}$ contours, respectively.
these wind anomalies, the EUCs are evidently weaker and appear primarily in the western basin (Fig. 9b). Currents from the ORAS4 reanalysis product are very similar to those from HYCOM (Figs. 9b,c).

To assess the roles played by the wind-driven equatorial wave dynamics, we examine the solutions to the LOM. Overall, Rossby and Kelvin waves directly forced by winds are the major cause for generating the EUC in the western basin, where the EUC often obtains its maximum (cf. Figs. 9e and 9d). Reflected Rossby waves from the eastern boundary (Fig. 9f), however, also have significant contributions in the western basin and play a role in generating the EUC in the eastern and central basin. For example, the 2002 EUC, which obtains its maximum in the central basin, results primarily from the reflected Rossby waves. The 2006 and 2010 EUCs also receive large contributions from the reflected Rossby waves (cf. Figs. 9d–f). To further quantify the above arguments, we obtain the time series of EUC core values from LOM_MR and contributions from directly forced and reflected Rossby waves (Fig. 10). As shown by Fig. 10 (red solid and dashed lines), all subsurface eastward currents from February to April are defined as EUC. The STD of the EUC core speed is 0.12 m s\(^{-1}\) for LOM_MR, 0.15 m s\(^{-1}\) for LOM_DAMP, and 0.11 m s\(^{-1}\) for (LOM_MR – LOM_DAMP) from 2001 to 2011. The large STD value from LOM_DAMP suggests the importance of directly forced Rossby and Kelvin waves to the EUC core, which is located in the western basin for most years (Figs. 9b–d). The correlation between the total EUC-core strength (Fig. 10, line with squares) and directly forced waves (line with circles) is 0.69, which is much larger than the correlation coefficient of 0.12 between the total EUC and reflected Rossby waves (line with triangles), even though the STD value of reflected Rossby waves (0.11 m s\(^{-1}\)) is comparable to the STD of LOM_MR (0.12 m s\(^{-1}\)). This is because most EUC cores are located in the western basin, which is dominated by the directly forced response and thus the high correlation.

Even though the EUC varies from year to year, it occurs every winter–spring and often extends across the equatorial basin. Thus, we further explore the fundamental dynamics of winter–spring EUC below using the monthly mean climatology from 2001 to 2011. To reveal the spatial structure of the equatorial waves, we show horizontal maps of pressure and currents averaged over the EUC depth (Fig. 11). To isolate the long-wave dynamics associated with the EUC, we consider the situation with only zonal wind stress forcing, under long wavelength approximation \([\partial_t + (A/c_n^2)]u_n = 0\) in Eq. (2b), and neglecting horizontal mixing. Under these assumptions and considering the fact that wind is exerted near the ocean surface and drives the subsurface EUC by setting up pressure gradient, and that \(y = 0\) at the equator, Eqs. (2a) and (2b) for EUC depth yield
\[
\left(\frac{\partial}{\partial t} + \frac{A}{c_n^2}\right)u_n + \frac{1}{p} p_{nx} = 0, \quad \text{and} \quad (3a)
\]
\[
u_n = -\frac{1}{\beta p^2} \frac{\partial^2 p_n}{\partial y^2}. \quad (3b)
\]
Note that we took the \(y\) derivative of Eq. (2b) to obtain Eq. (3b). In this simplified system, only long Rossby and Kelvin waves are retained, and their associated
subsurface zonal flow obeys equatorial “geostrophy,” as shown by Eq. (3b). Indeed, the $-\left(1/r_b\right)\left(\partial^2 p/\partial y^2\right)$ line agrees very well with the subsurface zonal current at the EUC depth range for all seasons (cf. the thick solid lines of gray and black of Fig. 8), suggesting the validity of long-wave approximation and the importance of long Rossby and Kelvin waves in affecting the EUC. Equation (3a) describes the relationship between subsurface zonal flow and zonal pressure gradient associated with the damped long waves.

During January, easterly wind components prevail in the western and central equatorial Indian Ocean.

Fig. 7. As in Fig. 4, but for the HYCOM model for 2001–11.

