The role of Equatorial Undercurrent in sustaining the Eastern Indian Ocean upwelling

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Abstract By combining volume transport and salinity analysis from 1958 to 2014, this paper investigates how the transient Equatorial Undercurrent (EUC) sustains the summer-fall equatorial eastern Indian Ocean (EIO) upwelling. On seasonal time scales, the EIO upwelling is mainly supplied by the salty water from the western basin through a buffering process: The winter-spring EUC carries the salty water from the western basin eastward, induces downwelling in the EIO, and pushes portion of the salty water below the central thermocline, which subsequently upwells to the central thermocline during summer-fall and sustains the EIO upwelling. On interannual time scales, enhanced upwelling occurs during positive Indian Ocean Dipole (+IOD) years. The strong summer-fall EUC associated with the +IOD supplies water for the intensified upwelling. This research provides new knowledge for basin-scale mass and property exchanges associated with the EIO upwelling, contributing to our understanding of three-dimensional ocean circulation and climate variability.

1. Introduction

In the eastern equatorial Pacific and Atlantic cold tongue regions, the quasi-permanent Equatorial Undercurrent (EUC) [e.g., McCreary, 1981] is the main source of persistent upwelling [Wyrtki and Kilonsky, 1984; Hormann and Brandt, 2007; Qin et al., 2015]. By contrast, in the Indian Ocean the EUC is a transient feature [e.g., Schott and McCreary, 2001; Chen et al., 2015a], and upwelling in the equatorial eastern Indian Ocean (EIO) is a seasonal phenomenon [e.g., Susanto et al., 2001] (Figure 1). On seasonal time scales, the equatorial EIO upwelling primarily occurs during boreal summer and fall (July–October [Chen et al., 2016]), which is mainly caused by the local southeast monsoon wind with significant contribution from equatorial wind forcing [Chen et al., 2016]. On interannual time scales, the EIO upwelling is associated with the Indian Ocean Dipole (IOD [Saji et al., 1999; Webster et al., 1999; Murtugudde et al., 2000; Yu et al., 2005; Nyadjo and McPhaden, 2014; Chen et al., 2016]) and El Niño–Southern Oscillation events [e.g., Susanto et al., 2001]. Sea surface temperature (SST) in the EIO upwelling region also exhibits evident intraseasonal variability [e.g., Vinayachandran and Saji, 2008; Shinoda et al., 1998; Qiu et al., 1999; Chen et al., 2015b]. Hereafter, seasons refer to those of the Northern Hemisphere.

On seasonal time scales, the EUC exists across the equatorial Indian Ocean basin near ~60–160 m depth during winter and spring, particularly from February to April, and it reappears during late summer to early fall (August–September) with a much weaker magnitude [e.g., Schott and McCreary, 2001; Chen et al., 2015a]. The Indian Ocean EUC is defined as an eastward flow with core located in the thermocline and beneath a westward or weaker eastward flowing surface current [Chen et al., 2015a]. Strong summer-fall EUC in the central and eastern basin only appears during positive IOD years [e.g., Reppin et al., 1999; Iskandar et al., 2009; Sengupta et al., 2007; Han et al., 2004; Swapna and Krishnan, 2008; Krishnan and Swapna, 2009; Zhang et al., 2014; Chen et al., 2015a]. Given that the EUC is a transient feature and EIO upwelling only occurs during summer-fall season, their relationship—particularly, the role played by EUC in sustaining the EIO upwelling—is unclear.

The western equatorial Indian Ocean is occupied by salty water from the Arabian Sea, whereas the EIO is occupied by fresher water due to strong precipitation over the warm pool and fresher water from the Bay of Bengal and the Indonesian throughflow [Han and McCreary, 2001; Qu et al., 2008] (Figure S1 in the supporting information). While the spring and fall Wyrtki Jets (WJs) [Wyrtki, 1973] transport the saltier water from the western basin to the east near the surface [Schott et al., 2009; McPhaden et al., 2015], whether or not the EUC can bring the saltier Arabian Sea water to the EIO in the subsurface to sustain the EIO upwelling remains unknown.

The goal of this paper is to understand what role the intermittent EUC plays in sustaining the EIO upwelling and how it provides salty water for the upwelling. Scientifically, addressing this issue will substantially advance our
understanding of the EIO upwelling and the role of the EUC in facilitating mass and salt exchanges between the western and eastern equatorial Indian Ocean. The research is a crucial step toward fully understanding the three-dimensional structure of the wind-driven shallow meridional overturning circulation, which links the subtropical subduction to the tropical upwelling [Schott et al., 2009]. Climatically, the research is important because the EIO upwelling occurs in the warm pool region, where the mean SST is high, and thus, atmospheric convection is more sensitive to SST changes [e.g., Sardeshmukh and Hoskins, 1988; Webster and Lukas, 1992; Yu et al., 2002; Wang and Mehta, 2008; Izumo et al., 2010; Kim et al., 2012]. The research will also contribute to our understanding of biological activities, which show considerable variability in the EIO upwelling region [e.g., Grumet et al., 2004; Varela et al., 2015].

