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Triggering of the positive Indian Ocean dipole events by severe cyclones over the Bay of Bengal

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ABSTRACT

In this paper, we suggest that positive Indian Ocean dipole events could be triggered by the occurrence of severe cyclones over the Bay of Bengal during April–May. All positive Indian Ocean dipole events during 1958–2003 are preceded by at least one such severe cyclone. Severe cyclones over the Bay of Bengal strengthen the meridional pressure gradient across the eastern equatorial Indian Ocean (EEIO) and hence lead to the intensification of the upwelling favourable southeasterlies along the Sumatra coast. Severe cyclones can also lead to a decrease in the integrated water vapour content and suppress convection over the EEIO. We suggest that the suppression of convection over the EEIO in turn, leads to the enhancement of convection over the western equatorial Indian Ocean (WEIO) and hence to the weakening of the westerlies along the central equatorial Indian Ocean (CEIO). This can lead to a positive feedback between suppression of convection over the EEIO and enhancement of the convergence and convection over the WEIO. If this positive feedback continues until the winds over the CEIO become easterlies, the convection over the EEIO remains suppressed for a period much longer than the synoptic scale. The strong upwelling caused by the easterlies along the equator and southeasterlies along the Sumatra coast decreases the sea surface temperature of the EEIO very rapidly and a positive dipole event gets triggered.

1. Introduction

Positive events of the Indian Ocean dipole mode appear to have a large impact on the climate of the region surrounding the Indian Ocean. The positive Indian Ocean Dipole (IOD) event of 1997 was associated with severe floods in eastern Africa and severe drought in Indonesia, and attracted great public attention (Birkett et al., 1999; Saji et al., 1999; Webster et al., 1999; Li et al., 2003). The excess summer monsoon rainfall over India during 1994 has been attributed to the positive IOD event (Behera et al., 1999; Gadgil et al., 2003, 2004) and it has been suggested that the Indian summer monsoon rainfall was close to the average in 1997, in spite of the occurrence of a very strong El Nino, because of the occurrence of a positive IOD event (Ashok et al., 2001; Gadgil et al., 2003, 2004). Clearly, it is important to understand the factors and processes that trigger these positive events and the mechanisms involved in their evolution. We address this problem in this paper. We show that all the positive IOD events during 1958–2003 are preceded by at least one such severe cyclone over the Bay of Bengal. We suggest that positive dipole events could be triggered by the occurrence of severe cyclones over the Bay of Bengal during April–May.

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The evolution of positive IOD events involves the suppression (enhancement) of convection over the eastern (western) equatorial Indian Ocean and decrease (increase) in the sea surface temperature (SST) of eastern (western) equatorial Indian Ocean (Fig. 1). While the variation of the SST of the western equatorial Indian Ocean (WEIO, 50°E–70°E, 10°S–10°N) arises primarily from variation in surface fluxes, ocean dynamics and coupling with the atmosphere play an important role in determining the variation of the SST of the eastern equatorial Indian Ocean (EEIO, 90°E–110°E, 10°S–EQ) (Murtugudde et al., 2000; Vinayachandran et al., 2002). Among the factors that determine the evolution of the SST of EEIO, the upwelling along the Sumatra coast, and the horizontal advection are known to be of particular importance in growth of the dipole events (Murtugudde et al., 2000; Vinayachandran et al., 2002).

The IOD is phase locked to the seasonal cycle in the Indian Ocean. IOD events generally start developing in the boreal spring–early summer which has been recognized as the time window in which the ocean atmosphere system is particularly sensitive to external forcing (Saji et al., 1999; Annamalai et al., 2003) and the peak strength of the anomalies generally occur during the boreal fall (Saji et al., 1999; Annamalai et al., 2003; Li et al., 2003). Positive events are characterized by a reversal of

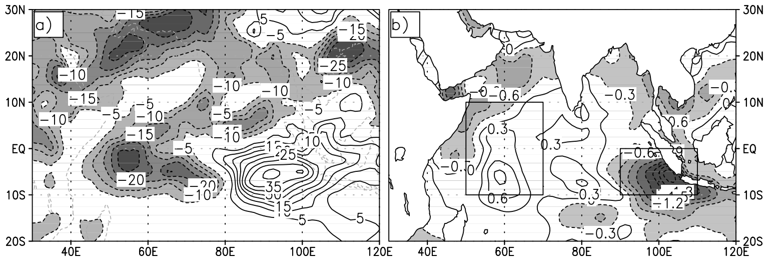

Fig. 1. Anomaly patterns of (a) NOAA AVHRR OLR (Wm^{-2}) and (b) NOAA optimal interpolated SST (°C) for July 1994. EEIO (90°E–110°E, 10°S–EQ) and WEIO (50°E–70°E, 10°S–10°N) are shown with boxes. Data are obtained http://www.cdc.noaa.gov/

zonal gradients in various atmospheric and oceanic parameters such as OLR, SLP, surface wind, SST, etc. (Vinayachandran et al., 1999; Murtugudde et al., 2000; Vinayachandran et al., 2002; Annamalai et al., 2003, etc.). Saji et al. (1999) defined an index for IOD events, the dipole mode index (DMI), which is the difference in SST anomalies between the WEIO and the EEIO. When the average September–November values of DMI are of magnitude one standard deviation or larger, positive IOD events are said to occur (Yamagata et al., 2004). During 1958–1999, positive dipole events occurred in 1961, 1963, 1967, 1972, 1977, 1982, 1994 and 1997 (Yamagata et al., 2004).

The role of the Ocean dynamics in the evolution of IOD events has been extensively studied (Han et al., 1999; Murtugudde et al., 2000; Vinayachandran et al., 2002, etc.). Vinayachandran et al. (1999) have shown that the easterly wind anomalies over the equatorial Indian Ocean excite eastward propagating equatorial Kelvin waves which reflect from the eastern boundary as long Rossby waves that propagate westward. Murtugudde et al. (2000) have shown that the Kelvin waves produced by the equatorial easterly wind stress anomalies raised the thermocline off the Sumatran coast during the development of the IOD event in 1997. The enhanced cooling in the east is caused by coastal upwelling that took place over a mixed layer that was shallower than usual, causing the SST anomalies to be maximum south of the equator, off Sumatra (Vinayachandran et al., 1999). Strengthened alongshore winds enhanced surface cooling by driving stronger upwelling there, which persisted till the end of 1997 (Murtugudde et al., 2000). This is further supported by heat budget estimates from the model temperature equation Vinayachandran et al. (1999), Murtugudde et al. (2000). It has been shown that downwelling Rossby waves in the preceding seasons of IOD event anomalously deepened thermocline in the WEIO Vinayachandran et al. (1999).

Positive IOD events are relatively rare (with eight events during the 42 yr period), and it s believed that they are triggered by some event during spring–early summer (Ueda and Matsumoto, 2000; Kajikawa et al., 2001; Annamalai et al., 2003). Annamalai et al. (2003) studied the roles of ENSO and Indonesian Through Flow (ITF) in the initiation of positive dipole events. They showed that a natural mode of coupled variability of the EEIO intensifies in boreal spring–early summer when ENSO like conditions exist over the west Pacific. They suggest that warm SST anomaly in the equatorial western-central Pacific Ocean leads to a change in the Pacific Walker circulation and induces subsidence and hence reduction of precipitation over the EEIO. They further suggest that, forced by this heat sink, an anticyclone develops in the lower atmosphere over the southeastern Indian Ocean which strengthens the alongshore wind off-Sumatra, leading to a decrease in SST by both upwelling and evaporation. The hypothesis they propose for triggering of the positive dipoles involves these processes along with the reduction of ITF. Ueda and Matsumoto (2000) studied the triggering of 1997 IOD event and suggested that the El Nino of 1997 triggered the IOD event. Another hypothesis for triggering of positive IOD events has been put forth by Kajikawa et al. (2001). They suggest that the intensification of the Hadley cell in the western Pacific between the South China Sea/Philippines Sea (SCS/PS) in the boreal summer, plays an important role in the onset of the IOD events. The positive convection anomalies over SCS/PS and negative convection anomalies over the EEIO and the maritime continent (which, they believe are likely to be forced by ENSO) appear in July and induce low-level cross-equatorial flow, with southeasterly flow along the Sumatra coast. The mechanism they suggest involves enhancement of upwelling in the EEIO by the anomalous alongshore wind leading to the initiation of the positive IOD event. According to Li et al. (2003) the ENSO-South Asian monsoon interaction and the shift in the Walker cell over the Indian Ocean as well as the enhancement of along-shore winds off Sumatra coast as a Rossby wave response to anomalous heat source over the western-central Pacific contribute to the initiation of IOD events. The simulations of Gualdi et al. (2003) with a coupled model suggest that the enhancement of sea level pressure over the tropical eastern Indian Ocean associated with ENSO is favourable for the development of a dipole.

