Weakening of spring Wyrtki jets in the Indian Ocean during 2006–2011

Sudheer Joseph,[1] Alan J. Wallcraft,[2] Tommy G. Jensen,[2] M. Ravichandran,[1] S. S. C. Shenoi,[1] and Shailesh Nayak[3]

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[1] Beginning in 2006, the Indian Ocean experienced climatologically anomalous conditions due to large-scale coupled air-sea interactions that influenced the surface circulation of the equatorial Indian Ocean. Here we present evidence from observations as well as a general circulation model to demonstrate that spring Wyrtki jets (WJ) were weak during the past 6 years and were even reversed to westward flow during 2008. We note that this weakening coincided with uniformly high sea level as well as positive east to west gradient anomalies along the equatorial Indian Ocean during the month of May each year, starting in 2006. The weakened jets occur in conjunction with the latitude of zero zonal wind (LUZ) being close to the equator during these years, resulting in weaker than normal zonal winds along the equator from 2006 and onward. We find that starting in 2006, the normal tendency of westward propagation of the annual harmonic mode switches to eastward propagation, coherent with the wind forcing. In comparison to the annual harmonic component of the zonal current, the weak WJs are mainly associated with the semiannual harmonic WJs, as evident from an amplitude reduction of that mode by at least 0.3 m s^{-1} during the post-2005 period. Our analysis demonstrates that the variance explained by the semiannual harmonic is reduced to half (30–40%) at the core of the WJ in 2006 and later years in comparison with earlier years when it was 70–80%.

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1. Introduction

[2] Since *Wyrtki* [1973] first discovered the narrow jet-like current that flows eastward during the intermonsoon periods in the equatorial Indian Ocean, numerous observational [*Rao et al.*, 1989; *Reppin et al.*, 1999; *Qiu et al.*, 2009] and modeling efforts [*O'Brien and Hurlburt*, 1974; *Jensen*, 1993; *McCreary et al.*, 1993; *Han et al.*, 1999; *Nagura and McPhaden*, 2010a, 2010b] have been made to better our understanding of this unique feature of the Indian Ocean. The wind drives the surface circulation of the Indian Ocean in tandem with equatorial Kelvin and Rossby waves. The surface currents in the equatorial Indian Ocean reverse direction four times a year, with a weak westward flow during winter and summer and a strong eastward flow during April-May and October-November. The strong eastward currents are now widely referred to as Wyrtki jets (WJs) [*Wyrtki*, 1973; *Schott and McCreary*, 2001]. During boreal summer (June to September) the southwest monsoon provides strong southwesterly winds across the equator and over the northern Indian Ocean. During boreal winter (December-January-February) surface winds blow from south Asia and cross the equator to meet the south Indian Ocean trades [e.g., *McCreary et al.*, 1993; *Schott and McCreary*, 2001; *Hastenrath and Polzin*, 2004]. During spring, from west to east, the mixed layer deepens, sea level rises, and the sea surface temperature (SST) increases [*Hastenrath et al.*, 1993; *McCreary et al.*, 1993]. The maximum speed of the WJs occurs between 65°E and 85°E, and the jets cause upper ocean divergence and upwelling in the west and convergence with downwelling in the east. Throughout the year, except during the boreal winter monsoon season, the westerly surface wind stress drives equatorward Ekman transports in both hemispheres, which cause downwelling along the equator [*Hastenrath et al.*, 1993; *Schott and McCreary*, 2001]. It was suggested [*Wyrtki*, 1973] that the eastward jets were directly forced by the equatorial westerlies during the intermonsoon period. This was confirmed by *O'Brien and Hurlburt* [1974] in their modeling effort using switched on winds, with an additional insight that the eastward jet gets canceled under the influence of reflected Rossby waves from the eastern boundary within a period of about 2 months after their genesis under the influence of westerlies. The strong semiannual jet-like response is a consequence of a basin resonance, theoretically formulated by *Cane and Moore* [1981] and applied to the

[1]Indian National Centre for Ocean Information Services, Ministry of Earth Sciences, Hyderabad, India.
[2]Oceanography Division, Naval Research Laboratory, Stennis Space Center, Mississippi, USA.
[3]Ministry of Earth Sciences, New Delhi, India.

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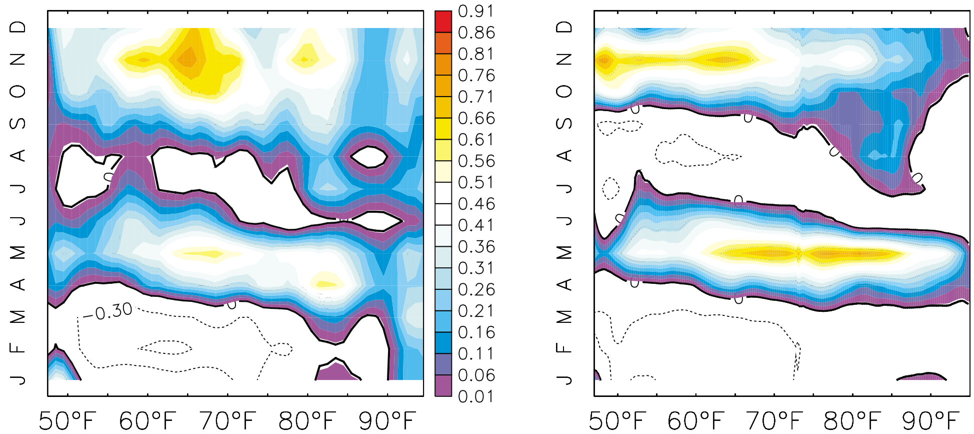

Figure 1. Climatology of zonal currents in the Indian Ocean from OSCAR. (left) Currents averaged from 1993 to 2005 and (right) the solution in year 10 from the climatological HYCOM run. Units are m s^{-1}.

Indian Ocean by *Jensen* [1993] to explain why the semiannual mode dominates the annual mode. The resonance is enhanced by shallow mixed layer depths, in particular during the fall [*Han et al.*, 1999].

[3] The WJs during spring and fall are comparable in magnitude [*Jensen*, 1993; *Han et al.*, 1999]. However, there is a general agreement in the literature that the fall jet is faster and more intense than the spring jet current. Current speeds reported in the literature vary in the range 0.5–0.9 m s^{-1} for the spring jet and from 0.75 to 1.2 m s^{-1} for the fall jet [*Knox*, 1976; *McPhaden*, 1982; *Han et al.*, 1999].

[4] In view of the unique status of Wyrtki jets in the equatorial circulation of the Indian Ocean, it is important to monitor their variability on different timescales, which will help in assessing the influence they may have on the ocean circulation in general and their impact on the climate at least on a regional scale. In this paper we present the variability of equatorial Wyrtki jets using currents from the Hybrid Coordinate Ocean Model (HYCOM) and from the Ocean Surface Current Analyses Real Time (OSCAR) for the period 2003 to 2011, and specifically focus on the anomalous behavior of the WJs from 2006 to 2011 when the spring WJs were unusually weak. We support our findings using current data from the Research Moored Array for African-Asian-Australian Analyses and Prediction (RAMA) moorings [*McPhaden et al.*, 2009] and sea level anomalies from Archiving, Validation and Interpretation of Satellite Oceanographic Data (AVISO).

2. Data

[5] For obtaining the equatorial zonal currents, we used the OSCAR analyses, which represents the average currents from the surface to 15 m [*Bonjean and Lagerloef*, 2002] available as 5 day averages on a 1° × 1° grid from http://www.oscar.noaa.gov/index.html. These currents are obtained by a linear combination of geostrophic components computed from altimeter and Ekman components derived from scatterometer winds. We further confirm our results using currents from HYCOM [*Bleck*, 2002] configured for the Indian Ocean (20°E to 125°E and 35°S to 31°N). This model has a hybrid vertical coordinate, which is isopycnal in the open, interior and stratified ocean, while using a z level coordinate in the mixed layer. It combines the advantages of isopycnic coordinates and z level coordinates in a unique way to improve the simulations. The present configuration is with a 0.25° × 0.25° horizontal resolution, 28 hybrid layers in the vertical, and a nonlocal K profile parameterization (KPP) is used for the boundary layer mixing scheme [*Large et al.*, 1994]. It has river run off included using the Naval Research Laboratory's monthly climatology [*Barron and Smedstad*, 2002]. The model thermohaline fields are initialized with Levitus climatology [*Antonov et al.*, 1998] and the General Bathymetric Chart of the Oceans (GEBCO) data is used for bottom topography. The model is relaxed to climatology at its southern and eastern boundaries. The model was spun up for 10 years using climatological forcing and subsequently run interannually from 2003 to June 2011 using 3 hourly atmospheric fields from the Navy Operational Global Atmospheric Prediction System (NOGAPS) available from 2003 and onward. For consistency with the OSCAR currents, HYCOM currents averaged between the surface and 15 m are used in our analysis. Since NOGAPS winds were used to force HYCOM, and QuikSCAT winds for the period 2000–2009 (until early November 2009, when the scatterometer failed) were used for OSCAR, we compared the two wind products and found they were in close agreement. Zonal currents on the equator from HYCOM were compared with daily 10 m current observations from the RAMA mooring at 90°E during the period from October 2006 until June 2011. With the exception of an extended period from November 2009 to July 2010, RAMA buoy observations were available for all years except for a few gaps of up to one month. HYCOM sea level anomalies were validated using the altimeter sea level anomalies from a delayed mode merged altimeter data product available from AVISO. However, for the month of May 2011, we relied on the near real-time product of AVISO altimetry. For comparison of the HYCOM model with observations in general, we refer to the validation reports by *Metzger et al.* [2008, 2010], which have shown that the model performs well on seasonal and interannual timescales.