2. Data and Method

Temperature, salinity, and current data from the European Centre for Medium-Range Weather Forecasting (ECMWF) Ocean Reanalysis System version 4 (ORAS4) [Balmaseda et al., 2013] during 1958–2014 with a horizontal resolution of 1° × 1° and 42 vertical levels are used to examine the EIO upwelling and its relationship with the EUC. Its vertical resolution is ~10 m in the upper 100 m, gradually increases to ~15 m at 150 m depth and ~20 m at 200 m depth. The ORAS4 data successfully capture the seasonal-to-interannual variability of the EIO upwelling [Chen et al., 2015b], EUC [Nyadjro and McPhaden, 2014; Chen et al., 2015a], and salinity variability in the equatorial Indian Ocean (Figures S2 and S3), and thus are suitable for our investigation. According to Australian Government Bureau of Meteorology (http://www.bom.gov.au/climate/IOD/positive/), positive IOD years from 1958 to 2007 are 1961, 1963, 1967, 1972, 1977, 1982, 1983, 1994, 1997, 2006, and 2007. Year 2011 is also identified as positive IOD [Chen et al., 2015a].

To investigate the role played by the EUC in sustaining the EIO upwelling, we define the EUC transport into the EIO as the zonal transport between $\sigma_\theta = 23.5$ and 25.5 kg m$^{-3}$ vertically and 2°S–2°N meridionally and averaged for 80°E–90°E. The ORAS4 data successfully capture the seasonal-to-interannual variability of the EIO upwelling [Chen et al., 2015b], EUC [Nyadjro and McPhaden, 2014; Chen et al., 2015a], and salinity variability in the equatorial Indian Ocean (Figures S2 and S3), and thus are suitable for our investigation. According to Australian Government Bureau of Meteorology (http://www.bom.gov.au/climate/IOD/positive/), positive IOD years from 1958 to 2007 are 1961, 1963, 1967, 1972, 1977, 1982, 1983, 1994, 1997, 2006, and 2007. Year 2011 is also identified as positive IOD [Chen et al., 2015a].

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Figure 1. Sketches for the winter-spring EUC to sustain the EIO summer-fall upwelling. The color in the sea surface represents SSTA, and the blue means negative SSTA. January–April currents using ORAS4 data from 1958 to 2014 are shown to represent surface currents and winter-spring EUC (vectors). The red box (95°E–114°E, 10°S–0°N) shows the upwelling area. The black line is used to show vertical sections of salinity and velocity in Figure 2, with letters A, B, and C showing the turns and end point. Their longitudes are 98.5°E, 105°E, and 114°E, respectively. The scale for EUC is 0.2 m s$^{-1}$. 

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transport is based on this definition. The EUC transport is also tested for fixed depths (e.g., at 60–160 m; horizontal black lines in Figure 2), and it is highly correlated with that between $\sigma_\theta = 23.5$ and $\sigma_\theta = 25.5$ kg m$^{-3}$ ($r = 0.92$) but has larger variability amplitude (figure not shown), because it may contain the bottom part of the WJs in some years (Figures 2b and 2d). The spring and fall WJs have peak transports of $14.9 \pm 2.9$ sverdrup (Sv) in May and $19.7 \pm 2.4$ Sv in November from August 2008 to December 2013 [McPhaden et al., 2015].

To quantify the EIO upwelling, we choose our study region as (95°E–114°E, 10°S–0°N) south of Sumatra and Java (red box in Figure 1). This area covers the prominent negative SST anomaly (SSTA) and high chlorophyll concentration corresponding to seasonal and interannual upwelling. For further justification of the boxed region please see Chen et al. [2016]. To calculate the volume transport associated with the EIO upwelling, we choose the top and bottom of the box at 60 m and 160 m, a depth range that contains the main thermocline and the EUC (Figure S4). Note that to measure the EIO upwelling, calculating vertical transport at a specified depth instead of a specified isopycnel is needed, because when isopycnals outcrop into the surface mixed layer during upwelling, their vertical velocities cannot be used to represent “upwelling” due to strong mixing in the mixed layer. To examine the relationship between upwelling and horizontal convergence, we also calculate the horizontal volume transport convergence into the box region integrated from 60 m to 160 m. It is calculated