Mechanisms which involve ENSO in the initiation of IOD events (Kajikawa et al., 2001; Annamalai et al., 2003; Gualdi et al., 2003, etc.) obviously do not operate in cases such as 1961 and 1994, in which the positive dipoles are generated in the absence of an El Nino or in the presence of La Nina as in 1967. Annamalai et al. (2003) have pointed out that the 1994 IOD event is not preceded by the trigger they propose, that is, by a warm SST anomaly in the western-central Pacific. Li et al. (2003) also point out that the factors they proposed as being important were absent in the initiation of 1961 and 1994 IOD events. Gualdi et al. (2003) note that in 1979 even though the SLP anomaly over the tropical eastern Indian Ocean was positive, a positive IOD event did not get triggered. Thus it appears that many of the favourable factors proposed may be neither necessary nor sufficient for the development of the dipole.

Studies with ocean models (e.g. Murtugudde and Busalacchi, 1999) and coupled models (Gualdi et al., 2003; Lau and Nath, 2004) suggest that local factors as well as ENSO could determine the sea surface temperature anomalies over the Indian Ocean and lead to the development of an IOD event. In the simulation of Lau and Nath (2004), an anomaly of sea level pressure situated south of Australia in the austral winter plays a role in the initiation of the IOD events which are not associated with ENSO. They show that the two factors they considered, viz., ENSO and the surface pressure changes south of Australia jointly account only about one third of the total variance of the Indian Ocean pattern index suggesting that other factors must be important in triggering some of the IOD events in the simulation.

In this paper, we propose a different hypothesis for the triggering of positive IOD events. We suggest that positive dipole events could be triggered by the occurrence of severe cyclones over the Bay of Bengal during April–May because of the associated enhancement of the southeasterlies off Sumatra and suppression of the integrated water vapour content and hence convection over EEIO. We suggest that the suppression of convection over EEIO due to the cyclone, in turn, leads to intensification of convection over WEIO and a positive feedback between convection and circulation over the equatorial Indian Ocean which can lead to the development of a positive dipole event. Clearly, the hypothesis we propose has to be tested with further studies using models.

The data analysed for this study are described in Section 2. Major features of the evolution of the strong positive events in the satellite era are discussed in Section 3. In Section 4, we consider the impacts of severe cyclones over the Bay of Bengal during April–May. In this section, we discuss the processes involved over different time scales. In Section 5, we present the case study of 1990 severe cyclone event, which did not trigger a positive IOD event. Role of pre-conditioning of Indian Ocean in the evolution of the positive IOD events is discussed in Section 6. Summary and conclusions are given in Section 7.

2. Data

Daily and monthly outgoing long-wave radiation (OLR, http://www.cdc.noaa.gov/) data derived from NOAA satellites have been used extensively as a proxy for deep convection in the tropics (Liebmann and Smith, 1996). Here we use NOAA OLR data for the period 1979–2003 with spatial resolution of 2.5×2.5 deg. Weekly and monthly SST data by (Reynolds and Smith, 1995) is also used in the study. The spatial resolution of this data set is 1×1 deg. This is readily available at http://www.cdc.noaa.gov/. Various daily average products of NCEP Reanalysis (Kalnay et al., 1996) such as zonal wind at the surface, specific humidity at various levels, sea level pressure, 0. are also used. Category and tracks of the cyclones over the Bay of Bengal are obtained from Naval Pacific Meteorology and Oceanography Center of the Joint Typhoon Warning Center (https://metoc.npmoc.navy.mil/jtwc/best_tracks/) and India Meteorological Department archives.

3. Evolution of IOD events

Of the positive dipole events identified on the basis of DMI, the events of 1961, 72, 94 and 97 were particularly strong, with maximum monthly DMI exceeding twice the standard deviation. Another strong IOD event in the recent period was in 1982. A dipole event was also triggered in 2003. However, it terminated abruptly in August (Rao and Yamagata, 2004). Hence its life span was short compared to other major IOD events, such as 1994 and 1997. We consider in detail the evolution of the IOD events of 1982, 1994, 1997 and 2003 in the satellite era, to gain an insight into the triggering of positive dipole events. The critical role played by the eastern parts of equatorial Indian Ocean in the development of positive IOD events has been pointed out in several studies (Murtugudde et al., 2000; Vinayachandran et al., 2002; Annamalai et al., 2003; Li et al., 2003). We, therefore, first consider the variation of the SST of and OLR over the EEIO during April–September of 2003, 1997, 1994 and 1982 (Fig. 2) to identify the period over which the transition to the dipole phase occurred.

In 1997, the OLR over EEIO was consistently above normal from mid May except for a very few days in July and August; the SST was above normal up to mid June and started decreasing very rapidly thereafter (Fig. 2b). It should be noted that in May, almost the entire equatorial Indian Ocean east of 50°E was above 29°C with positive SST anomalies over parts of the EEIO as well as WEIO (Figs. 3a and b).

The average SST of EEIO started decreasing very rapidly from mid June to mid July (Fig. 2b). By July 1997 the SST anomaly pattern had evolved to be that typical of a positive event (Fig. 3f). This SST anomaly pattern sustained and intensified until November when DMI reached a maximum. The variation of OLR in 2003 was similar to that in 1997 up to mid-August (Fig. 2c), with the OLR being below normal in early May and increasing rapidly around mid-May. The OLR then remained above normal until mid-August (Fig. 2c) except for a day or two in June and July. The SST started decreasing from mid-May but at a rate smaller than that in 1997. In mid-August the OLR dipped and the SST started increasing and the event of 2003 got terminated (Fig. 2c). Both in 1982 and 1994, SST anomaly of the EEIO was already negative in mid-April. However, the OLR anomalies were small or negative until the end of April, became positive on several days in May and consistently so from mid June, except for a few days towards the end of June (Fig. 2a). In 1994, the SST remained below normal throughout, decreasing rapidly from late May, and the SST anomaly pattern evolved to the typical SST anomaly pattern of the positive phase by July (Fig. 1). This SST anomaly pattern intensified until October when DMI

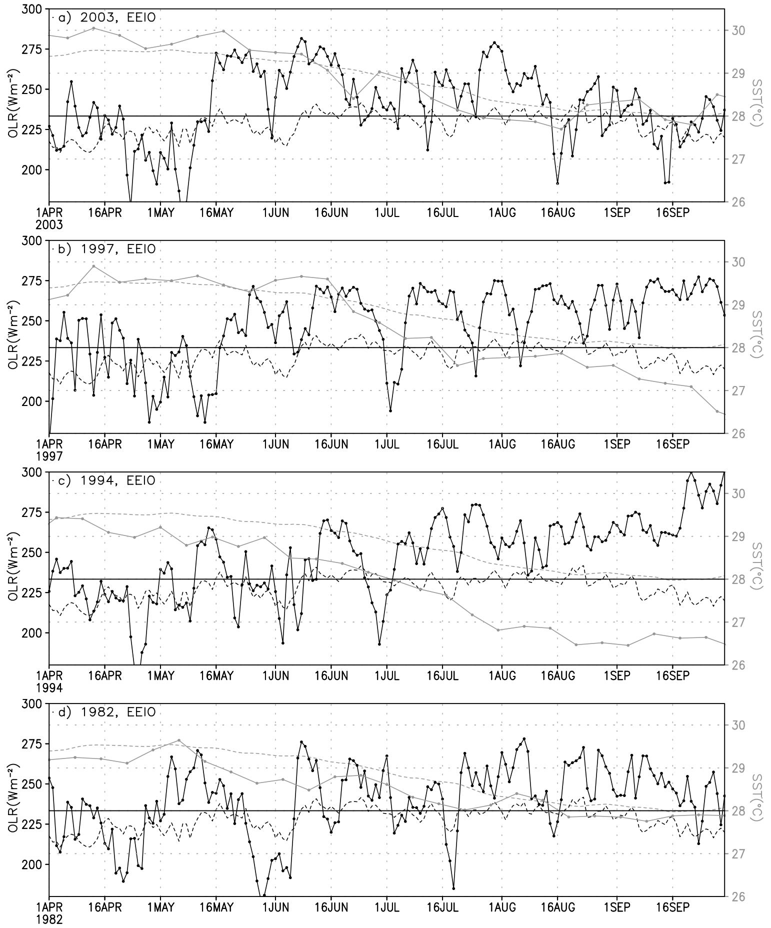

Fig. 2. Variation of weekly mean SST (°C) and daily mean OLR (Wm^{-2}) averaged over EEIO during 1 April–16 September for (a) 2003, (b) 1997, (c) 1994 and (d) 1982. Grey solid (dashed) curve represents the variation of actual SST (SST climatology) and black solid (dashed) curve represents the variation of actual OLR (OLR climatology).

reached its peak. On the other hand, in 1982, the cooling of EEIO was rather gradual.