Fig. 8. Area-averaged monthly zonal velocities for 0.5°S–0.5°N, 50°–95°E and 60–140 m from HYCOM (thin black line), ORAS4 (dashed line), and LOM (thick black line). Gray line is the second derivative of pressure term (PGD), $-\left(1/r_b\right)\left(\partial^2 p/\partial y^2\right)$, which equals $\alpha$ in Eq. (3b) under the long-wave approximation. The 2001–11 mean value (including all months) of the data is removed.
which excites eastward-propagating equatorial Kelvin waves and westward-propagating Rossby waves [Eq. (2a)]. After these waves radiate out, pressure increases in the west and the eastward pressure gradient force is set up, which tends to balance the westward (easterly) wind near the surface [Eq. (2a)]. Note that we neglect horizontal mixing and vertical friction and consider $f = 0$ on the equator in the above discussion. The setup of the eastward pressure gradient force takes about 1 month (February panel of Fig. 11), a time scale consistent with the 1-month lead of wind discussed earlier (Figs. 9a,b). The eastward pressure gradient force drives an eastward EUC [Eq. (3a); February panel of Fig. 11]. During February, the easterly wind component increases (Fig. 9a) and the eastward pressure gradient force, together with the EUC, strengthen and extend eastward during March (Fig. 11). Meanwhile, in the eastern equatorial basin, high pressure occurs in March, which corresponds to the relaxation of the equatorial easterly wind component (Figs. 1, 9a). Reflected Rossby waves from the eastern boundary are the primary cause for the eastward subsurface current in the eastern basin and have significant contributions to the EUC in the central and western basin from February to April (Figs. 9d–f). From April to May, equatorial westerlies prevail, which increase the subsurface pressure in the eastern basin and weaken the eastward EUC there.

Previous studies suggest that equatorial Kelvin and Rossby waves associated with the first and second baroclinic modes primarily determine the seasonal
variability of zonal surface currents in the equatorial Indian Ocean (Han et al. 1999; Nagura and McPhaden 2010a). Consistent with these studies, the first two baroclinic modes also dominate the surface Wyrtki jets in our LOM_MR (not shown). However, the situation is notably different in the subsurface layer. Currents at 60–140 m along the equator are well represented by the sum of the first eight modes (Figs. 12a,b). The low-order modes (e.g., modes \(n = 1, 2\)) have larger \(c_n\) (section 2b) and thus weaker friction because \(A/c_n^2\) is small [Eq. (3a)]. They are approximately the response of inviscid Kelvin and Rossby waves and are very important for generating the winter–spring EUC (Figs. 12c–e). The second baroclinic mode is the dominant mode of semiannual variability of equatorial zonal current, as is also shown by previous studies (e.g., Han et al. 1999; Nagura and McPhaden 2010a; Yuan and Han 2006). The intermediate modes (\(n = 3, 4, \ldots, 8\)) have relatively lower \(c_n\) and therefore larger friction. They are equatorial waves with significant damping and are also important for the winter–spring EUC (Figs. 12f,g), accounting for \(\sim 40\%\) of the climatological EUC amplitude near the EUC core during February–March. The higher-order modes (modes 9–25) have small \(c_n\) values and thus large friction. They are essentially in “pseudo Ekman balance,” which is the balance among friction, Coriolis force, and surface wind stress and have little contribution to the pressure gradient force and EUC.

The lag correlation between wind stress and EUC shown above suggests that equatorial currents need 1–2 months to set up the pressure gradient force in the equatorial basin. For the first baroclinic mode, it takes the equatorial Kelvin wave \(\sim 1.5\) months and first meridional mode Rossby wave \(\sim 4.5\) months to cross the equatorial Indian Ocean basin. For the second baroclinic mode, it takes the Kelvin wave \(\sim 1.5\) months and first meridional mode Rossby wave \(\sim 4.5\) months to cross the equatorial basin. Given that strong easterly winds occur primarily in the western basin and that directly forced waves are the major cause for the winter–spring EUC core, the time scale for the directly forced waves to affect the western basin EUC is much shorter than the time to cross the entire basin. More importantly the first eight modes contribute to the total EUC velocity (Figs. 12a,b), and the wind–EUC correlation reflects this relationship. As shown in Figs. 12a and 12b, from February to April, it takes the EUC signal (the sum of modes 1–8) less than a month to propagate from the central basin (\(\sim 70^\circ\)E) to the eastern boundary, somewhat more than a month from the central basin to the western boundary, and \(\sim 2\) months to cross the entire equatorial basin (Fig. 12b). This explains the 1–2-month time lag between EUC and surface wind correlations.