Figure 2. Vertical sections along the Indian Ocean equator and along Sumatra and Java coasts for monthly climatology of salinity (psu) and zonal current (m s$^{-1}$) from 1958 to 2014 using ORAS4 data (black line in Figure 1). The red curves in zonal current plots show $\sigma_\theta = 23.5$ and $\sigma_\theta = 25.5$ kg m$^{-3}$ lines, and the black dashed horizontal lines represent 60 m and 160 m depths. The arrows on salinity panels show zonal current, with the white arrows eastward and the blue arrows westward. The two vertical dashed lines show positions of turns A and B labeled in Figure 1.
as the sum of zonal transport difference at the west and east boundaries (west minus east) and meridional transport difference at the south and north boundaries (south minus north): $T_{95°E} - T_{114°E} + T_{10°S} - T_{10°N}$.

3. Results

Variability (seasonal plus interannual) of the horizontal volume transport convergence into the EIO (red box of Figure 1) for 60–160 m depths is dominated by the zonal transport at the west boundary of the box, with standard deviations (STD) being 3.8 Sv for the box area and 4.1 Sv for the west boundary. The correlation between the two is 0.84 (>95% significance). The transport convergence is highly correlated with the EUC transport, with correlation coefficient being $r = 0.75$ and STD being 3.8 Sv for the convergence and 3.6 Sv for the EUC (Figure 3a). Similar to the monthly time series, interannual anomalies of EUC transport and horizontal transport convergence are also highly corrected with $r = 0.71$, and their STDs are 0.97 Sv and 0.81 Sv (figure not shown), respectively. These results indicate that the horizontal transport convergence results largely from the equatorial transport in the thermocline layer. Their correlation reaches $r = 0.89$ using July–October mean data when only interannual variability of upwelling is considered and summer-fall EUC appears in the eastern basin particularly during positive IOD years.

When all seasons are considered, the anomalies of horizontal transport convergence into the EIO are negatively correlated with the anomalies of EIO upwelling—measured by vertical volume transport—when monthly data that contain seasonal-to-interannual variability are used, with correlation coefficients being $-0.43$, $-0.81$, and $-0.92$ (>95% significance), respectively, with upwelling at 60 m (UP60), 160 m (UP160), and their difference (UP160 – UP60) (Table 1). Upwelling anomaly at 160 m has larger amplitude with STD of 2.7 Sv, which is 41% higher than the 1.6 Sv STD at 60 m. These results demonstrate that the horizontal convergence (divergence) related to currents from 60 to 160 m depths results in downwelling (upwelling) in the EIO above 160 m. If only the upwelling season during July–October is considered, however, the horizontal convergence anomaly is positively correlated with interannual anomaly of upwelling at 60 m with a correlation coefficient of $r = 0.60$ (>95% significance), and negatively correlated with upwelling anomaly at 160 m with a weak correlation of $r = -0.31$ (Table 1), suggesting that on interannual time scales, horizontal transport convergence into the EIO in the main thermocline layer during July–October provides water source for upwelling into the surface mixed layer near 60 m, and to a lesser degree it leads to downwelling near 160 m. Note that positive correlation between transport convergence and upwelling at 60 m also appears during January–March (Table 1), but with lower correlation ($r = 0.45$) owing to the weaker upwelling that is induced by the remote equatorial winds in this season [Chen et al., 2016].

Similar to the horizontal transport convergence, interannual anomaly of the EUC transport is positively correlated with upwelling anomaly at 60 m with $r = 0.62$ (>95% significance) and weakly correlated with upwelling anomaly at 160 m with $r = -0.22$ during July–October (Table 1). The STD of July–October UP60 is 0.6 Sv and that of UP160 is 0.5 Sv. The above results suggest that the anomalous EUC transport dominates the anomalous horizontal transport convergence into the EIO, playing an important role in sustaining the EIO upwelling. Below, we first discuss seasonal upwelling and then its interannual variability.