The variation of the climatological averages of the wind along the Sumatra coast, the zonal component of the surface wind along the central equatorial Indian Ocean (CEIO, 60°E–90°E, 2.5°S–2.5°N) and SST of EEIO and WEIO are shown in Fig. 4a. From April, the mean wind parallel to the coast of Sumatra (region selected for this study is a strip parallel to Sumatra coast shown in Fig. 4b) is northward, leading to upwelling and hence a decrease in SST of the EEIO. During spring, WEIO is warmer than EEIO (Fig. 4a) and the spring jet (Wyrtki, 1973) in the equatorial Indian Ocean driven by westerlies along the equator advects warm water to the eastern Indian Ocean, leading to an increase in the SST of the EEIO and deepening of thermocline (Hastenrath, 2002; Vinayachandran et al., 2002). Whether the SST of the EEIO increases or decreases is determined primarily by the balance between the upwelling and advection. Since the mean wind along the equatorial Indian Ocean starts weakening from the beginning of May (Fig. 4a), the upwelling due to the wind along the coast of Sumatra dominates and SST of the EEIO starts decreasing (Fig. 4a). It reaches a minimum during August–September and then starts increasing again.

Positive IOD events are associated with weakening of westerly wind over the CEIO, until it becomes easterly. The easterly wind anomalies over the equatorial Indian Ocean excite eastward propagating upwelling Kelvin waves along the equator which on reaching the eastern boundary lift the thermocline (Vinayachandran et al., 2002). Thus the easterly wind anomalies over the CEIO enhance the impact of the southeasterlies off Sumatra and lead to more intense upwelling and a rapid decrease in the SST of EEIO. The triggering of positive IOD events, therefore, involves rapid strengthening of the southeasterly wind along the Sumatra coast and weakening of the westerlies (or the presence of easterlies) over the CEIO.

The variation of the along shore wind off Sumatra and the zonal wind over the CEIO during 1982, 1994, 1997 and 2003 is shown in Fig. 5. The development of the events of 1997 and 2003 was rather similar. In both the years, the southeasterlies along the Sumatra coast were very much below the normal in the first half of May, rapidly increased around mid-May to become stronger than normal and stayed above normal almost continuously for several weeks thereafter (Figs. 5d and f). The westerly surface wind over the CEIO strengthened in May and was well above the normal until mid May (Figs. 5c and e). In 1997, it then decreased

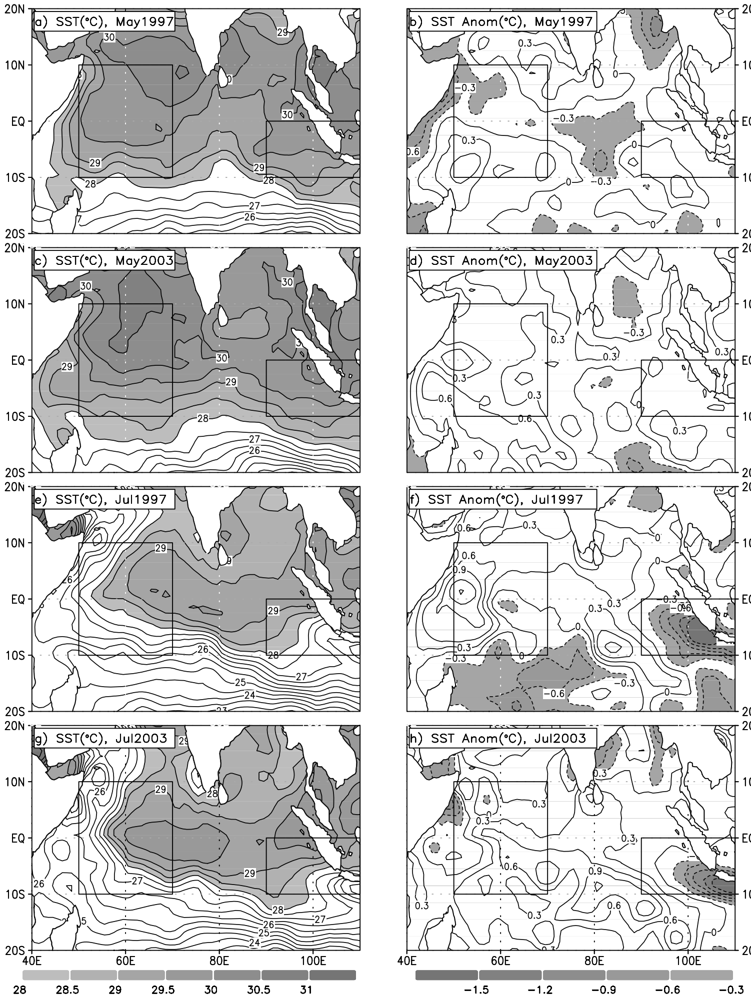

Fig. 3. SST (°C, left) and SST anomaly (°C, right) patterns for May and July of 1997 and 2003.

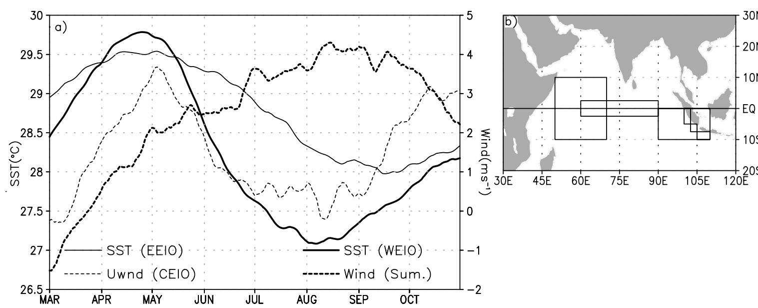

Fig. 4. Variation of the long-term mean of SST (°C) of EEIO and WEIO, surface wind (ms^{-1}) over CEIO (60°E–90°E, 2.5°S–2.5°N) and wind parallel to Sumatra coast (ms^{-1}) during 1 March–31 October. (b) The regions chosen for averaging are shown in the map.

rapidly from mid May up to mid-June when the zonal component became easterly. After mid-June, although there were fluctuations in the amplitude of the winds over these two regions, the southeasterly component remained above normal and the zonal wind over CEIO was easterly almost throughout the season (Figs. 5c and d). In 2003, the westerlies over the CEIO started weakening rapidly from mid-May and became easterlies for a few days from 20 to 26 May then almost continuously from early June. The easterlies persisted until the end of July over the CEIO. The July SST anomaly pattern (Fig. 3h) was typical of the positive dipole event. Thus the triggering of these IOD events occurred in mid-May and by mid-June the wind over CEIO as well as the wind off Sumatra were highly favourable for upwelling in EEIO. However, in 2003 convection revived over the EEIO in early August (Fig. 2c) and the surface wind over CEIO became westerly from the second week of August (Fig. 5e). At the same time, upwelling favourable wind along the Sumatra coast started weakening (Fig. 5f) and the IOD event of 2003 was

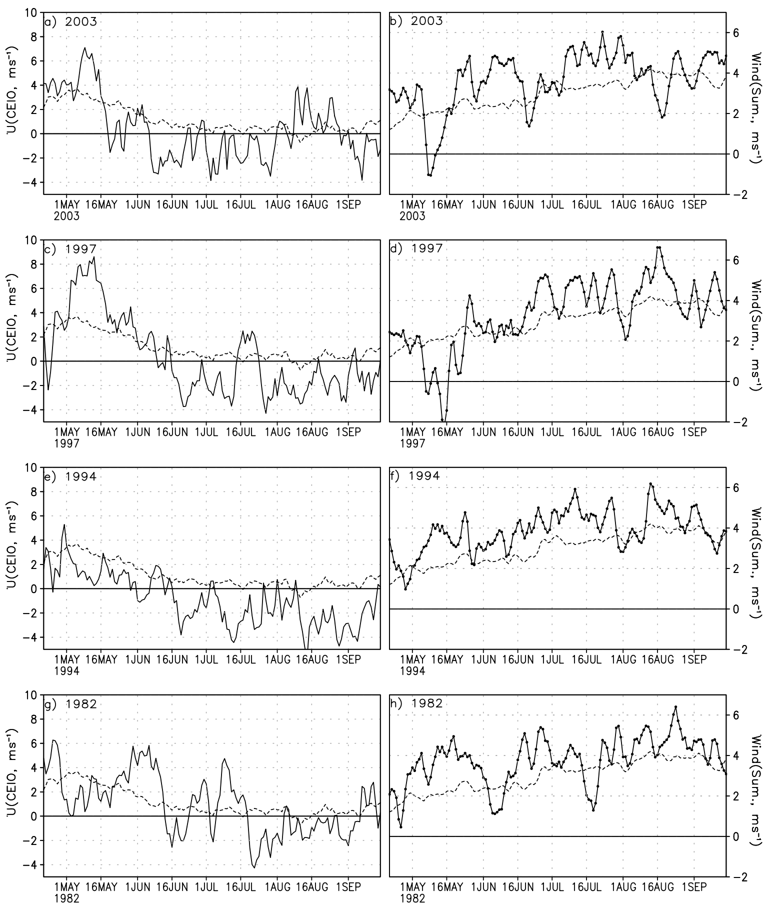

Fig. 5. Variation of the zonal component of daily mean surface wind (ms^{-1}) over CEIO (ms^{-1}) during 16 April–16 September of (a) 2003, (c) 1997, (e) 1994 and (g) 1982. Variation of the wind parallel to Sumatra coast (ms^{-1}) during 16 April–16 September of (b) 2003, (d) 1997, (f) 1994 and (h) 1982. Solid curves in are the actual zonal wind component and the dashed curves are respective climatologies.

terminated by the second week of August (Rao and Yamagata, 2004).

