Abrupt termination of Indian Ocean dipole events in response to intraseasonal disturbances

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Received 25 June 2004; accepted 20 September 2004; published 14 October 2004.

[1] We have investigated the role of atmospheric intraseasonal disturbances on termination of the Indian Ocean Dipole (IOD) using multiple datasets. We observed significant 30–60 day atmospheric disturbances associated with equatorial westerly zonal winds prior to termination of all IOD events except for the 1982 and 1997 events. The westerlies excite anomalous downwelling Kelvin waves that terminate the basin-wide coupled processes by warming the eastern Indian Ocean. The 1982 and 1997 IOD events coincide with the strongest El Niño events; during those years the intraseasonal disturbances are weak. *INDEX TERMS:* 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4231 Oceanography: General: Equatorial oceanography; 4572 Oceanography: Physical: Upper ocean processes. **Citation:** Rao, S. A., and T. Yamagata (2004), Abrupt termination of Indian Ocean dipole events in response to intraseasonal disturbances, *Geophys. Res. Lett.*, *31*, L19306, doi:10.1029/2004GL020842.

1. Introduction

[2] Indian Ocean Dipole (IOD), a coupled ocean-atmosphere phenomenon, in the tropical Indian Ocean [*Saji et al.*, 1999; *Rao et al.*, 2002a; *Yamagata et al.*, 2002] has vast impact on the climate of areas around the Indian Ocean [*Ashok et al.*, 2001; *Zubair et al.*, 2003; *Behera and Yamagata*, 2003; *Saji and Yamagata*, 2003, and references therein]. An IOD event is associated with anomalously cool (warm) SST in the southeastern (central and western) equatorial Indian Ocean. Normally an IOD event evolves in spring (May/June), peaks in fall (October) and terminates in early winter (December) [*Saji et al.*, 1999]. Just like the El Niño events all IOD events do not follow this evolution exactly. In particular, the termination phase of IOD events (see Figure 1).

[3] The role of intraseasonal disturbances (ISD), popularly known as the Madden-Julian Oscillations (MJO), has been discussed widely as a possible trigger of El Niño/Southern Oscillations (ENSO) events [e.g., *Luther et al.*, 1983; *Kessler et al.*, 1995]. Termination of El Niño events is also attributed to the intraseasonal disturbances [*Takayabu et al.*, 1999].

[4] In the central tropical Pacific, strong westerly wind bursts lasting from 1 to 3 weeks are found prior to every El Niño event [*Luther et al.*, 1983]. These westerly wind bursts are thought to be responsible for eastward advection of warm water to initiate El Niño event through exciting warm equatorial Kelvin jets [*Kessler et al.*, 1995; *McPhaden and Yu*, 1999]. *Takayabu et al.* [1999] suggested that an exceptionally strong MJO in May 1998 with an abrupt intensification of the easterly trade winds accelerated the termination of 1997–98 El Niño event. Many studies linked the activity of ISD in the tropical Indian Ocean to the active and break monsoon conditions over the Indian subcontinent [e.g., *Madden and Julian*, 1994; *Sperber et al.*, 2000]. Since the ISD propagate along the equator in the tropical Indian Ocean, it is of interest to discuss their relation with the IOD.

[5] In this article using multiple datasets, we examine a possible link particularly between occurrence of ISD and IOD termination. It is important to know this link because if the coupled phenomenon in the tropical Indian Ocean terminates before its usual evolution cycle, then the coupled phenomenon will no longer be able to impact the climate in the surrounding areas.

2. Data

[6] The following datasets were used in the analysis: (1) monthly SST from Global sea-Ice and Sea Surface Temperature data set (GISST) [*Rayner et al.*, 1996] (1958–1982), (2) optimum interpolated weekly SST (1982–1997) [*Reynolds et al.*, 2002], (3) 3-day rainfall over oceans and SST from Tropical Rain Measuring Mission (TRMM) Microwave Imager (TMI) (1998–2003), (4) merged sea surface height data from aviso, France (1993–2003). (5) surface zonal winds derived from National Center for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) reanalysis dataset (1950–2003) [*Kalnay et al.*, 1996] (6) surface zonal winds derived from European Center for Medium Range Weather Forecast reanalysis (ERA) dataset (1958–2001), (7) satellite derived winds from Quick Scat (2000–2003) and (8) Global Precipitation Climatology Project (GPCP) version 2 combined precipitation available at ftp://daac.gsfc.nasa.gov/. Anomalies of each dataset are obtained by removing the mean seasonal cycle for respective periods of the datasets. To study the intraseasonal disturbances, we have applied the Lanczos band pass filter with 100 weights and half-power frequency cutoffs of 20 and 90 days^{-1} to the rainfall, OLR and wind datasets.