3. Weakening of Spring Wyrtki Jets

[6] Our analysis of HYCOM currents from the tenth year of the climatological run and of the OSCAR average currents corroborates previous studies and demonstrates that

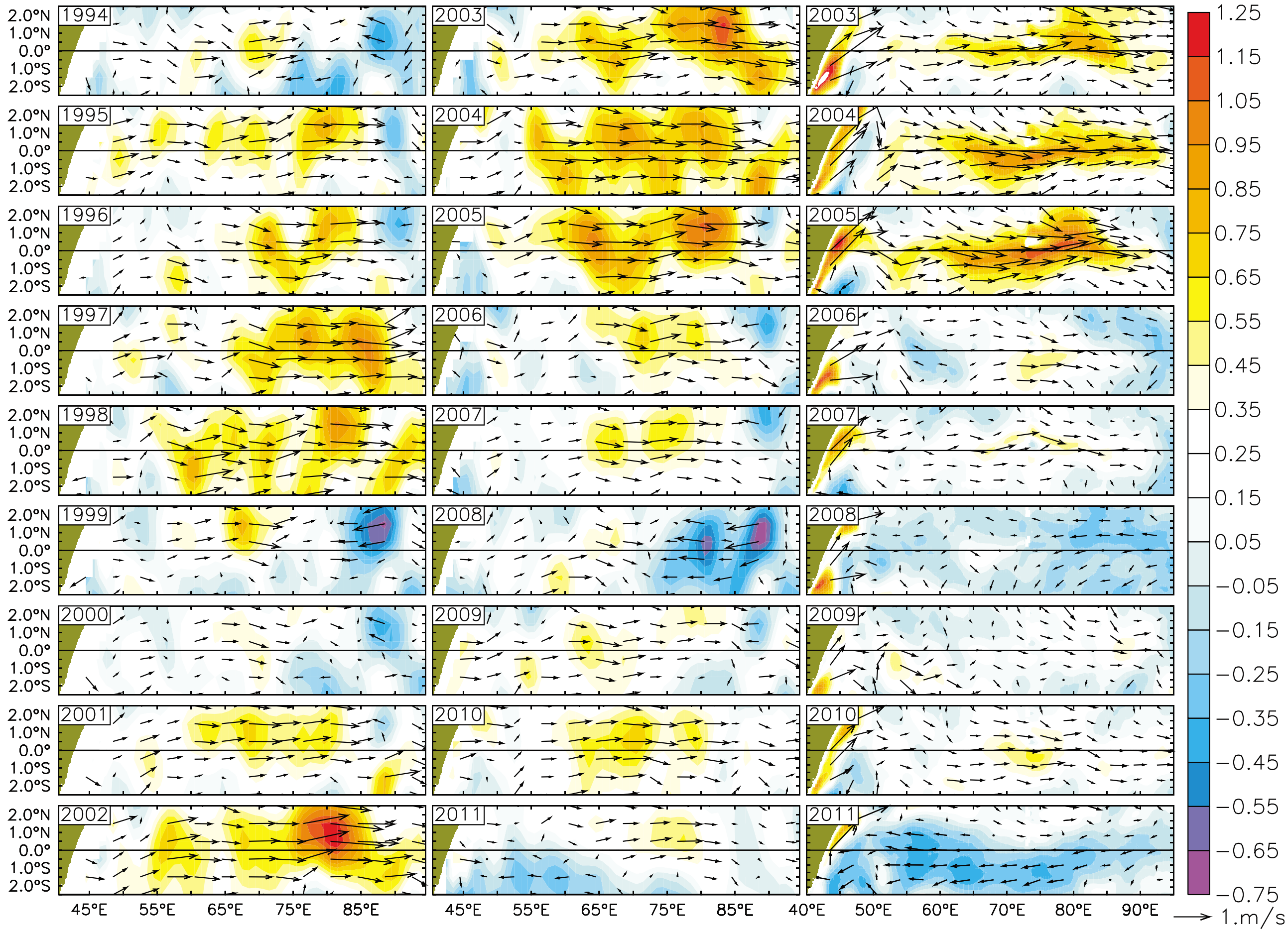

Figure 2. Interannual variability in spring WJs using monthly averaged current vectors for May (left and middle) from OSCAR during 1994–2011 and (right) from HYCOM during 2003–2011. The color shows the magnitude of the zonal component in m s^{-1}.

the fall jet current is faster than the spring jet (Figure 1). The HYCOM solution is showing slightly higher speeds and narrower jets than the OSCAR analysis, which is anticipated given the much higher model resolution (Figure 1).

[7] Compared to climatology, the May mean currents from OSCAR for the years 1994 to 2011, presented in Figure 2 show the characteristics of typical WJs with a maximum speed around 0.9 m s^{-1} for most of the years. Except for weak jets during 1994, 1999 and 2000, average or stronger than average WJs occur until 2006 when a prolonged period of weak spring WJs began, so far extending up to 2011. In the years 2002–2005 WJs were stronger than average in both HYCOM and OSCAR data with WJs covering a wider latitude range in the OSCAR data, compared to model currents (Figure 2, right). In contrast, in both model and OSCAR zonal currents, WJs are much weaker in spatial extent and intensity, appearing as areas with speeds in the range from 0.1 to 0.4 m s^{-1} during 2006 and 2007. In 2008, the spring WJ reversed direction to westward flow: A velocity of 0.2 m s^{-1} toward west occurs in model currents between 70°E and 90°E, while the westward speed is above 0.7 m s^{-1} in OSCAR currents for the same meridians. There are also other subtle differences among model and OSCAR data. For example, OSCAR currents show a slightly larger area of positive zonal current anomalies in 2009 and 2010 when compared to the HYCOM currents. The model shows a weak WJ with speed ranging from 0.1 to 0.4 m s^{-1} for those 2 years, while OSCAR data has currents up to 0.5 m s^{-1} at its core for the same period. For 2011 the model currents show a major westward component just south of equator from 40°E to 90°E, whereas OSCAR has the major westward component limited to 70°E. However, in both model and OSCAR the spring WJs are weaker than in 2010. Irrespective of the differences in details, both data sets show a general weakening starting in 2006 compared to the years 2003–2005 with strong spring jets as well as to the climatological mean spring WJs.

[8] Another significant feature in the HYCOM results is that a branch of the Somali Current, which acts as a source for the WJ, also is weaker in 2006 and in later years (Figure 2, right). Western boundary currents are not included in OSCAR as they cannot be computed accurately from altimetry near the coast.

4. Comparison With RAMA Buoy Currents

[9] In order to further confirm the validity of HYCOM and OSCAR currents, we have compared 10 m depth current meter records from RAMA mooring on the equator with

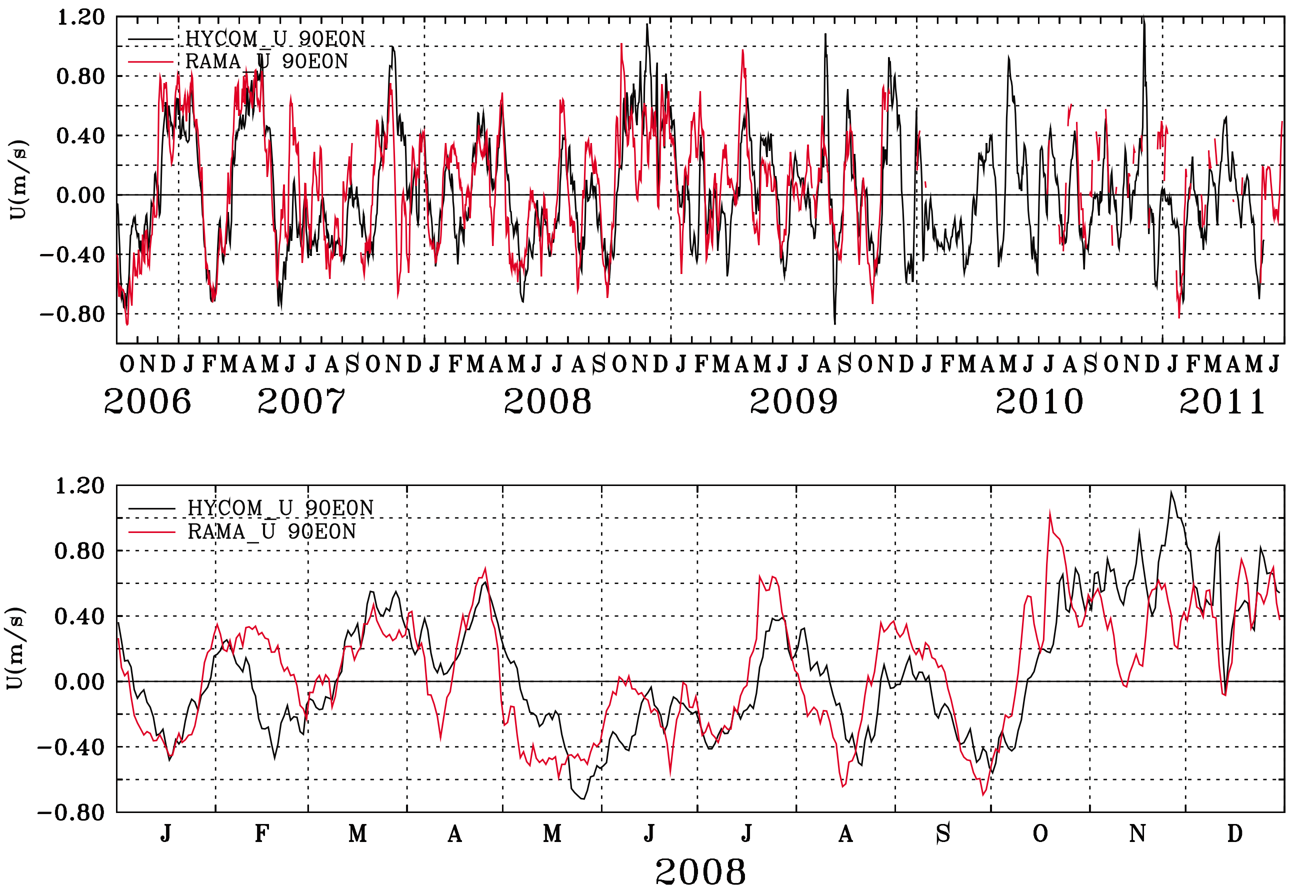

Figure 3. (top) Comparison of 10 m zonal currents from HYCOM (black) with zonal currents from the RAMA mooring on the equator at 90°E (red) from October 2006 to October 2011. (bottom) The zonal currents during 2008 in greater detail.