The 1994 event was different from that of 1997 in that the SST and thermocline depth of the EEIO and WEIO were favourable for the development of a positive event from spring (Fig. 2, Vinayachandran et al., 1999, 2002). The SST of the EEIO was already below normal in mid-April. However, it should be noted that the SST was still high enough to sustain convection and OLR was near normal. From May onwards, there were anomalous easterlies over the CEIO (suggesting that the spring jet was unusually weak); the thermocline in the EEIO was shallower than normal (Vinayachandran et al., 1999, 2002). In the first two weeks of May, the southeasterlies along the Sumatra coast became consistently stronger and the strength of the westerly wind over CEIO decreased rapidly (Figs. 5a amd b) and the positive dipole event was triggered. The zonal wind over the CEIO became easterly in the beginning of June (Fig. 5a). In 1982, the surface wind over CEIO remained anomalous easterly from early May for two weeks. However, it again turned as westerly during the last week of May. It became consistently easterly from mid July only. On the other hand, the wind along the Sumatra coast remained consistently strong southeasterly from early May itself. Thus, in 1997, 2003, 1994 as well as 1982, the triggering of the dipole event appears to have occurred in May and the transition to the dipole phase achieved by the end of June/early July.

4. Impact of severe cyclones over the Bay of Bengal during April–May

After the unanticipated failure of the Indian summer monsoon in 2002 (Gadgil et al., 2002), we were monitoring the developments over the Indian Ocean from May 2003. The major event in May 2003 was a severe cyclone over the Bay of Bengal, which was rather similar to the one that occurred in May of the last major positive dipole event in 1997. Subsequently, the southeasterlies off Sumatra strengthened, the westerlies over CEIO weakened and became easterlies in early June and it was announced on the IOD website (http://www.jamstec.go.jp/) that a positive dipole event had been triggered. This suggested that the cyclone over the Bay of Bengal might have played a role. We found that, in fact, each of the positive IOD events (Yamagata et al., 2002) were preceded by the occurrence of at least one severe cyclone over the Bay of Bengal (Table 1, Fig. 6). This led to an investigation of the role of severe cyclones over the Bay of Bengal in triggering of positive dipole events which is presented in this paper.

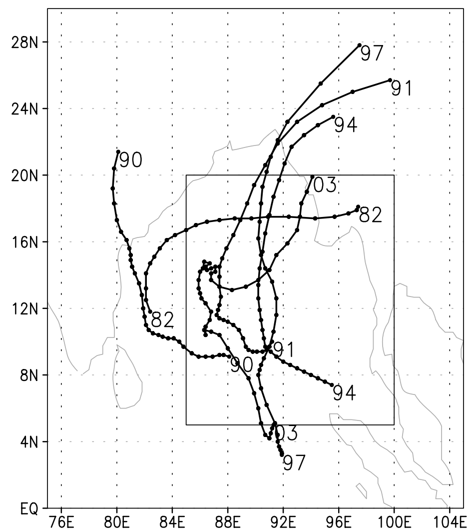

Fig. 6. Tracks of the systems which attained severe cyclone strength over the bay during April–May in the period 1982–2003.

It is interesting to note that the timing of the triggering of the events of 1982, 1994, 1997 and 2003 noted in the previous section coincides with the occurrence of the severe cyclone over the Bay of Bengal. Here we consider facets of the impact of severe cyclones over the Bay, which could be important for the development of positive IOD events. We note that although each of the positive IOD events (Yamagata et al., 2002) were preceded by the occurrence of at least one severe cyclone over the Bay of Bengal, the converse is not true. For example, in 1990, even though there was a severe cyclone over the Bay of Bengal in May, it did not trigger a positive IOD event. We consider this case in Section 5.

We find that severe cyclones over the Bay of Bengal in April–May can lead to atmospheric conditions which are favourable for the development of a dipole event, viz., (i) an intensification of the meridional sea level pressure (SLP) gradient across the EEIO and (ii) suppression of integrated water vapour content (IWVC, integrated from surface to 300 hPa) and hence convection over EEIO. We expect such impacts to last for a period of the order of the lifespan of the cyclone, that is, a few days. We show in Section 4.1 that in some years subsequent convective systems contribute to the intensification of the suppression of the convection over EEIO and prolonging the period to about two weeks. We suggest in Section 4.2, processes that can lead to continued suppression of convection over EEIO for longer periods of the order of a month and hence the triggering of a positive IOD event.

4.1. Processes involved over time scales of about 1–2 weeks

4.1.1. Impact on the surface pressure gradient. Severe cyclones over the Bay of Bengal are associated with a sharp decrease in the surface pressure which is generally reflected in the strengthening of the north-south pressure gradient across the EEIO. The surface wind parallel to the Sumatra coast tends to follow the north-south surface pressure gradient across the EEIO closely and intensifies with the intensification of the cyclone. This is clearly seen in the daily variation of the component of surface wind parallel to the west coast of Sumatra and difference in average surface pressure between the Indian Ocean south of EEIO (90°E–110°E, 20°S–10°S) and Bay of Bengal (henceforth the bay, 80°E–100°E, 5°N–20°N) for the severe cyclones in 1997 and 2003 (Fig. 7). The days on which the cyclone was active over the bay are marked on the SLP gradient curve with different symbols representing the strength of the cyclone. The gradient in SLP across the EEIO built up with the intensification of depression over the bay and its intensification to a severe cyclone (e.g. 13–19 May 1997) and the southeasterlies along the Sumatra coast also strengthened. Thereafter, the pressure gradient continued to be strong for a long period of time and hence the southeasterlies remained stronger than normal, except on a few occasions.

It is seen from Fig. 7 that there is a decrease in the meridional pressure gradient as well as the southeasterlies along the Sumatra coast in the initial stages of the depressions/cyclones in both the cases. This decrease is associated with an intensification of a convective system in the southern hemisphere. We find that in almost all the years (with the exception of 1990), the systems that intensified to severe cyclones over the Bay of Bengal during April–May were generated as organized convection over the WEIO. The evolution of the convection over the WEIO for all the cases (with the exception of 1990) was, in fact, similar to the generation of Madden-Julian Oscillation (MJO, Madden and Julian, 1971, 1972) as described by Wang and Rui (1990) involving eastward propagation and subsequent splitting into two systems-one on either side of the equator, followed by propagation of the northern system over the Bay of Bengal. Then the northern one intensified and a rapid increase in the pressure gradient and the southeasterly wind off the Sumatra coast was observed Fig. 7. These observations suggest that the genesis of the severe cyclone is likely to occur in an active phase of the MJO. Liebmann et al. (1994) also have shown that some of the cyclones in the tropical Indian Ocean basin are originated from MJO. However, while the MJO occurs every year, the intensification to the stage of a severe cyclone over the Bay of Bengal is achieved only in few years.

We expect the impact of the cyclone on the pressure gradient and the wind along the Sumatra coast to last for a few days. Long term mean of sea level pressure (SLP) in the month of May is shown in Fig. 8. It can be seen that there is a strong north-south SLP gradient across the equator, with low pressure over the northern hemisphere and high pressure over the southern hemisphere. This is associated with the mean monsoon system. Hence it is expected that the wind along the Sumatra coast will be upwelling favourable southeasterlies from April–May months. However, it can be seen from Fig. 8 (which is the SLP anomaly