3. Results

[7] Figure 1 shows the evolution of SST anomalies during six IOD events. Among them, 5 cases correspond to strong IOD events in 1961, 1967, 1982, 1994 and 1997 with strength of the Dipole Mode Index (measure of

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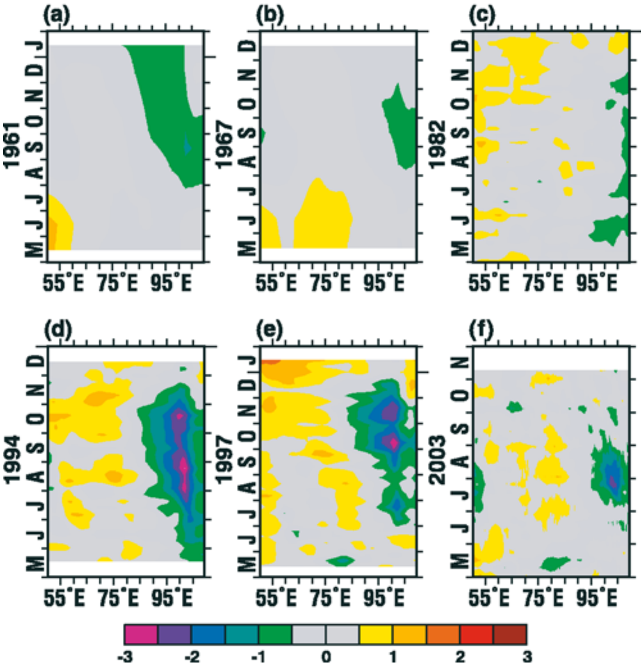

Figure 1. SST anomalies (deg. C) averaged between 10°S-Eq in the tropical Indian Ocean for (a) 1961, (b) 1967, (c) 1982, (d) 1994, (e) 1997 and (f) 2003 IOD events. Monthly GISST data is used for (a) and (b), weekly OI SST is used for (c), (d) and (e) and 3-day SST from TMI onboard TRMM is used for (f).

strength of the IOD defined as difference in SST anomalies between southeastern equatorial Indian Ocean and Western equatorial Indian Ocean) greater than one standard deviation [*Saji et al.*, 1999], and one case corresponds to a recent aborted event in 2003. Although the composite IOD event basically decays in boreal winter, exact date of termination of IOD events changes from event to event. Termination of an IOD event is determined when the strong gradient in SSTA in the zonal direction ceases. The 1961 IOD event terminated in January 1962; the 1967 event lasted until October; the 1982 event terminated in November; the 1994 event demised in November; the 1997 event continued until December. The IOD event in 2003 was aborted by the end of August. *Vinayachandran et al.* [1999], *Saji et al.* [1999], *Webster et al.* [1999], *Murtugudde et al.* [2000], and *Rao et al.* [2002a, 2002b] demonstrated that anomalous easterly winds along the equator excite anomalous upwelling Kelvin waves that lift the thermocline in the eastern Indian ocean, thereby initiating air-sea interaction in the tropical Indian Ocean. Therefore, using the NCEP/NCAR reanalysis data-set we first study possible roles of the equatorial zonal winds.