10 m HYCOM currents. As seen in Figure 3, the buoy observations and model currents are in reasonably good agreement, although RAMA currents in general are somewhat stronger than the model currents. The observed currents from RAMA moorings are clearly westward for the month of May in 2008, which is in agreement with the model and OSCAR currents, and provides a direct confirmation of that unusual event. The weak WJs during May in 2006 and later years (Figure 2) are also present in the currents from the RAMA buoys (Figure 3).

[10] In the RAMA buoy records the zonal currents in April are somewhat stronger than the May currents in some years, i.e., 2007–2009 and 2011. The HYCOM results for April (not shown) reveal that the area of eastward currents are limited to the region 65°E–90°E rather than basin wide.

5. Dynamics of Wyrtki Jet Variability

[11] There is ample evidence from the literature that the WJs are primarily wind driven as their genesis is coherent with the biannual westerlies along equatorial Indian Ocean [*Wyrtki*, 1973; *Jensen*, 1993; *Han et al.*, 1999; *Qiu et al.*, 2009; *Nagura and McPhaden*, 2010a]. There have been studies which found a weakening of the Wyrtki jet during the 1997–1998 positive Indian Ocean Dipole (IOD) event [*Vinayachandran et al.*, 1999; *Murtugudde et al.*, 2000; *Rao et al.*, 2002; *Vinayachandran et al.*, 2002; *Thompson et al.*, 2006; *Subrahmanyam et al.*, 2011] but which emphasized the fall jet. A few other studies related IOD to a weakening of spring jets or fall jets [*Yamagata et al.*, 1996; *Tozuka et al.*, 2007; *Chu*, 2010]. While our analysis shows that spring jets were weak during 1994, in agreement with *Yamagata et al.* [1996], we found no weakening of spring jets during the major IOD event of 1997–1998, which is in disagreement with the findings by *Chu* [2010]. However, we do find a considerable weakening in fall jets in 1997–1998, which has been shown by *Chu* [2010] (Figures 2, 4, and 5).

[12] There have been previous studies that reported weaker WJs [*Murtugudde et al.*, 2000; *Grodsky et al.*, 2001; *Vinayachandran et al.*, 2002; *Jensen*, 2007; *Nagura and McPhaden*, 2010b]. These studies suggested that weaker trades during IOD events caused an anomalously weak fall WJ. However, to our knowledge, a prolonged departure from the mean climatological state of WJs, which we present here, has not previously been described in the literature. This unique event is marked by its interannual persistence as well as the large magnitude of the zonal wind anomalies. As described in the seminal work of Saji and others in 1999 [*Anderson*, 1999; *Saji et al.*, 1999; *Webster et al.*, 1999], IOD signatures are observed to be phase locked with seasons. Typically SST anomalies appear in June, intensify

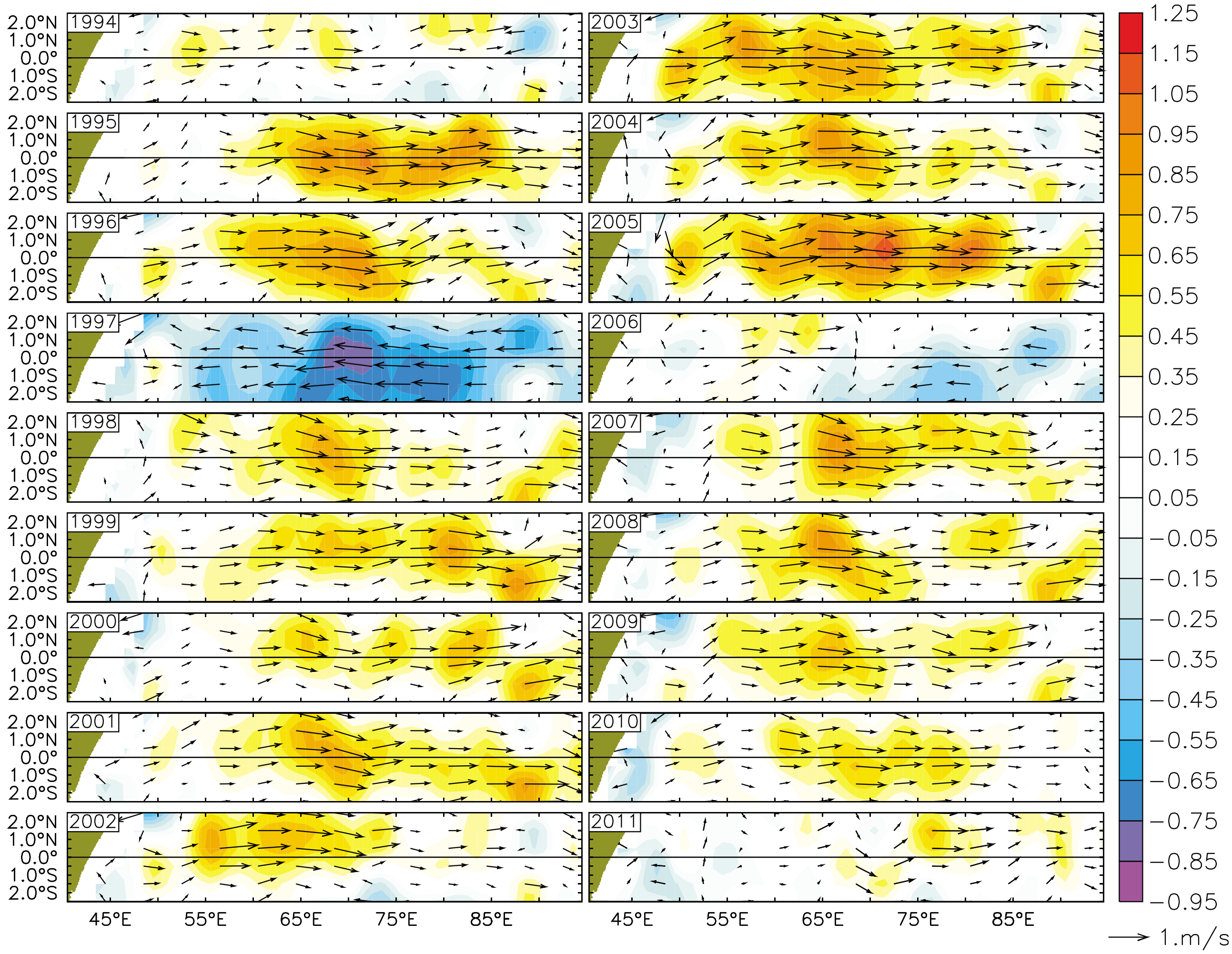

Figure 4. Fall Wyrtki jets from OSCAR currents for the period 1994–2011. For each year an average for currents during October and November was computed. For 2011, data were only available until 16 October. The color shows the magnitude of the zonal components. Units are m s^{-1}.

in subsequent months and reach maximum in October. The typical characteristics of IOD events involve (1) warmer western Indian Ocean with enhanced atmospheric convection, (2) anomalously cool SST anomalies in the eastern Indian Ocean near Sumatra, and (3) abnormal easterly winds along the equator and southeasterly winds near the coast of Sumatra, which lift up the thermocline leading to upwelling of colder waters and as a result suppressed atmospheric convection.

[13] The available history of WJs from 1994 to 2011 (Figures 2, 4, and 5) shows a reasonable association of weak fall WJs to the events of positive IODs with a concurrent El Niño (1994, 1997, 2006) in agreement with previous work [e.g., *Yamagata et al.*, 1996; *Vinayachandran et al.*, 1999; *Murtugudde et al.*, 2000; *Chu*, 2010; *Nagura and McPhaden*, 2010b]. However, weakening of spring jets is apparently not closely associated with El Niño or IOD, as is evident from the relatively intense spring WJ during the strongest El Niño–IOD event of the century in 1997. Further, the 2006 IOD event is reported to be the third strongest IOD in the last 30 years [*Horii et al.*, 2008] and co-occurred with a weak El Niño. It is noteworthy, that Cai and collaborators [*Cai et al.*, 2009] describe the years 2006, 2007 and 2008 as three consecutive years of positive IOD events, and an early excitation of easterlies is common to all 3 years. However, according to their study, each of these events differ in details. In climatological average years, westerlies dominate the equatorial Indian Ocean during spring time and positive zonal winds along the equator generate eastward WJs.