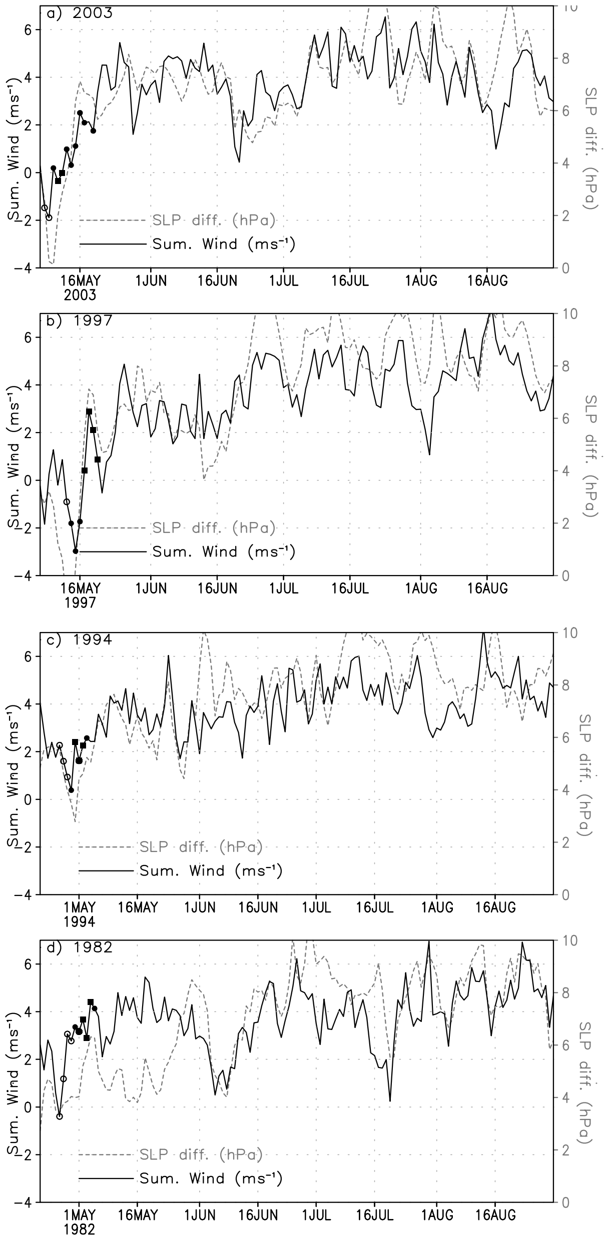

Fig. 7. Variation of the zonal component of surface wind off Sumatra coast (ms^{-1}) and difference in SLP (hPa) averaged over the bay (80°E–100°E, 5°N–20°N) and region south of EEIO (90°E–110°E, 10°S–20°S) during and after the severe cyclones of 1982, 1994, 1997 and 2003. The symbols on the surface wind curve represent the intensity of the bay system: open circle for depression, large closed circle for cyclone and square for severe cyclone.

averaged during severe cyclones in 1982, 1994, 1997 and 2003), that the severe cyclones over the Bay decrease the SLP over the bay further, which enhances the SLP gradient across the EEIO. Hence it is the interaction of these transient systems (severe cyclones) with the mean monsoon system that enhances and maintains the strong SLP gradient across the EEIO and hence upwelling favourable southeasterlies along the Sumatra coast as seen in Fig. 7.

We believe that the more important impact may be the one on the IWVC over the EEIO, which we consider next.

4.1.2. Impact on the integrated water vapour content over EEIO. The time series of the IWVC averaged over the EEIO during and after the cyclones in May 1997 and May 2003 are shown in Fig. 9. It is seen that there is a sharp reduction in the IWVC over the EEIO with the intensification of the bay systems in 1997 and 2003. The degree of suppression of the IWVC depends upon the track of the cyclone and the intensity. Rosendal (1998) suggested that deep convection associated with cyclones or ITCZ leads to strong descent of air in the opposite hemisphere. Composite pattern of OLR anomaly and latitude-height sections of vertical and meridional velocity anomalies during 3 d for each of the severe cyclones are shown in Fig. 10a as negative OLR anomaly. Intense convection over the Bay of Bengal associated with the severe cyclones is clearly seen as an intense negative OLR anomaly. Latitude-height section of the zonally averaged vertical (pressure) velocity over 85°E–110°E, shown in Fig. 10b clearly depicts the strong ascent associated with the severe cyclones. The ascending air from this region of deep convection then diverges at upper levels (Fig. 10c). This is evident from latitude-height section of meridional velocity averaged over 85°E–110°E (Fig. 10b). It can be further seen from latitude-height section of vertical velocity that the southward flowing air from the severe cyclone then descents over the EEIO. This descent is observed even at lower levels. The reduction of IWVC over EEIO seen in Fig. 9 is due to the descent of this dry air. This leads to a suppression of convection over EEIO (Fig. 10a).

In 1994, the intensity of the cyclone decreased rapidly relative to those in 1997 and 2003. However, soon thereafter, deep convection was generated again over the bay. The decrease in the IWVC over the EEIO by the combined effect of these two systems was only a little less than that in 1997 (Fig. 9c). In 1982 also, the duration and intensity of the severe cyclone over the bay was not as high as in 1997. However, like in 1994 there was another cyclone subsequently (during 30 May–5 June) over the Bay and the combined effect of the two systems resulted in comparable suppression of IWVC over the EEIO (Fig. 9d).

The suppression of IWVC over the EEIO after the second week of May in 1997 and 2003 is exceptionally large in comparison to all other years (Fig. 9). In both these years, the value of IWVC over the EEIO became lower than that in any other years by 22 May and continued to be very low until mid June. Even though the initial suppression of IWVC is associated with severe cyclones over the Bay of Bengal, we need to understand why it continued to be suppressed. Kajikawa et al. (2001) had suggested that the anomalous convection over the SCS/PS during June–July leads to suppression of convection over the EEIO and the maritime continent. The suppression of convection over

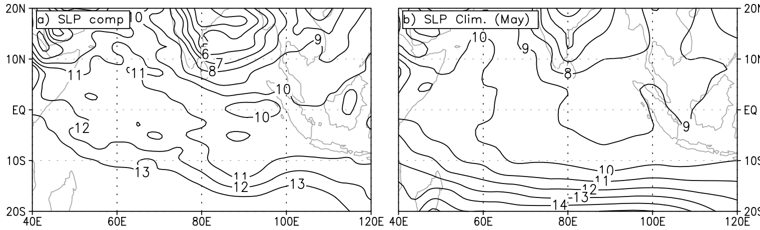

Fig. 8. (a) Composite map of SLP (−1000 hPa) during the severe cyclones which preceded positive IOD events. Composite is made by considering 3 days during which cyclones were severe in 1982, 1994, 1997 and 2003. (b) SLP climatology (−1000 hPa) for the month May.

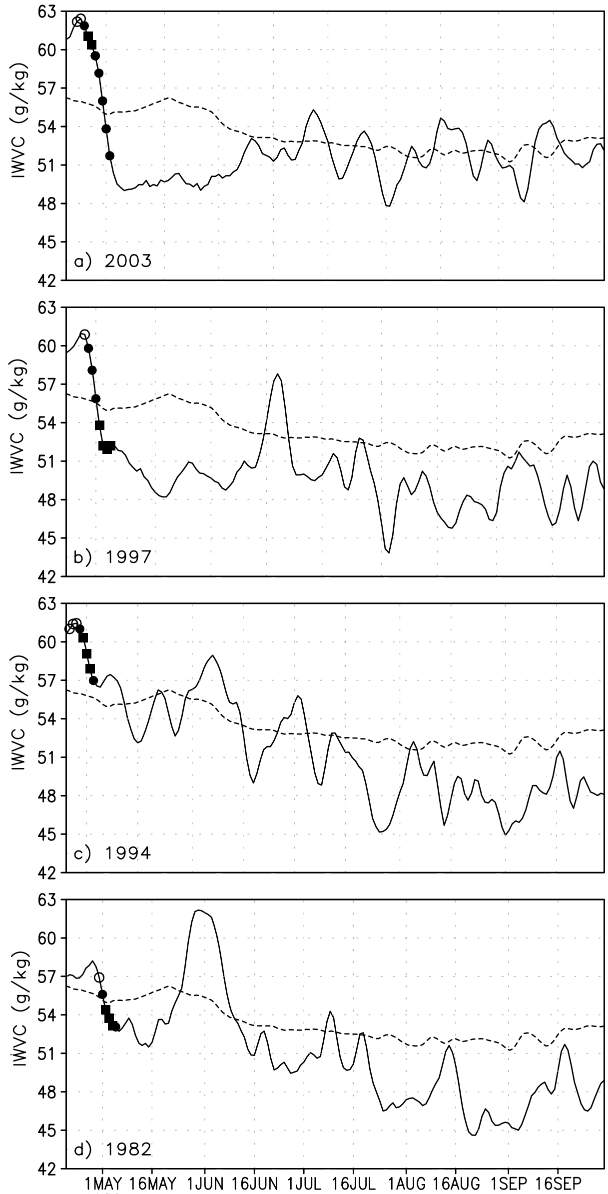

Fig. 9. Daily variation of the IWVC (g kg^{-1}) over EEIO during and after the severe cyclones in (a) 2003, (b) 1997, (c) 1994 and (d) 1982. A three point smoothing is performed while plotting. The symbols on the IWVC curve represent the intensity of the bay system: open circle for depression, large closed circle for cyclone and square for severe cyclone.