On seasonal time scales, summer-fall upwelling exists in the EIO during June–October [Susanto et al., 2001; Chen et al., 2016] but only weak EUC exists during August–September [Chen et al., 2015a] (Figure 2d). What is the water source that sustains this seasonal upwelling in the EIO? To answer this question, we obtain the climatological seasonal cycle for the 1958–2014 period. Upwelling over the EIO exhibits clear semiannual variability at both 60 m and 160 m depths (red and black curves of Figure 3b), with peaks occurring in summer-fall (particularly from June to September) and winter (December to February). The strength of the upwelling is stronger at 160 m than at 60 m for both seasons. Given that cold SST and increased chlorophyll concentration associated with upwelling only occur during summer-fall and not in winter [Chen et al., 2016], here we focus only on the summer-fall upwelling. During June–August, seasonal upwelling at 60 m is primarily supplied by upwelling at 160 m, a season when no strong eastward EUC exists (Figure 2d; compare the red and black curves with the pink and blue curves in Figure 3b). In September, upwelling weakens (Figure 3b) and downwelling starts to occur at 160 m ($-0.1$ Sv); meanwhile, the EUC (pink curve) induces convergence (blue curve) in the EIO, which supplies water for the weak upwelling at 60 m.

An intriguing question is as follows: What is the water source for upwelling at 160 m during June–August? Since positive EUC transport and horizontal transport convergence into the EIO generally lead to downwelling near
Figure 3. (a) Time series of monthly EUC transport (ET, red curve) (see section 2 for definition) and Horizontal Volume Transport Convergence (HVT) integrated for 60–160 m into the box of Figure 1 from 1958 to 2014. (b) Monthly climatological vertical volume transport associated with upwelling at 60 m and 160 m depths (UP60 and UP160), together with HVT into the EIO and ET from 1958 to 2014 using ORAS4 data. (c) Same as Figure 3b but for positive IOD (PIOD) years. (d) Interannual anomalies (with the 1958–2014 climatology removed) of July–October mean vertical volume transport associated with UP60 (dotted red), UP160 (dotted black), together with HVT (blue) and ET (red) from 1958 to 2013. The vertical lines mark the positive IOD year. Note that 1987 was regarded as a moderate PIOD by some previous studies [e.g., Cai et al., 2014]. (e) Monthly climatological salinity anomaly (with the annual mean of all years removed) averaged for 60–160 m in the EIO (box of Figure 1) from 1958 to 2014 using ORAS4 data. (f) Same as Figure 3e but for 160–260 m. (g) Time evolution of monthly climatological salinity distribution in the upper 300 m at (0°N, 95°E), the west boundary of the EIO box. The white dashed horizontal lines are 60 m, 160 m, and 260 m depths. The vertical bars in Figures 3b, 3c, 3e, and 3f show confidence intervals for the true population mean at 95% significance level obtained by t test. The white arrows in Figure 3g indicate downwelling and upwelling.
160 m (e.g., from February to April; Figure 3a and Table 1), we test the following “buffering” hypothesis. The winter-spring EUC, which exists across the equatorial Indian Ocean with large magnitude occurring in February–April (Figure 2b), carries the salty water from the western basin eastward (Figures 2a and 2b) and induces downwelling in the EIO, pushing portion of the salty water downward out of the central thermocline to below 160 m. After which the spring WJ appears but cannot extend down to 160 m depth (Figure 2b). Due to weak horizontal circulation, the salty water stays there until summer, when seasonal upwelling occurs and the downwelled salty water below 160 m upwells to the central thermocline (60–160 m), which subsequently provides water for upwelling into the surface mixed layer near 60 m.

This buffering process and the impact of EUC on the EIO seasonal upwelling can be further verified by salinity variations. The seasonal salinity anomaly over the EIO shows an evident annual cycle within the central thermocline, with high salinity during summer-fall upwelling season and low salinity for the rest of the year (Figure 3e). Since salinity in the EIO shows less variation below \(\sigma_\theta \approx 26.5 \text{ mg m}^{-3}\) (~260 m; Figure S5), we choose another box for 160–260 m depths with horizontal area being the same as the red box of Figure 1. In contrast to the annual cycle of the central thermocline layer (60–160 m), salinity in this deeper box exhibits clear semiannual variability, with larger amplitudes occurring during April–June and July–September (Figure 3f).

The strong positive salinity anomaly during April–June and weaker positive salinity anomaly during October–December cannot be attributed to the spring and fall WJs, because the WJs cannot extend down to the 160–260 m depths, fresher water (~33.0 practical salinity unit (psu)) dominates the mixed layer in the EIO, and the sole salinity maximum in the EIO occurs in the central thermocline (Figure 2). Further analysis suggests that the positive salinity anomalies below 160 m over the EIO during spring and fall (Figure 3f) are associated with variations of the EUC, and the positive anomaly during October–December is induced by the stronger summer-fall EUC during positive IOD years, which occupies the eastern basin until December (Figure S6). If positive IOD years are excluded, negative salinity anomaly at 160–260 m dominates October–December (Figure S7) for nonpositive IOD years, which are ~80% of the 57 years during 1958–2014.