[14] For 2006, the fall jet is much weaker than normal and also weaker than the fairly weak 2007 and 2008 fall jets (Figures 2 and 5). This is in agreement with the literature referred above. In contrast, the weakening of the spring jet in 2006 is less pronounced compared to the weakening of the spring WJs in 2007 and 2008 (Figures 2 and 5). Another notable exception was during 1999 and 2000 when spring WJs were weak, but neither El Niño nor IOD occurred (Figures 2 and 5). Rather, both 1999 and 2000 were years of moderate and weak La Niña's, respectively [*Anyamba et al.*, 2002; http://ggweather.com/enso/oni.htm]. Thus from the available data it may be noted that weakening episodes in spring jets may be more closely associated with evolving or dissipating phases of La Niña or a preoccurrence or postoccurrence of La Niña with a concurrent IOD event

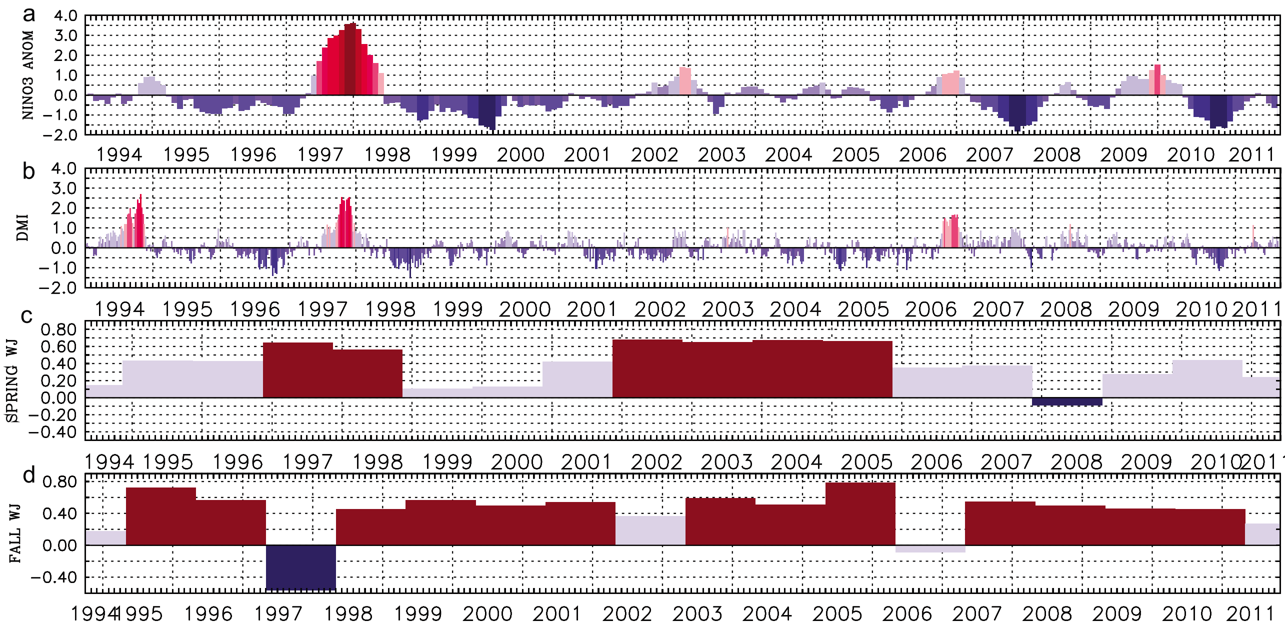

Figure 5. (a) Time series from 1994 to 2011 of Niño3 index, (b) dipole mode index or DMI, (c) magnitude of spring WJ using the May average of zonal currents along the equator and (d) magnitude of fall WJs using the September–October average of zonal currents along the equator. Units for currents are m s^{-1}.

(Figure 5). However, exceptions are found, as this was not the case in 2008 and 2009–2010.

6. Equatorial Winds

[15] A change with latitude of prevailing winds from easterlies in the Southern Hemisphere Indian Ocean to westerlies in Northern Hemisphere exists from spring throughfall. This pattern is influenced by El Niño–Southern Oscillation (ENSO) and IOD as trade winds change during these events [e.g., *Saji et al.*, 1999]. The transition from easterlies to westerlies takes place in a narrow latitude band. This allows us to define the latitude of zero zonal wind, which we will refer to as the latitude of U zero, (LUZ). The farther south of the equator the LUZ occurs, the broader a band of westerlies develops around the equator [*Hastenrath and Polzin*, 2004]. The May monthly mean of NOGAPS equatorial zonal winds, averaged from 60°E to 90°E is presented in Figure 6a for the years 2003 to 2011, and for

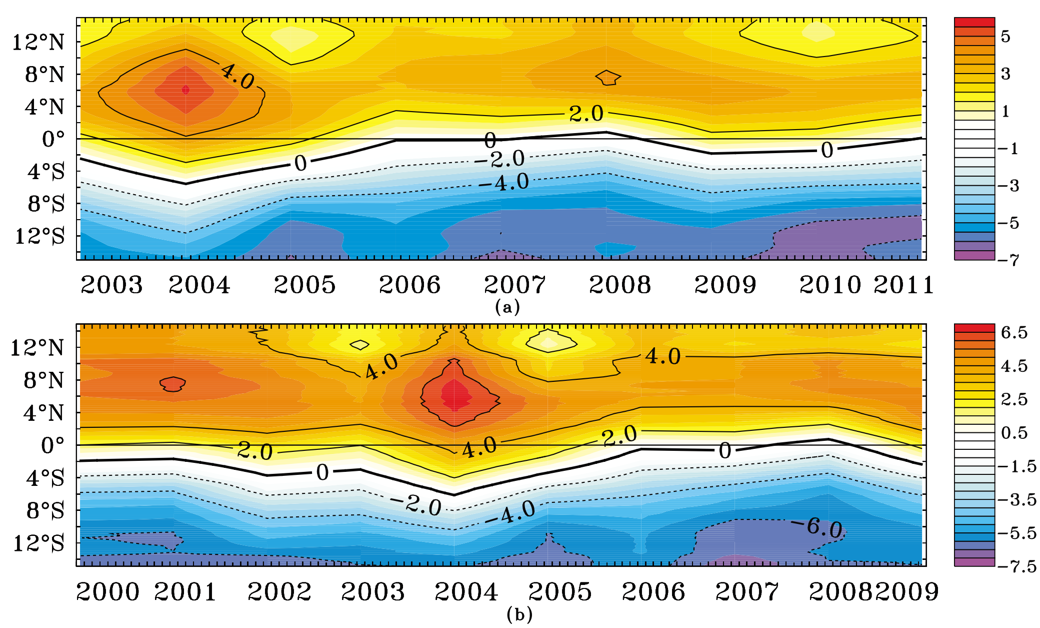

Figure 6. Interannual and meridional variability of May monthly averaged zonal winds (a) from NOGAPS for the years 2003–2011 and (b) for QuikSCAT for the years 2000–2009. The average from 60°E to 90°E in the equatorial Indian Ocean is shown for both data sets. Contour levels are in m s^{-1}.

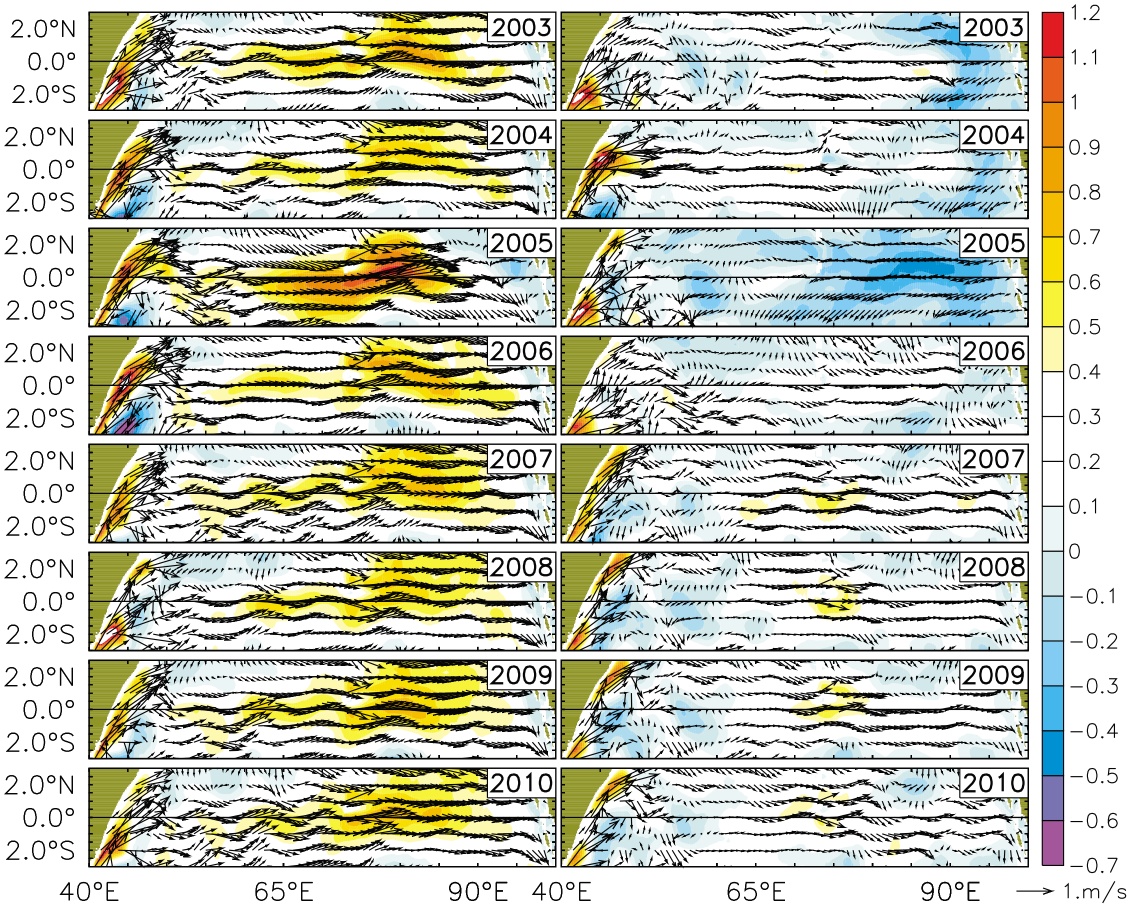

Figure 7. Interannual variability in spring WJs shown as May average zonal currents (left) in HYCOM experiment 2 forced by repeated 2003 winds and (right) in HYCOM experiment 3 forced by winds from 2006 through 2009, followed by 3 years of 2010 winds. Magnitude of the zonal currents are shown in color. Unit is m s^{-1}.