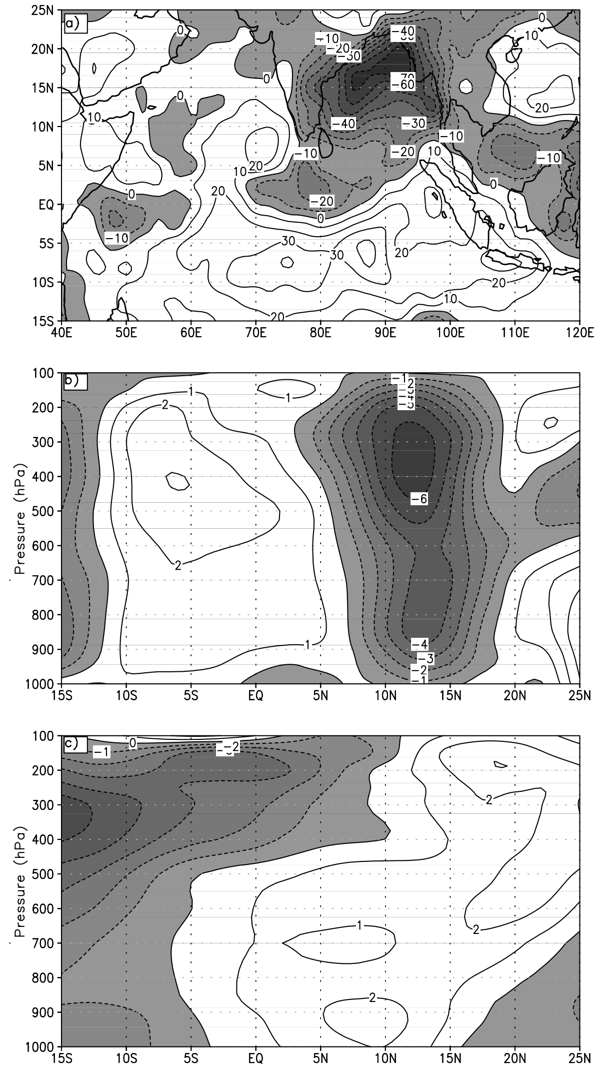

Fig. 10. Composite maps of (a) OLR anomaly (Wm^{-2}) and latitude-height sections of (b) vertical velocity anomaly ($\times 10^5$ hPa s^{-1}) and (c) meridional velocity anomaly (ms^{-1}), averaged over 85°E–110°E. Composites are made by considering 3 days during which cyclones were severe in 1982, 1994, 1997 and 2003.

the EEIO due to deep convection over the SCS/PS is evident from the negative correlation between OLR over the SCS/PS, and that over the EEIO in May (Fig. 11). Interestingly, the cyclones over the Bay of Bengal, both in 1997 and 2003, were followed by deep convection associated with typhoons over the SCS/PS.

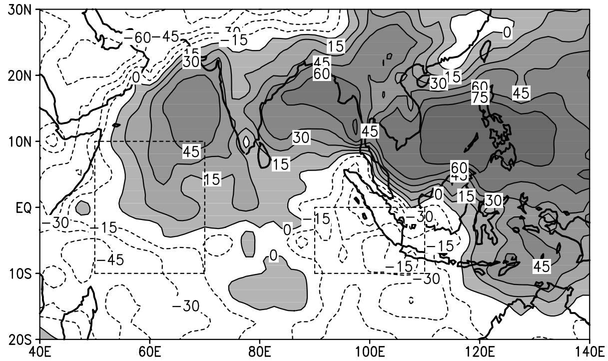

Fig. 11. Correlation (×100) of OLR with respect to OLR over the SCS/PS (110°E–130°E, 5°N–20°N) for May during the period 1979–2004. Correlation coefficients greater than 0.33 are significant at 95% confidence level.

Daily variation of average OLR over the bay, SCS/PS (110°E 130°E, 5°N 20°N) and EEIO during May–June period for 1997 and 2003 are shown in Fig. 12. It can be seen that associated with the intensification of deep convection over the Bay of Bengal, the convection over EEIO started weakening in both the years. This is followed by intensification of convection over the SCS/PS, which is associated with further suppression of convection over the EEIO (Fig. 12). Hence the continuation of suppression of convection over the EEIO up to early June in 1997 and 2003 could be a combined effect of deep convection over the Bay of Bengal and SCS/PS.

It should be noted that suppression of convection over the EEIO is the first step in the triggering mechanisms proposed by Annamalai et al. (2003), who attribute it to warm SST anomalies over equatorial western-central Pacific, and Kajikawa et al. (2001), who suggest that it arises due to enhanced convection over the SCS/PS. Kajikawa et al. (2001) suggest that the suppression convection leads to enhancement of the upwelling favourable winds off Sumatra. We suggest that the suppression over EEIO associated with severe cyclones over the bay could also lead to the weakening of the westerly winds over CEIO and hence triggering of the dipole events.

4.2. *Processes involved on a longer time scale*

In this section we suggest how the suppression of convection over EEIO which is an atmospheric conditions favourable for the triggering of IOD events, could be sustained for a period of time considerably longer than the synoptic scale of a few days.

In order to study the daily variation of convection over regions such as the EEIO and WEIO we use a convection index (CI), which is a measure of the intensity and the horizontal extent of deep convection over a region. On the daily scale, regions of deep convection can be delineated on the basis of OLR. However, average values of OLR for a region on the weekly and larger scales can be relatively high despite the occurrence of some days of deep convection over some subregions. Hence we use CI, which is derived from the 2.5 × 2.5 deg resolution OLR data as follows. We assume that, on any day, deep convection occurs only over the grid points for which OLR is below 200 Wm^{-2} and take the difference of OLR value from 200 Wm^{-2} to represent the intensity of deep convection at the grid point. The CI for a specific region (such as the Bay), for a particular day is calculated as the sum of the intensity of convection over all grid points with deep convection (i.e. over all the grid points with OLR less than 200 Wm^{-2}.

In order to understand the impact of the suppression of convection over EEIO, consider the variation of the climatological

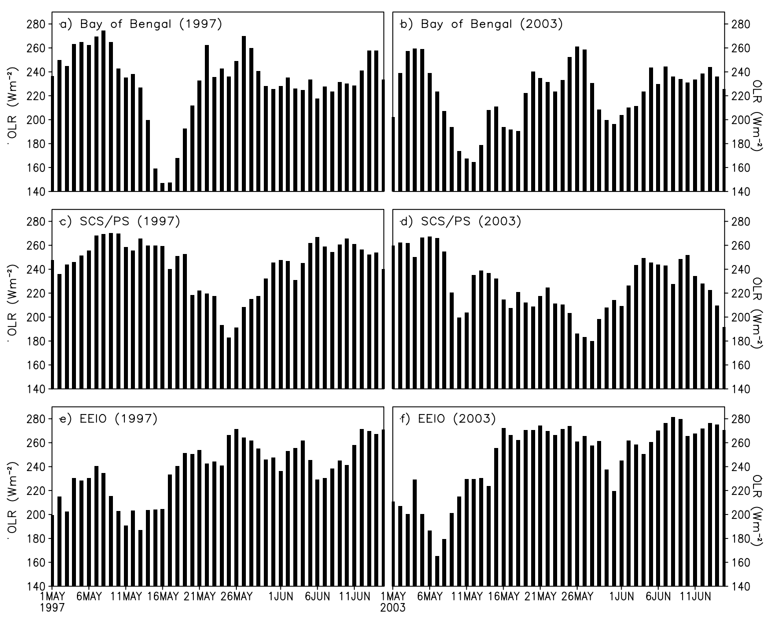

Fig. 12. Daily variation of OLR (Wm^{-2}) over Bay of Bengal (85°E–95°E, 5°N–20°N), SCS/PS (110°E–130°E, 5°N–20°N) and EEIO for May–June 1997 (left) and 2003 (right)

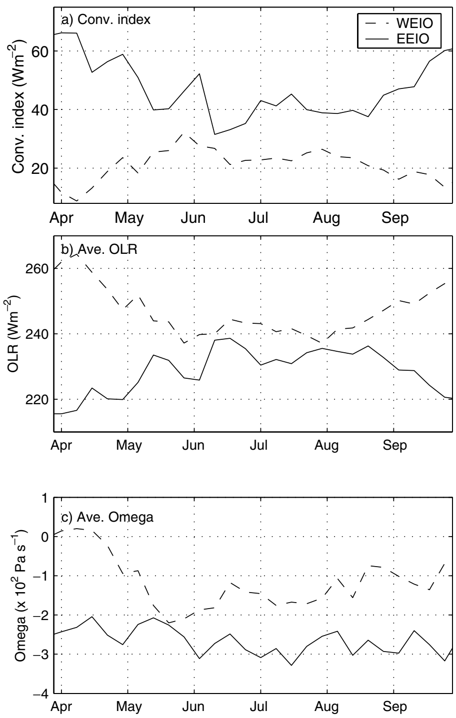

Fig. 13. Daily variation of convection index (in units of 5×10^2 Wm^{-2}) over Bay of Bengal (85°E–95°E, 5°N–20°N), SCS/PS (110°E–130°E, 5°N–20°N) and EEIO for May–June 1997 (left) and 2003 (right). Variation of the climatology of (a) weekly average CI (in units of 5×10^2 Wm^{-2}) and (b) average OLR (Wm^{-2}) and (c) daily mean vertical velocity ($\times10^3$ hPa s^{-1}) at 500 hPa over EEIO (solid) and WEIO (dashed) during April–September

average OLR and the CI over WEIO and EEIO (Fig. 13). The average OLR over EEIO is less than the average OLR over WEIO (Fig. 13a) and the average CI over EEIO is higher than the average CI over WEIO throughout the boreal spring and summer (Fig. 13b). Thus climatologically, convection is more favoured over EEIO than over WEIO. Consistent with this, the mean sea level pressure over the EEIO is lower than that over WEIO and the mean zonal component of the wind over CEIO is westerly (Fig. 4a). We note that the average OLR over WEIO is high during March–mid-April, then decreases up to the end of May and then oscillates around 240 Wm^{-2}, up to the end of August. Weekly average OLR of 240 Wm^{-2} and above would be taken to indicate the absence of deep convection. However, the CI over WEIO increases from less than 10×10^{-3} Wm^{-2} in April to over 20×10^{-3} Wm^{-2} in May and stays at that level until the end of July (Fig. 13b). This suggests that deep convection does occur over WEIO, albeit intermittently. This is consistent with the climatological vertical velocity over the EEIO being close to that over WEIO (Fig. 13c), during May–August. In fact the intermittent occurrence of convection over WEIO and EEIO is clearly seen in the variation of the CI over these regions during 1990 (Fig. 14) for which the average monthly OLR values over WEIO and EEIO are close to the climatological averages.