c. Dynamics: Summer–fall EUC reappearance

During summer and fall, eastward subsurface flow occurs every year along the Indian Ocean equator (Fig. 13). The EUC (with a subsurface maximum), however, disappears in 2010 but reappears in all other years (section 3). Recall that not all eastward subsurface currents in the depth range 60–140 m can be defined as EUC, since some flow at shallow depths in this range results from the downward extension of surface Wyrtki jets (Figs. 2–4). The eastward subsurface flow generally has a good correspondence with the equatorial easterly wind that appears in the eastern and western basins except for 2010, when westerly wind prevails in this
negative IOD year (Figs. 13a,b, 14a). Warm water accumulation in the eastern basin and the development of zonal temperature gradient (warmer in the east compared to the west) may be a key element that contributed to the strengthening of westerly anomalies (Krishnan et al. 2006). The eastward subsurface current simulated by the LOM has an overall agreement with the HYCOM solution, even though significant differences exist between the two (cf. Figs. 13b,d). Directly forced Kelvin and Rossby waves dominate the currents in the western basin for all years, and they also have significant contributions in the central-eastern basin in 2006, 2007, and 2011 (cf. Figs. 13d and 13e) when positive IOD events occur (Fig. 14a). Strong easterly wind anomalies associated with the IOD events have been suggested to be important in causing the summer–fall reappearance of EUC (section 1). Different from the winter–spring EUC that often reaches its maximum in the western basin (Fig. 9), summer–fall eastward subsurface flow often reaches its maximum in the central-eastern basin. Evidently, reflected Rossby waves from the eastern boundary make larger contribution to current maxima in this region than the directly forced response (Figs. 13d–f).

Here, we investigate further the idea that equatorial winds associated with the IOD affect the EUC during summer and fall. The September–November (SON) mean DMI suggests that the positive IOD events with larger than 0.5 STD occur in 2002, 2006, 2007, and 2011 (Fig. 14a), which is verified by the spatial distributions of SST anomaly (SSTA) for individual years (not shown). For instance, in 2002 obvious positive and negative SSTAs appear in September in the western node and eastern node regions, respectively. These SSTAs correspond to easterly wind anomalies in the central-eastern equatorial Indian Ocean. The positive SSTA in the western node becomes weaker in October, whereas the negative SSTA becomes stronger corresponding to the stronger easterly wind anomaly in the eastern equatorial Indian Ocean (also in November). Even
though the “negative SSTA” obtains its maximum magnitude from 5° to 10°S and never reaches the equator, the area-averaged SSTA in the eastern node subtracted from the SSTA in the western node is ~1 STD of the DMI, and thus we identify 2002 as a “weak IOD” year. The 2011 case is somewhat different. It is characterized by strong easterly wind anomalies along the equator with weak southeasterly wind anomalies in the tropical southeastern basin. Cai et al. (2009) suggested that 2008 is also a positive IOD event. Considering that large easterly wind anomalies occur during June–August but sharply reduce afterward, we identify 2008 as an aborted IOD event (Fig. 14b). Atmospheric intraseasonal disturbances may lead to termination of the IOD event (Rao and Yamagata 2004), which is beyond the scope of this study. A strong negative IOD event occurred in 2010 and a weaker one occurred in 2005.

To reveal the different characteristics of the summer–fall EUC associated with the IOD, we perform composite analyses for zonal currents, SSTA, and wind anomaly during August–October for positive IOD years (2002, 2006, 2007, 2008, and 2011), negative IOD years (2005 and 2010), and normal years (2001, 2003, 2004, and 2009). Note that the aborted IOD of 2008 is included in the positive IOD composite.

A common feature of EUC for each panel (left column of Fig. 15) is the existence of a maximum in the western basin. A striking difference exists between positive IOD and negative IOD years (Figs. 15a,b). During a positive IOD, the EUC extends across the equatorial basin (Fig. 15a). An eastern basin maximum occurs at a shallower core depth than that of the western basin maximum, and the EUC flows below the westward surface current (Fig. 15a) due to forcing by strong easterly wind anomalies, which shoal the thermocline and thus EUC core depth, drive an eastward pressure gradient force and EUC, and force a negative SSTA in the eastern equatorial basin (Fig. 15e). In contrast, during a negative IOD, the EUC disappears in the eastern basin with a strong eastward flow near the surface (Fig. 15b), which is driven by the strong westerly wind anomalies that also induce positive SSTA in the eastern basin (Fig. 15f). The normal years share similar characters with the negative IOD years but with a relatively weak eastward flow near the surface (cf. Figs. 15b and 15c). By performing statistical Monte Carlo tests at each grid of zonal current sections, we find that the mean difference in zonal velocities in the eastern basin (75°–95°E) between the positive IOD years and other years exceeds the 90% significance level. However, zonal currents in the western basin present less difference during these years. Monte Carlo tests also suggest that no obvious difference in zonal currents exists between the negative IOD and normal years. Rao and Yamagata (2004) considered 2003 as an aborted positive IOD event. By examining the wind and EUC for this year, we found that westerly wind anomalies dominate the equatorial Indian Ocean during August–September, and the EUC presents similar features as 2001, 2004, and 2009.