QuikSCAT winds for the years 2000 to 2009 in Figure 6b. It is evident from Figure 6b that for the period 2002–2005, the zonal winds at the equator were westerly during the month of May, and the LUZ was located much farther south of the equator than usual. This was also the case for 2003, although it has been characterized as an aborted dipole year [*Rao et al.*, 2009]. The WJs in both HYCOM and OSCAR appear close to typical WJs during this period (Figure 2). Beginning in 2006, the position of the LUZ shifted toward the equator. In 2008, the LUZ was located to the north of the equator resulting in easterly equatorial winds. This is in the opposite wind direction of normal years, and forced an anomalous reversal of the WJ to westward flow. We note that 2006 had signals of a weak El Niño with a concurrent IOD, while 2007 had a moderate La Niña and a concurrent IOD and was followed by a third IOD, unrelated to ENSO, in 2008 [*Cai et al.*, 2009; http://ggweather.com/enso/oni.htm]. The period 2009–2010 had a strong El Niño followed by a moderate La Niña in 2010–2011 (Figure 5). These events may have imparted strong influence on equatorial winds, which were weak starting from 2006 up to 2011, resulting in weak WJs during those years. In both model and OSCAR zonal currents, WJs appear much weaker for these post-2005 years though there was a small regain of strength in 2010. In both NOGAPS and QuikSCAT products (Figures 6a and 6b) there are a remarkable interannual variability in location of equatorial wind reversals. It is clearly discernible that the strength of the zonal wind along equator was close to zero for successive years starting in 2006. Westerly winds along the equatorial waveguide play an important role in generating eastward propagating Kelvin waves during the month of May in the equatorial Indian Ocean [*Rao et al.*, 2010]. Absence or weakening of westerlies along the equator has a profound influence on the Kelvin wave propagation during those years. The weak winds caused a drastic reduction in the strength of WJs, in particular during 2008, which showed a dramatic reversal of the WJs in concurrence with easterly zonal winds.

7. Numerical Experiments With Modified Wind Forcing

[16] Based on the discussion above, it is our hypothesis that the interannual change in wind forcing is the cause of the weak WJs found in our analysis of OSCAR and HYCOM results. In order to assess the importance of winds on WJs we have designed two additional experiments. The model results in section 6 were based on our control run,

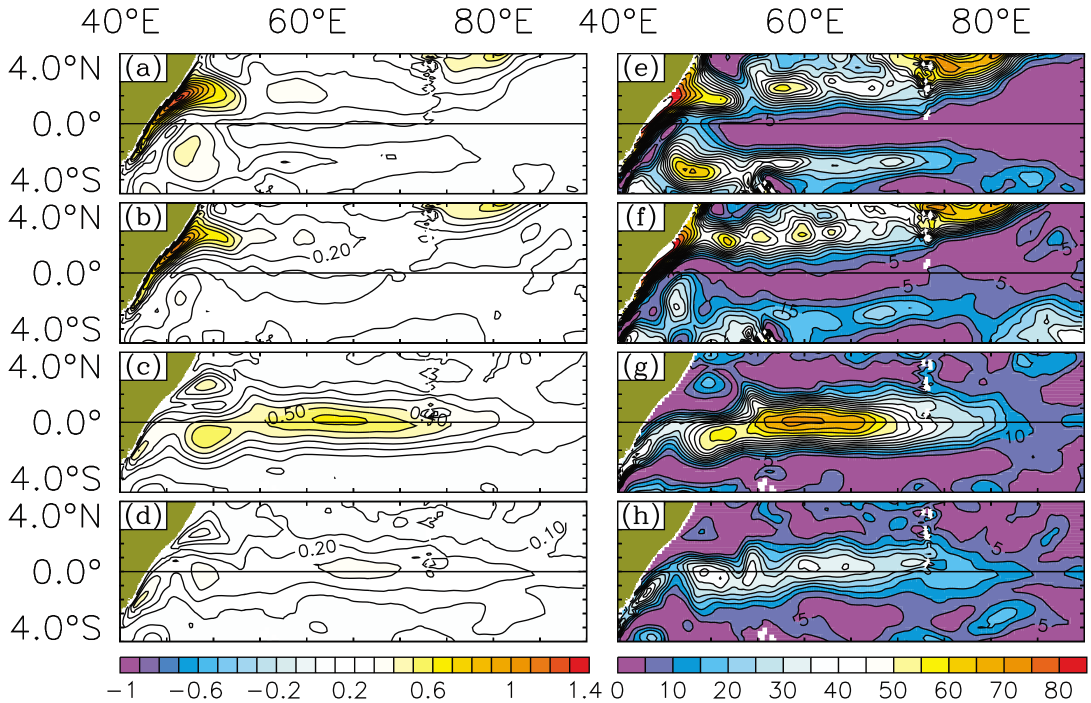

Figure 8. (left) Amplitude in m s^{-1} and (right) percent variance explained of annual (Figures 8a, 8b, 8e, and 8f) and semiannual (Figures 8c, 8d, 8g, 8h) harmonics of zonal currents for (8a, 8c, 8e, 8g) years 2003–2005 and (8b, 8d, 8f, 8h) years 2006–2010.

where HYCOM was forced with realistic winds and buoyancy fluxes from 2003 up to 2011. In the second experiment, we address a hypothetical situation in which 2003 winds persisted for the entire 7 year period. The heat flux and net precipitation varied from year to year as in the control run. We will refer to this experiment as the "strong westerly wind case". In the third experiment, we used the initial condition from 1 January 2003, but used the winds from 2006 to 2009 as forcing during the first 4 years, followed by 3 years repeatedly forced with 2010 winds, while buoyancy forcing was the same as for the control run. We will refer to this as the "weak westerly wind case". The WJs generated from the control experiment are presented in Figure 2 (right). In Figure 7 we present the 2003 repeat experiment or "strong westerly case" (Figure 7, left) and the result from the third experiment, the "weak westerly wind case" (Figure 7, right). In the "strong westerly wind case", the May WJs are well developed for all years, although oceanic initial conditions and buoyancy fluxes are the same as in the control (Figure 2). The WJs are strong and eastward like normal years. In the third experiment, "weak westerly wind case", weak spring WJs develop each year for 5 years in the model ocean. This demonstrates that wind forcing is the cause of weak WJs in the Indian Ocean in the control case. It also infers that the wind pattern starting in 2006 has changed compared to earlier years, which may be associated with the decadal frequency modulations of IOD and ENSO [*Tozuka et al.*, 2007; *Wang et al.*, 2009] or imply a change of state of the coupled atmosphere-ocean system on an even longer timescale.

[17] The atmospheric forcing causing weak WJs can be examined in light of the finding of *Hastenrath and Polzin* [2004], which suggests that the latitude of recurvature of winds in equatorial Indian Ocean is a function of the relative competition between an opposing pressure gradient force and Coriolis acceleration. When a steep eastward pressure gradient and slow trade wind are present, the LUZ can be located far south of equator resulting in a wider than usual latitude range of westerlies in the equatorial region. Without a strong eastward pressure gradient term, the recurvature latitude and therefore LUZ will be located close to the equator [*Hastenrath and Polzin*, 2004]. A recent study on the temporal asymmetry of El Niño and La Niña by *Okumura et al.* [2011] finds that during La Niña the anomalous easterly winds, forced by surface precipitation anomalies, exist in the Pacific and are much stronger compared to the westerly anomalies induced by the Indian Ocean cooling. They suggest this is the reason for the prolonged duration of La Niña compared to El Niño. The stronger easterlies from the Pacific may act as a deterrent for development of strong westerlies over the Indian Ocean during La Niña years, resulting in the LUZ being close to equator and consequently weak spring WJs.

8. Impact of Weak WJs on Annual and Semiannual Harmonics

[18] WJs are known to have a stronger semiannual harmonic component than annual harmonic due to basin resonance in response to semiannual winds [*Jensen*, 1993; *Han et al.*, 1999; *Hastenrath and Polzin*, 2004]. To assess the impact of weakening of WJs on these two major components of seasonal variability, we performed two independent harmonic analyses of the zonal currents from HYCOM for the periods 2003–2005 (3 years) and for 2006–2010 (5 years). The amplitude and variance of both harmonics are presented in Figure 8.

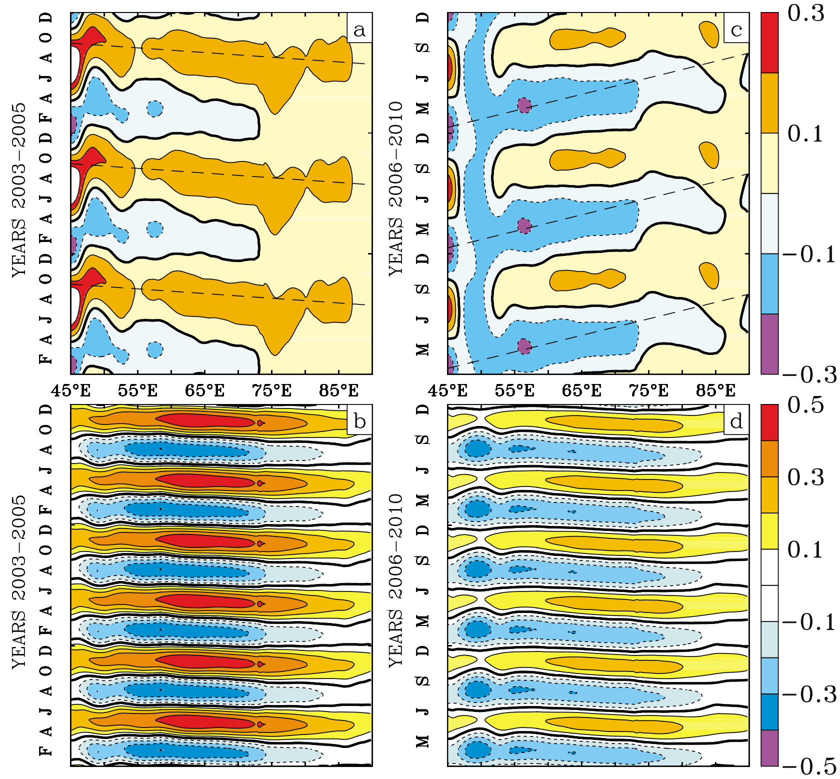

Figure 9. Hovmöller plot of reconstructed zonal currents at the equator using (top) annual and (bottom) semiannual harmonics for the years (a and b) 2003–2005 and (c and d) 2006–2010. Units are m s^{-1}.