It is seen from Fig. 14c that the major difference from climatology during the dipole event of 1997 is that convection over WEIO is favoured vis-a-vis that over EEIO. On the other hand, there are some years in which convection over EEIO is favoured compared to that over WEIO (eg. 1985, as shown in Fig. 14a). The average OLR (the CI) and vertical velocity over EEIO are close to that over WEIO during mid-May to mid-August (Fig. 13) suggesting that the atmospheric conditions are almost equally favourable over the two regions for supporting convection in this period. The strength of the climatological westerly winds over CEIO, which reflects the east-west gradient in the convection, also decreases from mid-May and the magnitude is small (less than 1 ms^{-1}) throughout June–August (Fig. 4). Since both EEIO and WEIO are favourable for convection from May onwards, suppression of convection over the EEIO leads to enhancement of convection over the WEIO and weakening of surface westerlies over the CEIO as seen from Fig. 15. This implies anomalous convergence over the WEIO leading to enhanced convection which in turn strengthens the anomalous easterlies over the CEIO further, until the winds become easterly (Fig. 15). As the surface easterlies converge over the WEIO, convection is favoured over WEIO relative to EEIO. This east-west gradient in convection maintains the surface easterlies over the CEIO. These surface easterlies trigger upwelling favourable Kelvin Waves in the EEIO. Together with the coastal upwelling due to anomalous southeasterlies along the Sumatra coast, these Kelvin Waves leads to anomalous cooling in the EEIO. We suggest that this positive feedback between convection and circulation can lead to suppression of convection over EEIO for periods of several weeks and hence trigger a positive IOD event. This hypothesis has to be tested further with detailed dynamic and thermodynamic budget studies.

5. Severe cyclones which were not followed by an IOD event

Some of the severe cyclones over the bay in April–May are not followed by a positive IOD event. An example is the cyclone of 1990. It is interesting to note that the track of the cyclone of 1990 was different from the other tracks in that they were primarily over the southwestern part of the bay, that is, west of 85°E (over which the cyclone became severe) and culminated over the Indian region (Fig. 6). Average OLR anomaly and latitude-height sections of vertical and meridional velocity anomalies during 11–13 May 1990 (when the cyclone had moved away from the southern Bay), are shown in Fig. 16. In contrast to the patterns in Fig. 10, there is deep convection and strong ascending motion over the EEIO while the eastern parts of Bay of Bengal is characterized by positive OLR anomaly and associated

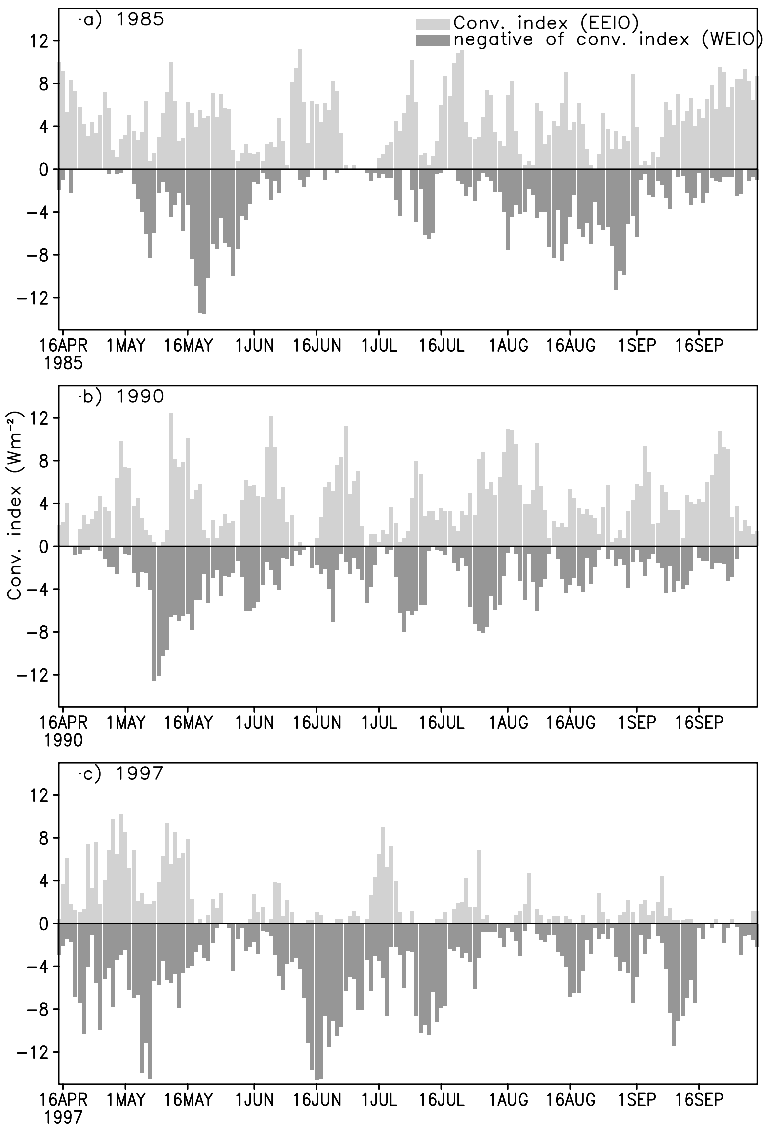

Fig. 14. Daily variation of CI (in units of 5×10^2 Wm^{-2}) over the EEIO (light grey) and negative of the CI of WEIO (dark grey) over the CEIO during 20 April–31 July for (a) 1985, (b) 1990 and (c) 1997.

descending motion. Not surprisingly, the magnitude of the suppression of the IWVC over the EEIO associated with this cyclone was much smaller than of that in 1997 or 2003 (Fig. 15). The meridional pressure gradient across EEIO and hence the magnitude of the enhancement of southeasterlies off Sumatra coast, associated with these cyclones was also not as large. Thus for a severe cyclone over the bay to trigger a dipole event, the impacts in terms of suppression of the IWVC over the EEIO and enhancement of southeasterlies have to be large.

6. Evolution of IOD events: Role of pre-conditioning of equatorial Indian Ocean

It is well known that the impact of surface winds on the SST depends upon the depth of thermocline. The upwelling in the EEIO work efficiently to decrease the SST only when the thermocline is shallow (Vinayachandran et al., 1999). The role of the depth of the upper layer of the ocean can be clearly seen by comparison of the cases of 1997 and 2003. Although the wind forcing in 2003 is almost identical to that in 1997 (Fig. 5), it can be seen that the cooling in the EEIO during June–July 2003 is not as much as in June-July 1997 (Fig. 3a and b). In 1997, the SST in the EEIO dropped well below 28 °C by mid July and continued to be below this threshold thereafter. On the other hand, in 2003, SST in the EEIO was above 28 °C up to the end of July and slightly below 28 °C in the first two weeks August. With the revival of convection over the EEIO in mid August, SST started increasing and the IOD event of 2003 was terminated. In 1994 the spring jet was very weak and the thermocline in the EEIO was very shallow during April–May