[8] Figure 2 shows the wavelet spectrum of zonal winds, derived from NCEP/NCAR reanalysis, averaged over a region between 70–90°E and 5°S–5°N for the years of 5 strong and 1 recent aborted IOD events as mentioned above. The general feature in all panels of Figure 2 is absence of significant ISD activity during the duration of active IOD. However, just prior to the termination of IOD, significant ISD activity in 30–60 days band is observed for 1961, 1967, 1994 and 2003 events (shown as ellipses in Figure 2). But for the 1982 & 1997 events absolutely no significant ISD activity is observed prior to the termination. The wavelet spectrums of zonal winds derived from ECMWF, also show similar features (figure not shown). Striking similarities between the analyses of two datasets enhance the confidence of results presented here. The wavelet analysis of OLR anomalies averaged over a region between 10°S-Eq. and 90°–110°E (to represent the convection in the Indian Ocean warm pool) for the IOD events of 1982, 1994, 1997 and 2003 years shows that the structure of the wavelet spectrum of OLR anomalies in the eastern Indian Ocean (not shown) is similar to the wavelet spectrum of the equatorial zonal winds (Figure 2). This similarity confirms that the zonal winds in the equatorial region are directly linked with the convective activities. It may be noted here that OLR data is continuously available since 1979.

[9] Figures 3a and 3b show the rainfall and surface wind anomalies averaged over a region between 10°S–10°N. We observe eastward propagating rainfall anomalies during November in 1994 and August in 2003. Usually, rainfall systems associated with the ISD are observed over the warm water pool region from the Indian Ocean to the central Pacific, but diminish over the eastern Pacific [*Wang and Rui*, 1990; *Takayabu et al.*, 1999]; only dry dynamical waves propagate around the globe [*Takayabu et al.*, 1999]. In Figure 3 we confirm this and see continuous propagation of rainfall anomalies from the Indian Ocean region to the central Pacific. When there is a heat source at the equator, we observe easterly wind anomalies to the east of the heat source and westerlies to the west of the heat

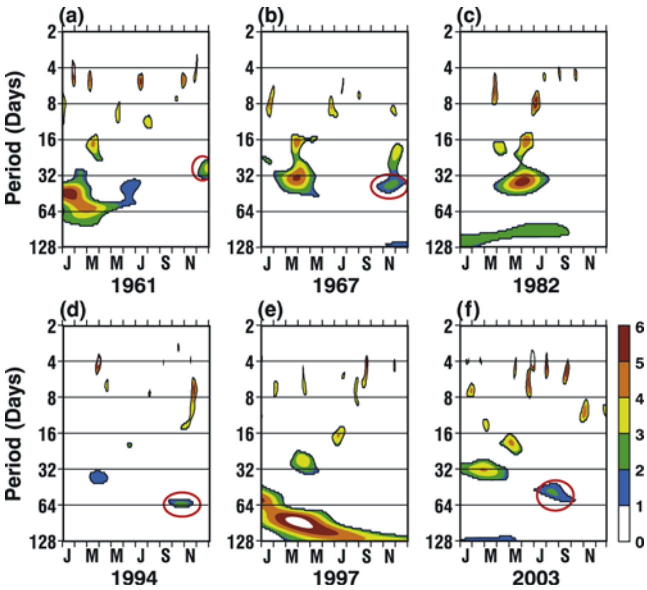

Figure 2. Wavelet spectrum of equatorial daily zonal wind anomalies for 1961, 1967, 1982, 1994, 1997 and 2003 IOD events. Daily zonal winds from NCEP/NCAR reanalysis are used. The wavelet spectrum is normalized with the global wavelet spectrum and normalized spectrum is plotted only if the spectrum is significant at 95% confidence limit. Ellipses (Thick line) show the regions of significant 30–60 day oscillations in zonal wind prior to the termination of IOD events.