In addition to the IOD, equatorial wind variability associated with Indian monsoon rainfall has also been
suggested to affect the summer–fall EUC of IOD years (Swapna and Krishnan 2008; Krishnan and Swapna 2009). For the period of 1988–2011 when both CCMP winds and NOAA SST data are available, the correlation coefficient is $0.81$ between SON-mean DMI and JJA-mean equatorial zonal wind anomalies and $0.91$ between SON-mean DMI and SON-mean equatorial zonal wind anomalies averaged over $5^\circ S$–$5^\circ N, 50^\circ$–$95^\circ E$, with both correlations exceeding 95% significance. When all months are included, the correlation drops to $0.56$ using monthly DMI and monthly CCMP winds. As a comparison, the correlation between summer monsoon rainfall and equatorial zonal wind anomalies is $0.34$ during June–September and $0.44$ (above 95% significance) for June–August. These results suggest that equatorial easterly wind anomalies associated with strong (weak) Indian summer monsoon could help to intensify (weaken) the zonal wind anomalies and therefore summer–fall EUC associated with IOD events, a result consistent with existing studies (Swapna and Krishnan 2008; Krishnan and Swapna 2009). Interannual variability of equatorial zonal wind over the Indian Ocean, however, is predominantly determined by zonal SST gradients associated with IOD events as shown by their high correlations, with Indian monsoon rainfall playing a less important role.

Following Swapna and Krishnan (2008), we use all-India rainfall anomalies to define strong or weak monsoons. Excess (deficit) monsoons are defined when the June–September all-India rainfall is more (less) than 10% of the long-term climatological normal, and those within 10% of the long-term norm are defined as “normal monsoon years.” For our period of interest from 2001 to 2011, only 2007 is a year of positive IOD and strong monsoon (Fig. 14), which does correspond to a fairly strong eastern basin EUC during summer and fall.

![Image of zonal wind stress](image_url)

**Fig. 13.** As in Fig. 9, but for July–December. The core areas of the EUC in (b) are marked by the word EUC. The year 2010 without EUC mark shows just subsurface eastward flow.
Year 2002 is a positive IOD and weak monsoon year (Fig. 14). The EUC, however, is comparable to that of 2007 with a maximum occurring in the eastern basin (Fig. 13). It shares similar EUC features with the positive IOD and normal-to-strong monsoon years (2006, 2007, 2008, and 2011), except for a weaker but thicker and deeper core in the eastern basin (not shown). This result suggests that the situation for weak monsoon years can be rather complex.

Does an EUC exist in weak monsoon years? Composite analyses of the three weak monsoon years 2002, 2004, and 2009 show that the EUC is not weaker than that of normal years (Figs. 15c,d). The western basin is dominated by easterly wind anomalies associated with weak monsoon (Fig. 15b), which can drive a westward surface flow and an eastward EUC. The EUC can extend to the central-eastern basin, albeit with weaker magnitude compared to positive IOD years.

As the situation for winter–spring EUC, the summer–fall EUC in the LOM is weaker than that in the HYCOM. However, the LOM can reproduce the EUC in the western basin and interannual variability of the EUC in the eastern basin (Fig. 16). For all cases, the western basin EUC maximum results primarily from the directly forced...
Kelvin and Rossby waves (cf. the solid and dashed lines of Fig. 16), whereas the eastern basin EUC is dominated by reflected Rossby waves from the eastern ocean boundary (dashed–dotted lines). It is only during the positive IOD years that direct wind forcing also has significant contribution to the eastern basin EUC (Fig. 16a).

5. Summary and discussion

The EUC is a transient feature in the Indian Ocean, which was thought to regularly occur during spring but reappear in summer–fall only in some years—the years when both positive IOD event and strong Indian monsoon occur. Because of the time and space limitations of observational data, EUC spatial structure and temporal variability are unclear and have not yet been systematically investigated. Combining data from moorings and ocean reanalysis with modeling experiments using HYCOM and LOM, this study provides a systematic investigation on the structure and dynamics of the Indian Ocean EUC, including its seasonal and interannual variability, for the period of 2001–11. We also examined the EUC structures for both spring and summer–fall using ORAS4 reanalysis products for the entire 1958–2011 period.