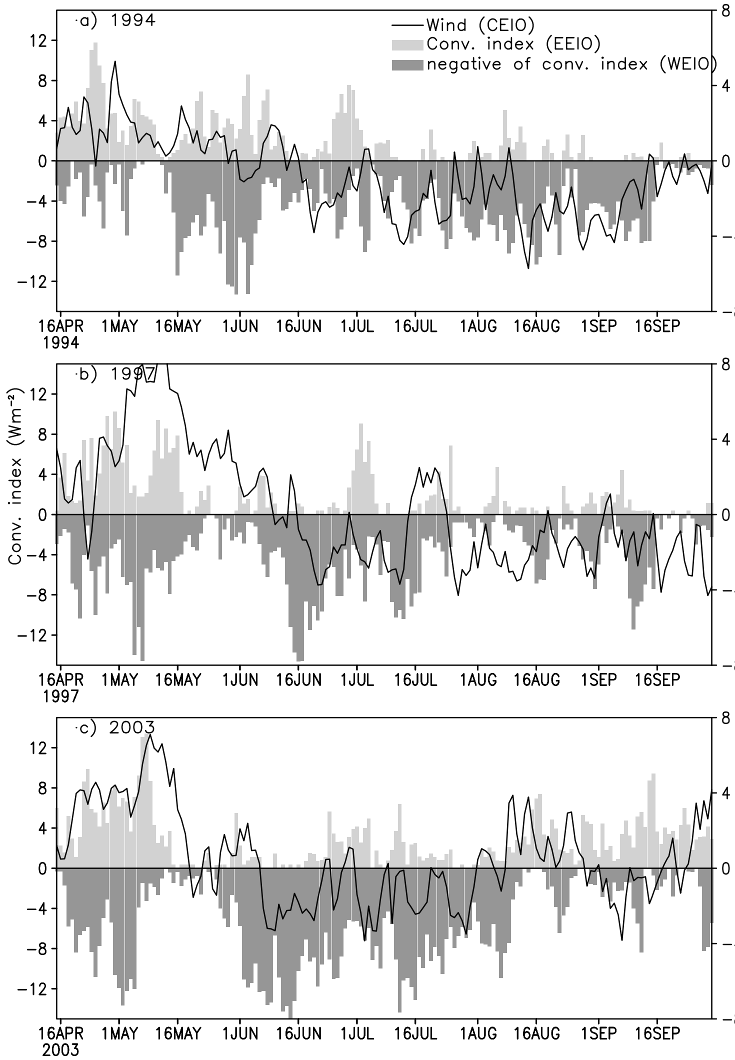

Fig. 15. Daily variation of CI (in units of 5×10^2 Wm^{-2}) over the EEIO, negative of the CI of WEIO and surface zonal wind (ms^{-1}) over the CEIO during 15 April–30 September for (a) 1994, (b) 1997 and (c) 2003.

(Vinayachandran et al., 1999, 2002). The SST of the EEIO started decreasing from early April itself. A severe cyclone (which was weaker than those in 1997 and 2003) occurred in the last week of April and this decrease in SST amplified and a positive IOD event was triggered. The experience of 1994 and 2003 events suggests that even though the severe cyclones can trigger positive IOD events, its further evolution into mature phase depends on other factors like the background state of the ocean. Using coupled model simulations, Behera et al. (2006) have shown that the cold anomalies in the western tropical Indian Ocean during the boreal fall of the year preceding IOD events, which do not cooccur with El Nino events, propagate to the EEIO. They suggest that, this creates a favourable condition for the triggering of IOD events. Clearly, it is important to understand the role of preconditioning of the Indian Ocean in the development of IOD events.

7. Conclusions

In this paper we suggest that severe cyclones over the Bay of Bengal during April–May could trigger positive dipole events. We show that each positive IOD event is preceded by such a cyclone though not all severe cyclones are followed by IOD events. We show that severe cyclones can lead to the intensification of southeasterlies off-Sumatra coast and suppression of water vapour content and hence convection over the EEIO. Whether the cyclone does trigger a dipole event depends on the intensity and track which determine the extent to which the water vapour content over EEIO is suppressed. We suggest that the once the suppression of convection over the EEIO occurs, it can lead to more convection over the WEIO. This leads to the weakening of westerly surface wind over the CEIO and hence enhancement of convergence to the WEIO. This positive feedback can

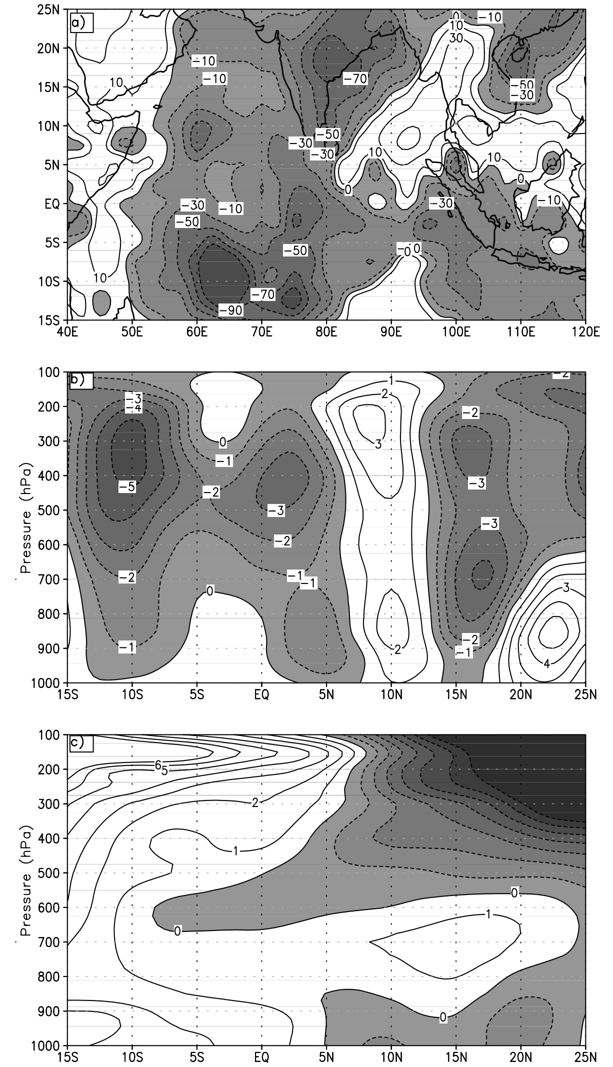

Fig. 16. Average patterns of (a) OLR anomaly (Wm^{-2}) and latitude-height sections of (b) vertical velocity anomaly ($\times\ 10^3$ Pa s^{-1}) and (c) meridional velocity anomaly (ms^{-1}), averaged over 85°E–110°E during 11–13 May 1990.

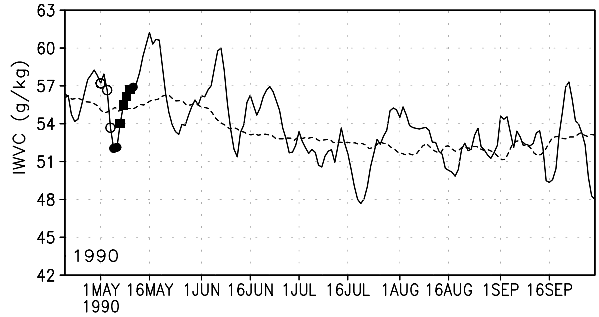

Fig. 17. Variation of the IWVC (g kg^{-1}) over EEIO during and after the severe cyclone in 1990. The symbols on the IWVC curve represent the intensity of the bay system: open circle for depression, large closed circle for cyclone and square for severe cyclone

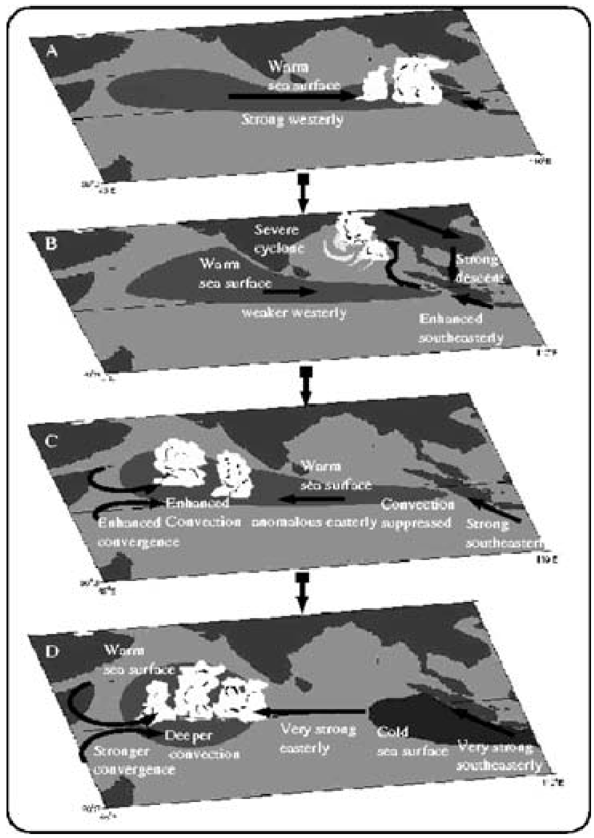

Fig. 18. A schematic diagram representing the evolution of a positive IOD event.

lead to further enhancement of convection over the WEIO until the wind over the CEIO becomes easterly and the convection over the EEIO remains suppressed for periods much larger than the synoptic time scale characterizing the cyclone. The strong upwelling caused by the easterlies along the equator and southeasterlies along the Sumatra coast decrease the SST of EEIO very rapidly and a positive dipole event is triggered (Fig. 18). Clearly, the hypothesis proposed has to be tested with studies with models.

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