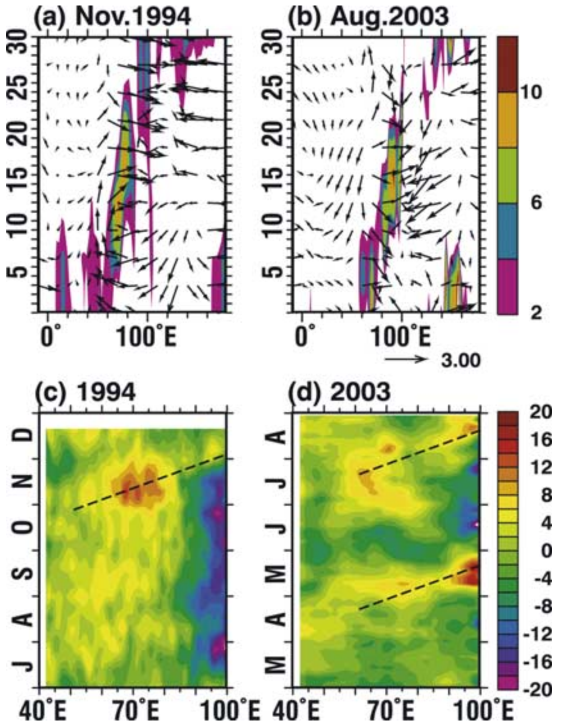

Figure 3. 20–90 day filtered anomalies of rain (shading, mm/day) and winds (vectors, m/sec) averaged between 10°S–10°N for (a) 1994 and (b) 2003 IOD events. SSH anomalies (cm) for (c) 1994 and (d) 2003 IOD events from merged SSH dataset. GPCP pentad rainfall is used for (a) and TMI rainfall is used for (b) Path of propagation of downwelling Kelvin-wave is shown with dashed line.

source [*Matsuno*, 1966; *Gill*, 1980]. This is due to the Kelvin wave response to the east and Rossby wave response to the west of the heat source. The intensification of twin cyclones, associated with the Rossby wave response, results in strong westerly wind bursts [*Takayabu et al.*, 1999]. We observe these wind bursts in the equatorial Indian Ocean during the above mentioned 30–60 day oscillation in zonal winds (Figures 3a and 3b). However, easterlies to the east of the rainfall anomalies are weak when the convection reaches the central Indian Ocean.

[10] These strong westerly wind anomalies drive anomalously strong downwelling Kelvin waves, which deepen the thermocline in the eastern Indian Ocean. These oceanic Kelvin waves are clearly seen in Figures 3c and 3d for both the IOD events in 1994 and 2003. The deepened thermocline in the eastern Indian Ocean therefore influences the SST in the eastern Indian Ocean by reducing the role of cold water entrainment and hence the coupled phenomenon in the eastern Indian Ocean ceases. On the other hand, strong easterlies (Figures 3a and 3b) in the equatorial Indian Ocean prior to the movement of convection to the central Indian Ocean maintain the coupled phenomenon in the tropical Indian Ocean (Figures 3c and 3d). Since sea surface height data derived from satellites is available only after 1993, we can not show the evidence for presence of downwelling Kelvin waves for 1961 and 1967 IOD events. However, from our understanding of equatorial ocean dynamics, we can expect similar ocean response to the observed ISD activity in 1961 and 1967 events. In addition to the equatorial ocean dynamics, air-sea fluxes associated with ISD may also play an important role in terminating the IOD event. Weak easterlies to the east of convection result in decreased latent heat flux, and clear sky conditions in the east further enhance the solar radiation (not shown). Combined effect of these fluxes results in warm SSTs in the eastern Indian Ocean.

4. Summary and Discussion

[11] We have shown occurrence of strong 30–60 day intraseasonal disturbances (ISD) of zonal winds prior to the termination of all major and one recent aborted Indian Ocean Dipole (IOD) events (Figure 2). Two IOD years with no action of this significant ISD signal prior to the termination are 1982 and 1997. This may be due to the co-occurrence of the strongest El Niño events in 1982 and 1997; *Slingo et al.* [1999] showed that the MJO activity was weaker during these two El Niño events. Interestingly, all other IOD events considered in this study occur in absence of strong El Niño events. These IOD events are normally referred as pure IOD events [*Rao et al.*, 2002a]. Therefore, we conclude that termination of pure IOD events is primarily due to the 30–60 day ISD activity in equatorial zonal winds in the tropical Indian Ocean. This 30–60 day ISD of zonal winds is part of the global Madden-Julian oscillation, which is associated with eastward propagation of convection accompanied by strong westerlies (Figures 3a and 3b) due to intensified twin cyclones of Matsuno-Gill type circulation.

[12] We also analyzed other major IOD events since 1958. In agreement with the above conclusion, we observe significant ISD activity prior to the termination of 1977 pure IOD event (figure not shown). During other two IOD events (1963 and 1972), we do not observe significant ISD activity prior to the termination; this is once again in agreement with the above conclusion as these two events co-occurred with El Niño events.