Consistent with existing studies, the Indian Ocean EUC indeed regularly occurs in each winter–spring, particularly from February to April, and it can exist in the western, central, and eastern equatorial basin. The EUC generally extends from about 60 m to more than 200 m, with a core depth near D20 within the
thermocline. Rossby and Kelvin waves directly forced by winds are the major cause for generating the EUC in western basin, where the EUC core is located for most of the years we studied. Reflected Rossby waves from the eastern boundary, however, also have a significant deterministic role in producing the EUC in the eastern and central basin (Figs. 9, 10). The EUC associated with the equatorial Kelvin and long Rossby waves satisfies equatorial geostrophy, with \( u \approx -(1/\beta)(\partial^2 p/\partial y^2) \) [Fig. 8; Eq. (3b) of section 4b]. Easterly wind excites equatorial Kelvin and Rossby waves and increases the subsurface pressure in the western equatorial basin, which drives an EUC in the western basin [Fig. 11; Eq. (3a) of section 4b]. The EUC strengthens and extends eastward subsequently and finally weakens by the high subsurface pressure in the eastern basin forced by the equatorial westerly winds during April–May. Different from the situation near the surface, where the current is controlled by the first two baroclinic modes, the current in the subsurface layer is affected not only by the low-order modes 1–2, which essentially represent inviscid Kelvin and Rossby waves’ effects, but also by the intermediate modes 3–8. Equatorial waves associated with these intermediate modes are subjected to significant damping, and they contribute \( \sim 40\% \) of the winter-spring EUC core amplitude (Fig. 12).

The EUC often reappears during summer–fall, particularly from August to October, and can also be observed in July (e.g., 2008) and December (e.g., 2006) in some years. Indeed, the summer–fall EUC reappears in all years from 1958 to 2011, except for 2010 in ORAS4 data. In 2010, the summer–fall EUC completely disappeared. By analyzing vertical sections of ORAS4 zonal currents along the equator for individual years, we find that the summer–fall EUC reappears at different regions of the equatorial Indian Ocean. It usually occurs in the western basin but also is present in the central and
eastern basin during positive IOD years (Fig. 15). During the negative IOD and normal years, the EUC does not exist in the eastern basin and only occurs in the western basin. During all years, the western basin EUC is caused primarily by direct, wind-driven Rossby and Kelvin waves. By contrast, the eastern basin EUC during positive IOD years results largely from reflected Rossby waves from the eastern boundary, with directly wind-forced waves also having significant contributions. During negative IOD and normal years, eastward subsurface flow induced by eastern boundary–reflected Rossby waves is canceled by the westward flow directly driven by winds (Fig. 16).

Zonal wind anomalies in the equatorial Indian Ocean associated with strong Indian monsoon rainfall may help to intensify the zonal wind anomalies and thus summer–fall EUC associated with positive IOD events, consistent with existing studies. Note, however, that interannual variability of equatorial zonal wind during summer–fall is predominantly determined by the zonal SST gradients associated with IOD events, with monsoon rainfall variability playing a less important role. The situation for weak monsoon years appears to be more complex. Composite analysis shows that the summer–fall EUC during weak monsoon years can be stronger than that of normal years. This is because the easterly wind anomalies in the western equatorial basin associated with the weak monsoon can drive an eastward pressure gradient force and thus EUC, which can extend eastward to the central and eastern basin (Fig. 15).

From 1948 to 2012, there were 14 weak monsoon years: 1951, 1965, 1966, 1972, 1974, 1979, 1982, 1985, 1986, 1987, 2002, 2004, and 2009. The NCEP wind anomalies based on the 14 yr show easterlies extending across the Indian Ocean, especially in the western basin (west of 65°E; Fig. 17a). By analyzing 54-yr ORAS4 products from 1958 to 2011, we do not find that the EUC weakens significantly or disappears during the last 13 weak monsoon years. The available NOAA SST anomalies for 1951, 1965, 1966, 1968, 1972, 1974, 1979, 1982, 1985, 1986, 1987, 2002, 2004, and 2009 suggest that the SST tends to be higher in the western basin during weak monsoon years (Fig. 17b), which could aid the formation of the stronger easterly wind anomaly in the western basin by creating a westward pressure gradient force through pressure gradient and vertical diffusion mechanisms.

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