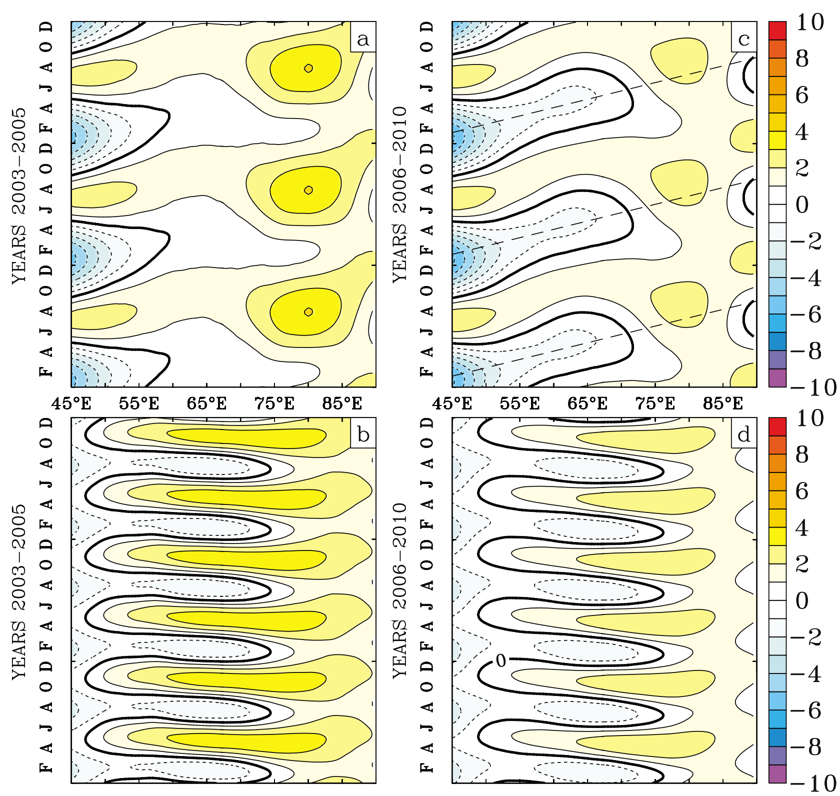

Figure 10. Hovmöller plot of reconstructed zonal winds along the equator using (top) annual and (bottom) semiannual harmonics for period (a and b) 2003–2005 and (c and d) 2006–2010. Units are m s^{-1}.

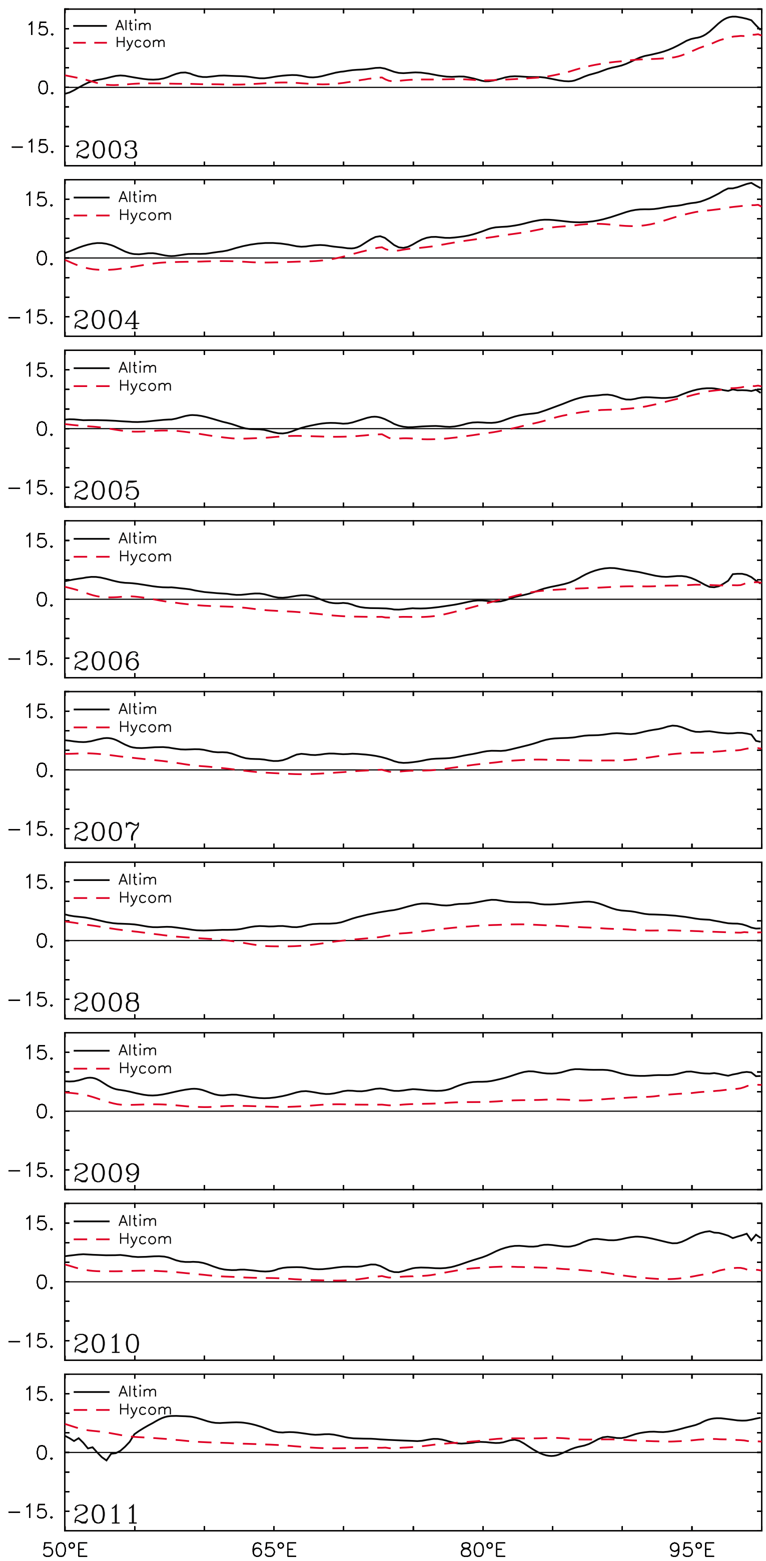

Figure 11. Comparison of HYCOM sea level anomalies (red) with those observed by altimeter (black) along the equator during the period 2003–2011. Units are cm.

[19] It is clearly discernible from Figures 8a, 8b, 8e, and 8f that the weak WJs after 2005 have affected both annual and semiannual components, though the impact is larger on the semiannual harmonic (Figures 8c, 8d, 8g, and 8h). Figure 8a shows that variability of the Somali Current provides the main contribution to the annual component. After 2005, the Somali Current was only at about half of its strength in comparison with earlier years when it had a maximum core

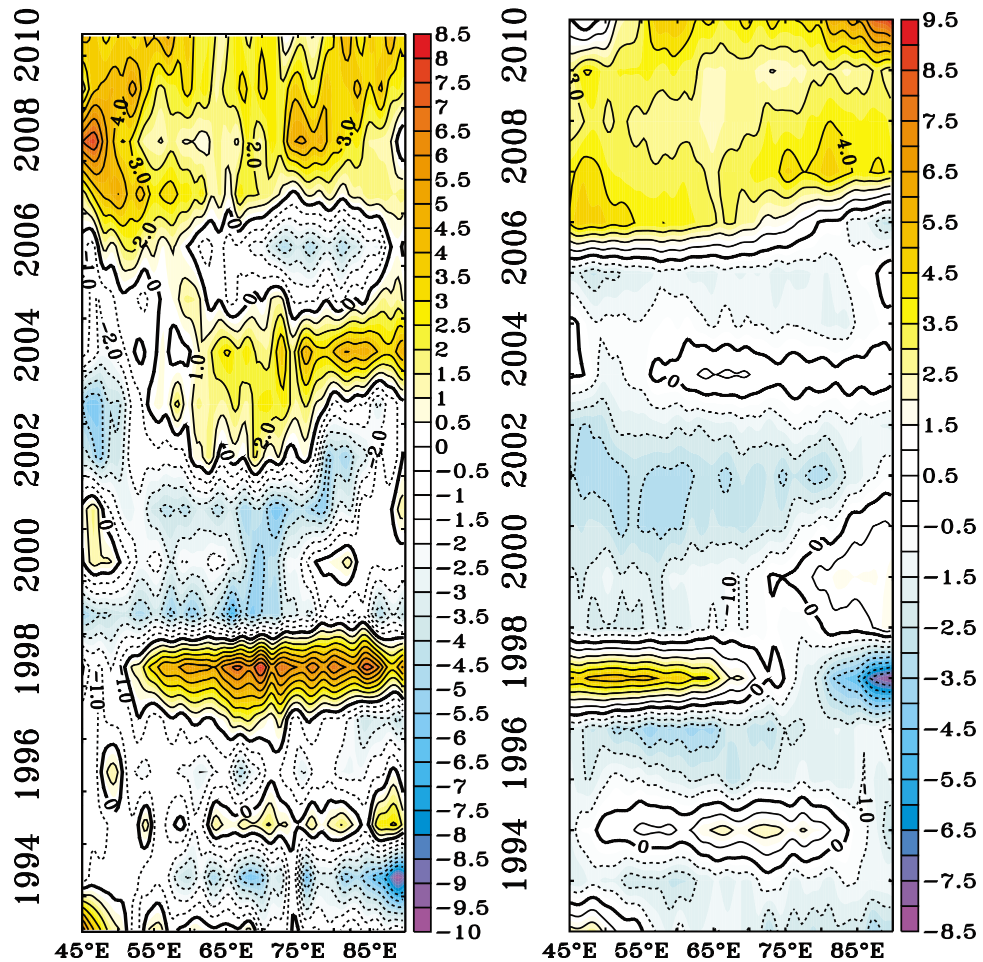

Figure 12. Hovmöller diagram of altimeter sea level anomalies along the equator from 1993 to 2010. (left) Data for the May monthly mean for each year averaged over the latitude band from 1°S to 1°N. (right) Annual anomalies using the same spatial averaging. Units are cm.