[13] The strong anomalous westerlies associated with the 30–60 day ISD of equatorial zonal winds excite anomalous downwelling Kelvin waves (Figures 3c and 3d) that reach the eastern Indian Ocean within a month. Owing to the downwelling Kelvin waves, the thermocline in the eastern equatorial Indian Ocean deepens. The deepened thermocline warms the overlying SST by reducing the role of cold water entrainment (major source for cool SSTA during an IOD event [*Murtugudde et al.*, 2000; *Yu and Rienecker*, 2000]); therefore the air-sea coupled phenomenon ceases abruptly. The role of air-sea fluxes on termination of the IOD is also expected. Normally, an ISD event is associated with clear sky conditions and weak easterlies to the east of strong convection associated with an ISD event. The combined effect of weak winds and clear sky further enhances the warming of the eastern Indian Ocean.

[14] On the other hand, termination of IOD events that co-occur with El Niño events is due to the reduction of latent heat flux (in response to reduced wind speeds; probably due to onset of northeast monsoon) and also due to increased insolation [*Murtugudde et al.*, 2000; *Yu and Rienecker*, 2000]. During the termination phase of these IOD events, thermocline remains shallower than normal, but with reduced upwelling [*Murtugudde et al.*, 2000].

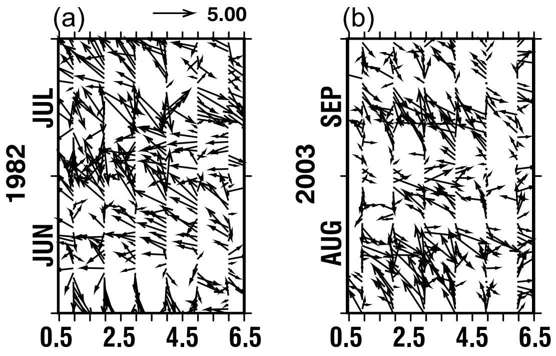

Figure 4. Daily wind vectors (m/sec) averaged in different boxes along the coast of Sumatra and Java for (a) 1982 and (b) 2003 IOD events. Units along the x-axis are box numbers 1. (95–100°E, 2°S-Eq.), 2. (95–100°E, 2°–4°S), 3. (97–102°E, 4°–6°S), 4. (99–104°E, 6°–8°S), 5. (104–108°E, 8°–10°S), and 6. (108–112°E, 8°–10°S).

[15] Not only termination, but also onset of an IOD event is different from event to event for all the major IOD events described above. A clear example is the 1994 IOD event. This event started in May, earlier than other IOD events. One notable feature in Figure 2 is that absence of strong ISD activity in zonal winds prior to the onset of 1994 IOD event. This may explain the earlier onset of 1994 event. Another interesting feature in Figure 2 is presence of strong ISD activity in early June of 1982. This ISD did not terminate the IOD event in 1982; nevertheless it weakened the eastern Indian Ocean cooling for some time in July. This IOD event revived and lasted until November, probably because local winds along the coast of the eastern Indian Ocean favor upwelling (Figure 4a). These winds might have preserved the eastern Indian Ocean cooling. However, ISD activity in August, 2003 abruptly terminated the 2003 IOD event. The local winds along the coast of eastern Indian Ocean during this time are favorable for downwelling (Figure 4b). Therefore, the timing of an ISD event and the local conditions in the eastern Indian Ocean, together, are important factors in determining the termination of an IOD event. Further, most of the IOD events (except 1967 and 2003 events) considered in this study terminate in the northeast monsoon season. During this time the local winds along the coast of eastern Indian Ocean favor downwelling [*Saji et al.*, 1999; R. Suzuki et al., Seasonal and interannual variabilities of the Indian Ocean climate, submitted to *Journal of Physical Oceanography*, 2004]. Therefore it is appropriate to envisage the role of northeast monsoon in termination of IOD events. These issues need to be addressed further in detail.

[16] **Acknowledgments.** We thank W. S. Kessler for his kind guidance on using the Lanczos filter. Useful comments from H. Nakamura, Y. Masumoto, K. Ashok and S. K. Behera are acknowledged. We are grateful to the two anonymous reviewers for their constructive comments. FERRET freeware is used for the present work.

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