When the winter-spring EUC carries salty water from the western basin to the EIO during February–April (Figure 2a; pink curve of Figure 3b), downwelling related to the EUC induces high-salinity anomaly in the deeper layer during April–June (Figures 3f and 3g) because of high-salinity input at 160 m (about \(\sigma_\theta = 25.5 \text{ mg m}^{-3}\)) and low-salinity output at 260 m (Figures S5a and S5b). Subsequently, the salty water in the deeper box upwells to the central thermocline during the summer-fall upwelling (Figure 3b), and negative salinity anomaly appears in the deeper layer (Figures 3f and 3g), while positive anomaly appears in the thermocline layer (Figures 3e and 3g).

During positive IOD years, intensified upwelling occurs in the EIO during summer-fall, and strong eastward EUC appears in the central and eastern basin [Nyadjro and McPhaden, 2014; Chen et al., 2015a] (pink curve of Figure 3c and Figure S6). Is the EUC sufficient to provide water source for the enhanced EIO upwelling during these years? Due to the fact that the EIO seasonal upwelling occurs earlier than the strong summer-fall EUC associated with the positive IOD, the buffering effect still accounts for seasonal upwelling during June–July and contributes partly in August (Figure 3c). The strong summer-fall EUC associated with the IOD, however, can supply part of the upwelling water into the surface mixed layer in August and most of the upwelling

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<th>Table 1. Correlation Coefficients Between Horizontal Volume Transport Convergence/Divergence (HVT) Integrated for 60–160 m and Vertical Volume Transport Due to Upwelling/Downwelling in the EIO Box Region Shown in Figure 1a.</th>
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* Positive (negative) values indicate volume convergence into (divergence from) the EIO and upwelling (downwelling).
water for September–October; meanwhile, the EUC induces downwelling at the thermocline bottom during September–November (Figure 3c). While the summer-fall EUC transport is not sufficient to sustain the total strong upwelling (seasonal upwelling + positive interannual anomaly) for positive IOD events (labeled by dashed vertical lines in Figure 3d), it is sufficient to provide water source for sustaining the intensified interannual upwelling anomaly (compare the solid red and blue with the dotted red curves in Figure 3d). Composite analysis for positive IOD years is conducted and the results clearly support this point (Figure S8). For nonpositive IOD years including negative IOD years, the situation is similar to the climatological seasonal upwelling discussed above (Figure S9).

4. Summary and Discussion

Using ORAS4 data from 1958 to 2014, we examine the role played by the intermittent EUC in sustaining the seasonal and interannual EIO upwelling. On seasonal time scales, the winter-spring EUC during February–April carries the salty water from the western basin eastward in the central thermocline layer (Figure 2), induces downwelling in the EIO and pushes part of the salty water downward to the depth below ~160 m (Figures 2 and 3). The salty water (positive salinity anomaly) stays there until summer, and then upwelling brings the salty water to the central thermocline, which sustains upwelling into the surface mixed layer during summer-fall (Figure 3). As a result, salinity within the central thermocline over the EIO shows an annual cycle, with positive salinity anomaly occurring during summer-fall upwelling season. These results are summarized in Figure 1. On interannual time scales, enhanced upwelling occurs during positive IOD years (Figures 3c and 3d). The intensified summer-fall EUC associated with the positive IOD events transports sufficient amount of salty water from the western basin eastward, which significantly increases the EIO salinity in the central thermocline and sustains the enhanced upwelling anomaly (Figure 3d). Nyadro and McPhaden [2014] suggested that positive EUC transport anomalies appear in years following negative IOD events because of equatorial wave processes. Figure S10 shows that the enhanced EUC transport helps to sustain the weak interannual upwelling anomaly during summer-fall of those years.

Note that the salty Arabian Sea water from the western equatorial basin through the “buffering effect” (section 3) may not be the only water source that sustains the EIO seasonal upwelling. Waters from other sources, such as the fresher Indonesian throughflow and outflow from the Bay of Bengal, may also contribute. Since the mid-1990s, an obvious freshening trend occurs in the southeastern tropical Indian Ocean [Du et al., 2015; Llovel and Lee, 2015], which may lead to freshening in the upwelling region. However, the fact that the upwelled water is “salty” strongly suggests that the salty Arabian Sea water certainly is an important source. Carefully designed experiments using high-resolution ocean general circulation models in the future will help to decipher the possible water sources that sustain the EIO upwelling. This study takes a crucial step forward toward a thorough understanding of the three-dimensional structure of the Indian Ocean shallow meridional overturning circulation, which plays an important role for maintaining the Indian Ocean heat and salt balance.

Acknowledgments

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