speed of 1.4 m s^{-1} (Figure 8b). Before 2006 the annual signal originating from the Somali Current can be traced as a continuous amplitude band along the 0.2 m s^{-1} contour as far east as 85°E (Figure 8a), whereas in later years (Figure 8b), the contours are discontinuous due to a much weaker contribution from the Somali Current. The amplitude of the semiannual harmonic is more strongly affected by the weakening of the WJs after 2006 as is evident from Figure 8d, which is just about a third of the amplitude found in earlier years (Figure 8c). The variance explained by the semiannual harmonic in 2006 and in later years is just 30% at the core of the WJ, which amounts to reduction of more than half compared to earlier years, where the semiannual harmonic explained more than 70% of the total variance (Figures 8g and 8h).

[20] The computed annual and semiannual components were used to reconstruct the zonal currents and winds for the two periods discussed above and are presented in Figures 9 and 10. The slope of the isotach of the annual component of zonal currents in Figure 9a shows that for the period before 2006, there is evidence for westward propagation with a phase speed about 0.9 m s^{-1}, which is attributable to westward propagating Rossby waves. Annual zonal wind harmonics do not show any propagation (Figure 10a) and hence the westward propagation in zonal currents is purely oceanic. However, starting in 2006, the signal has a negative sign and the slope of the isotachs is of opposite direction for zonal currents (Figure 9c), which shows wave propagation toward the east. The phase speed estimated from the axis of the slope of the isotachs is close to 0.2 m s^{-1} which is far below the expected Kelvin wave speed at 2.8 m s^{-1}. However, the reconstructed zonal wind shows a propagating long tongue which has the same slope as the current isotachs (Figure 10c). Thus, Figures 9c and 10c suggest that the slow propagation seen in zonal currents starting in 2006 is associated with the winds. In case of the semiannual component, propagation is not evident from the reconstructed zonal currents. However, there is a clear decrease in amplitude starting in 2006, which correlates well with the decrease in the semiannual harmonic of zonal wind speed. Before 2006 the semiannual amplitude reached a maximum close to 0.5 m s^{-1} at longitudes from 60°E to 70°E, but was just 0.1 m s^{-1} in later years.

9. Response of Equatorial Sea level

[21] In order to investigate the sea level response associated with the WJ currents, we have compared the sea level

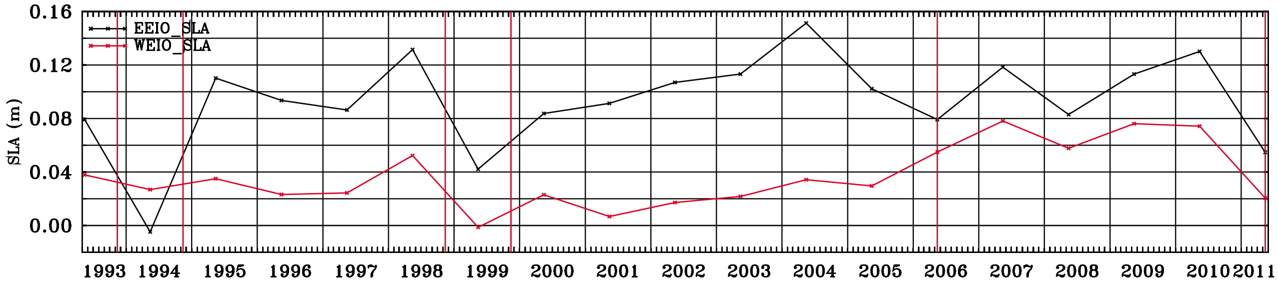

Figure 13. Variability of sea level in the eastern Indian Ocean (50°E to 55°E, 2°S to 2°N average) and in the western Indian Ocean (90°E to 95°E, 2°S to 2°N average) from 1993 to 2011. The rectangular boxes show periods of reduced gradients. Units are m.

anomaly from altimeter with model sea level anomalies after removing the mean over the period of analysis. We present the response of the sea level anomaly from both altimeter and HYCOM in Figure 11. As described in the literature [*Rao et al.*, 1989; *Schott and McCreary*, 2001], WJs carry warm upper layer waters eastward, lowering sea level and decreasing mixed layer thickness in the west while increasing those quantities in the east. During the period 2003–2005, there is clear evidence for water piling up along the eastern boundary both in altimeter and HYCOM anomalies, which is a direct consequence of strong WJs during those years. However, from 2006 the sea level anomaly along the eastern boundary is much lower compared to the years 2003–2005, which is consistent with weaker WJs after 2006. This appears to be due to the fact that mean sea level remained high in western Indian Ocean during 2006–2011, a deviation from the lower sea level usually seen in the west during normal years.

[22] In an attempt to search for similar events in earlier years, mean sea level anomalies for the month of May for each year are computed starting from 1993. They are used to compute an equatorial sea level anomaly by averaging across the equator from 1°S to 1°N. It was used to construct the Hovmöller diagram presented in Figure 12. Figure 12 (left) shows contours of the May sea level anomalies, and Figure 12 (right) presents the annual mean computed the same way. Figure 12 clearly demonstrates that a regime shift appears to take place in 2006. The later years are unique compared to positive anomaly events in the past, during 1994–1995, 1997–1998 and 2003–2004, by having continuously positive anomalies along the equator and from year to year. In 2006 and later the sea level is uniformly high from east to west along the equator and ranged 3–4 cm above the May mean as well as above the annual mean. In earlier years the sea level anomaly was negative, in particularly in the western equatorial Indian Ocean. Exceptions are the dipole year of 1997 and the aborted IOD event of 2003. For the anomalous period beginning in 2006, the east-west sea level gradient became negligible due to the high sea level anomalies present in the west. This span of positive sea level anomalies in the equatorial Indian Ocean is the longest lasting event seen in the available history from altimeter data. Figure 13 shows the May sea levels in two areas: The east equatorial area (EEIO) is from 50°E to 55°E, 2°S to 2°N and the western equatorial area (WEIO) is from 90°E to 95°E with the same latitude extent as the EEIO. Figure 13 summarizes the conclusions in this section, that the west to east gradients have significantly decreased since 2006. Our analysis of sea level anomaly data provides additional support for the weakening of WJs starting in 2006.

10. Summary and Conclusions

[23] Based on OSCAR currents and model results supported by current meter mooring data and sea level data, there is clear evidence for a weakening of spring Wyrtki jets during the period from 2006 to 2011. The weakening is apparently associated with La Niña events or IOD years preceding La Niña years, when the location of the latitude of zero zonal winds (LUZ) is close to the equator. The observed weakening of the WJs partitions its loss of energy to both its annual and semiannual harmonics with maximum effect on the semiannual component, which also may be influenced by a slight reduction in intensity of fall jets. Before 2006, dominant annual westward wave propagation was found in reconstructed model currents using an annual harmonic with a phase speed of 0.9 m s^{-1}, which is equivalent to that of baroclinic mode 1 equatorial Rossby waves. However, after 2006, the expected Rossby wave propagation was not observed in the reconstructed zonal currents. Instead, an eastward propagation with a phase speed of about 0.2 m s^{-1} with a correlated propagation of the NOGAPS zonal wind was found. Thus the eastward waves appear to be wind forced waves generated in the years since 2006. The existence of this eastward propagation has to our knowledge not been previously reported in the literature. Starting in 2006, it is evident that equatorial zonal winds were close to zero due to the proximity of the LUZ to the equator and resulted in weaker WJs than normal. Two model experiments using alternative wind forcing show that the wind is the trigger for this response. The 2003 wind repeat experiment or "strong westerly wind case" demonstrates that the state of the ocean would return to near-normal conditions with the generation of westerlies at least as strong as the 2003 winds. It is evident from our analysis that the sea level of the equatorial Indian Ocean responded to the weak WJs by the presence of low east-west gradients during the 2006–2011 period as compared to its reconstructed history from 1993 using altimeter data. These results warrant further efforts to assess the impact of weakened jets on the Indian monsoon and other coupled ocean-atmosphere processes.

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References

Anderson, D. (1999), Extremes in the Indian Ocean, *Nature*, *401*(6751), 337–339.

Antonov, J., S. Levitus, T. P. Boyer, M. Conkright, T. O' Brien, C. Stephens, and B. Trotsenko (1998), *World Ocean Atlas 1998*, vol. 3, *Temperature of the Indian Ocean*, *NOAA Atlas NESDIS*, vol. 29, 166 pp., NOAA, Silver Spring, Md.

Anyamba, A., C. J. Tucker, and R. Mahoney (2002), From El Niño to La Niña: Vegetation response patterns over east and southern Africa during the 1997–2000 period, *J. Clim.*, *15*(21), 3096–3103.

Barron, C. N., and L. F. Smedstad (2002), Global river inflow within the Navy Coastal Ocean Model, in *Oceans 02 MTS/IEEE*, pp. 1472–1479, Inst. of Electr. and Electron. Eng., New York.

Bleck, R. (2002), An oceanic general circulation model framed in hybrid isopycnic-Cartesian coordinates, *Ocean Modell.*, *4*(1), 55–88.

Bonjean, F., and G. S. E. Lagerloef (2002), Diagnostic model and analysis of the surface currents in the tropical Pacific Ocean, *J. Phys. Oceanogr.*, *32*(10), 2938–2954.

Cai, W., A. Pan, D. Roemmich, T. Cowan, and X. Guo (2009), Argo profiles a rare occurrence of three consecutive positive Indian Ocean dipole events, 2006–2008, *Geophys. Res. Lett.*, *36*, L08701, doi:10.1029/2008GL037038.

Cane, M., and D. Moore (1981), A note on low-frequency equatorial basin modes, *J. Phys. Oceanogr.*, *11*, 1578–1584.

Chu, P. C. (2010), Observational studies on association between eastward equatorial jet and Indian Ocean dipole, *J. Oceanogr.*, *66*(3), 429–434.

Grodsky, S. A., J. A. Carton, and R. Murtugudde (2001), Anomalous surface currents in the tropical Indian Ocean, *Geophys. Res. Lett.*, *28*, 4207–4210.

Han, W., 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*(9), 2191–2209.

Hastenrath, S., and D. Polzin (2004), Dynamics of the surface wind field over the equatorial Indian Ocean, *Q. J. R. Meteorol. Soc.*, *130*(597), 503–517.

Hastenrath, S., A. Nicklis, and L. Greischar (1993), Atmospheric-hydrospheric mechanisms of climate anomalies in the western equatorial Indian Ocean, *J. Geophys. Res.*, *98*(C11), 20,219–20,235, doi:10.1029/93JC02330.

Horii, T., H. Hase, I. Ueki, and Y. Masumoto (2008), Oceanic precondition and evolution of the 2006 Indian Ocean dipole, *Geophys. Res. Lett.*, *35*, L03607, doi:10.1029/2007GL032464.

Jensen, T. G. (1993), Equatorial variability and resonance in a wind-driven Indian Ocean model, *J. Geophys. Res.*, *98*, 22,533–22,552.

Jensen, T. G. (2007), Wind-driven response of the northern Indian Ocean to climate extremes, *J. Clim.*, *20*(13), 2978–2993.

Knox, R. A. (1976), On a long series of measurements of Indian Ocean equatorial currents near addu atoll., *Deep Sea Res Oceanogr Abstr.*, *23*(3), 211–221.

Large, W. G., J. C. McWilliams, and S. C. Doney (1994), Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization, *Rev. Geophys.*, *32*(4), 363–403.

McCreary, J. P., P. Kundu, and R. L. Molinari (1993), A numerical investigation of dynamics, thermodynamics and mixed layer processes in the Indian Ocean, *Prog. Oceanogr.*, *31*, 181–244.

McPhaden, M. J. (1982), Variability in the central Indian Ocean. part I: Ocean dynamics, *J. Mar. Res.*, *40*, 157–176.

McPhaden, M. J., G. Meyers, K. Ando, Y. Masumoto, V. S. N. Murty, M. Ravichandran, F. Syamsudin, J. Vialard, L. Yu, and W. Yu (2009), RAMA: The Research Moored Array for African-Asian-Australian monsoon analysis and prediction, *Bull. Am. Meteorol. Soc.*, *90*(4), 459–480.

Metzger, E. J., O. M. Smedstad, P. Thoppil, H. E. Hurlburt, A. J. Wallcraft, D. S. Franklin, J. F. Shriver, and L. F. Smedstad (2008), Validation test report for the Global Ocean Prediction System v3.0–1/12° HYCOM/NCODA: Phase I, *NRL Memo. Rep. NRL/MR/7320-08-9148*, Nav. Res. Lab., Washington, D. C.

Metzger, E. J., H. E. Hurlburt, A. J. Wallcraft, J. F. Shriver, T. L. Townsend, O. M. Smedstad, P. G. Thoppil, D. S. Franklin, and G. Peggion (2010), Validation test report for the Global Ocean Forecast System v3.0–1/12° HYCOM/NCODA: Phase II, *NRL Memo. Rep. NRL/MR/7320–10-9236.*, Nav. Res. Lab., Washington, D. C.

Murtugudde, R., J. P. McCreary Jr., and A. J. Busalacchi (2000), Oceanic processes associated with anomalous events in the Indian Ocean with relevance to 1997–1998, *J. Geophys. Res.*, *105*(C2), 3295–3306, doi:10.1029/1999JC900294.

Nagura, M., and M. J. McPhaden (2010a), Wyrtki jet dynamics: Seasonal variability, *J. Geophys. Res.*, *115*, C07009, doi:10.1029/2009JC005922.

Nagura, M., and M. J. McPhaden (2010b), Dynamics of zonal current variations associated with the Indian Ocean dipole, *J. Geophys. Res.*, *115*, C11026, doi:10.1029/2010JC006423.

O'Brien, J. J., and H. E. Hurlburt (1974), Equatorial jet in the Indian Ocean: Theory, *Science*, *184*(4141), 1075–1077.

Okumura, Y. M., M. Ohba, C. Deser, and H. Ueda (2011), A proposed mechanism for the asymmetric duration of El Niño and La Niña, *J. Clim.*, *24*(15), 3822–3829, doi:10.1175/2011JCLI3999.1.

Qiu, Y., L. Li, and W. Yu (2009), Behavior of the Wyrtki Jet observed with surface drifting buoys and satellite altimeter, *Geophys. Res. Lett.*, *36*, L18607, doi:10.1029/2009GL039120.

Rao, R. R., R. L. Molinari, and J. F. Festa (1989), Evolution of the near-surface thermal structure of tropical Indian Ocean: 1. Description of mean monthly mixed-layer depth, and sea surface temperature, surface current, and surface meteorological fields., *J. Geophys. Res.*, *94*, 10,801–10,815.

Rao, R. R., M. S. Girish Kumar, M. Ravichandran, A. R. Rao, V. V. Gopalakrishna, and P. Thadathil (2010), Interannual variability of kelvin wave propagation in the wave guides of the equatorial Indian Ocean, the coastal Bay of Bengal and the southeastern Arabian Sea during 1993–2006, *Deep Sea Res., Part I*, *57*(1), 1–13, doi:10.1016/j.dsr.2009.10.008.

Rao, S. A., V. V. Gopalakrishna, S. R. Shetye, and T. Yamagata (2002), Why were cool SST anomalies absent in the Bay of Bengal during the 1997 Indian Ocean dipole event?, *Geophys. Res. Lett.*, *29*(11), 1555, doi:10.1029/2001GL014645.

Rao, S. A., J.-J. Luo, S. Behera, and T. Yamagata (2009), Generation and termination of Indian Ocean dipole events in 2003, 2006 and 2007, *Clim. Dyn.*, *33*, 751–767.

Reppin, J., F. 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*(7), 15,495–15,514.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata (1999), A dipole mode in the tropical Indian Ocean, *Nature*, *401*(6751), 360–363.

Schott, F., and J. P. McCreary (2001), The monsoon circulation of the Indian Ocean, *Prog. Oceanogr.*, *51*, 1–123.

Subrahmanyam, B., V. Murty, and D. M. Heffner (2011), Sea surface salinity variability in the tropical Indian Ocean, *Remote Sensing of Environment*, *115*(3), 944–956, doi:10.1016/j.rse.2010.12.004.

Thompson, B., C. Gnanaseelan, and P. Salvekar (2006), Variability in the Indian Ocean circulation and salinity and its impact on SST anomalies during dipole events, *J. Mar. Res.*, *64*(6), 853–880, doi:10.1357/002224006779698350.

Tozuka, T., J. J. Luo, S. Masson, and T. Yamagata (2007), Decadal modulations of the Indian Ocean dipole in the SINTEX-f1 coupled GCM, *J. Clim.*, *20*(13), 2881–2894, doi:10.1175/JCLI4168.1.

Vinayachandran, P. N., N. H. Saji, and T. Yamagata (1999), Response of the equatorial Indian Ocean to an unusual wind event during 1994, *Geophys. Res. Lett.*, *26*(11), 1613–1616, doi:10.1029/1999GL900179.

Vinayachandran, P. N., S. Iizuka, and T. Yamagata (2002), Indian Ocean dipole mode events in an ocean general circulation model, *Deep Sea Res., Part II*, *49*(7–8), 1573–1596.

Wang, X., D. Wang, and W. Zhou (2009), Decadal variability of twentieth-century El Niño and La Niña occurrence from observations and IPCC AR4 coupled models, *Geophys. Res. Lett.*, *36*, L11701, doi:10.1029/2009GL037929.

Webster, P., A. Moore, J. Loschnigg, and R. Leben (1999), Coupled ocean-atmosphere dynamics in the Indian Ocean during 1997–98, *Nature*, *401*(6751), 356–360.

Wyrtki, K. (1973), An equatorial jet in the Indian Ocean, *Science*, *181*, 262–264.

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*(C5), 12,465–12,473.

T. G. Jensen and A. J. Wallcraft, Oceanography Division, Naval Research Laboratory, Code 7320, Bldg. 1009, Stennis Space Center, MS 39529, USA.

S. Joseph, M. Ravichandran, and S. S. C. Shenoi, Indian National Centre for Ocean Information Services, Ministry of Earth Sciences, Hyderabad, Andhra Pradesh 500090, India. (sjo@incois.gov.in)

S. Nayak, Ministry of Earth Sciences, Prithvi Bhavan, Lodhi Road, New Delhi, Delhi 110 003, India.