Influence of El Nino Southern Oscillation and Indian Ocean Dipole in biennial oscillation of Indian summer monsoon

4.1 Introduction

The main contributors to the interannual variability of Indian summer monsoon rainfall (ISMR) is the large scale forcing from El Nino Southern Oscillation (Webster and Yang, 1992; Ju and Slingo, 1995; Webster et al., 1998), tropical Tropospheric Biennial Oscillation (Meehl, 1997; Chang and Li, 2000; Meehl and Artblaster, 2002a; Meehl et al., 2003; Wu and Kirtman, 2004; 2007) and Indian Ocean Dipole (Saji et al., 1999; Webster et al., 1999; Ashok et al., 2004). During an El Niño event, the locus of maximum SST in the Pacific Ocean shifts eastward, bringing more precipitation over the central and eastern Pacific Ocean. During these periods the eastern Indian Ocean, Indonesia, and south Asia are in the subsiding part of the Walker circulation that has shifted eastward from its climatological position (Webster et al., 1998). Clearly, there is a connection between the Asian monsoon and ENSO, but it is not possible to predict the strength of the monsoon solely from the phase of ENSO, as these monsoon–ENSO correlations have variable lag-lead times (Webster and Yang, 1992). With regard to monsoon–ENSO relationships, Torrence and Webster (1999) noted the demise of ENSO-based predictive relationships in the 1920–1960 period. During the 1997/98 El Niño events, the Indian rainfall remained essentially normal (Webster et al., 1998;
Kumar et al. (1999) gives the evidence of weakening of the reverse monsoon-ENSO relationship in recent decades. Years of heavy Indian summer monsoon rainfall tend to be followed by years of diminished rainfall. This well-known biennial periodicity of monsoon rainfall is extensively related to tropospheric biennial oscillation (TBO), which appears in a wide range of atmospheric variables including rainfall, surface pressure, wind, and SST of tropical regions (Meehl, 1987). Meehl (1997), Meehl and Arblaster (2001; 2002a) described an air–sea negative feedback mechanism in which warm spring SSTs in the Indian Ocean enhance atmospheric convection. The ensuing stronger monsoon leads to greater than average wind strength, increased Ekman transports and vertical mixing, and higher heat loss by evaporation throughout the summer monsoon season, which causes subsequent cooling of the ocean surface. The low ocean temperatures persist for one year until the next summer season. The lowered SSTs are associated with less convection than the previous spring, producing weakened winds, reduced heat loss by evaporation, and less mixing, which leads to higher SSTs than the year before. The cycle is thus repeated. Clearly a better understanding of the ocean–atmosphere and land–atmosphere feedbacks that may give rise to the TBO can improve predictions of monsoon strength.

Indian Ocean Dipole mode (Saji et al., 1999) is the coupled ocean-atmosphere interaction mode in Indian Ocean, which may induce unusual rainfall not only in surrounding areas, but other parts of the globe also (Saji and Yamagata, 2003). Positive IOD is characterized by anomalously cool SST in the southeast equatorial Indian Ocean off Sumatra due to unique coastal upwelling and anomalous warming in the west equatorial Indian Ocean. Cool water in the east Indian Ocean gives easterly anomalies along the equator. This conditions develop in the boreal summer season and peak in boreal autumn season. The reverse situation is the negative Indian Ocean dipole mode.

An index to quantify the IOD has been defined (Saji et al., 1999) as the SST difference between the tropical western Indian Ocean (50°E-70°E, 10°S-10°N) and the tropical southeastern Indian Ocean (90°E-110°E, 10°S-equator). Ashok et al. (2001) using a 41
month sliding correlation of both IOD index and Nino3 SST with ISMR showed that correlations have decadal variations and in decades when IOD-ISMR correlation is strong nino3-ISMR correlation is weak and vice versa. The presence of a positive IOD has facilitated normal or excess rainfall over the Indian region during the summers, despite the simultaneous occurrence of the negative phase of the Southern Oscillation by bringing IOD induced convergence over Indian Ocean replacing the ENSO induced divergence there (Behera et al., 1999; Webster et al., 1999). On the other hand, during some years the prevailing negative IOD and El Niño have combinedly caused an anomalously deficit rainfall during the monsoon season. The ISMR anomalies also depend on the relative intensities of the IOD and the El Niño/La Niña events.

The period of ENSO is estimated to be of 3-7 years. So these unusually long period episodes may weaken the biennial signal (Reason et al., 2000). But there is enough evidence for the biennial tendency of the ENSO episodes (Kiladis and Van Loon, 1988; Kiladis and Dias, 1989; Rasmusson et al., 1990). Thus the definition of TBO years will include some of the ENSO onset years also. In addition, there are other years, including many Indian Ocean dipole (or zonal mode) events, that contribute to biennial transitions. Ropelewski et al. (1992) related the TBO to the southern oscillation. Meehl and Arblaster (2001, 2002a) projected Pacific Ocean SST as the major contributor for TBO mechanism. Based on modeling studies, Chang and Li (2000), Li et al. (2001), suggested that Indian Ocean alone can produce TBO transition for monsoon, but the TBO amplitude will be weaker. By including Pacific in the model the anomalies become more pronounced. Meehl et al. (2003) showed that with and without ENSO onset years, TBO cycle has only difference in the magnitude of anomalies. According to him there exist some other remote mechanisms like ENSO, which can contribute to biennial variability. Thus earlier studies have the opinion that with and without ENSO years there may be only a difference in magnitude of the observed anomalies in TBO composites. In fact TBO and ENSO signal strength is closer in magnitude in Indian Ocean than in Pacific (Reason et al., 2000). Almost all the previous studies explain the TBO cycle on the basis of ENSO biennial cycle. Yu et al. (2003) argued that Pacific Ocean SST anomaly is more important in in-phase Indian to Australian monsoon
transition, while Indian Ocean plays a dominant role in out of phase transition from Australian to Indian summer monsoon. Wu and Kirtman (2004) with the help of coupled GCM produced biennial cycle for wet minus dry years accompanied by ENSO years. They are of the opinion that there is biennial cycle that are not associated with ENSO, during which other factors like Indian Ocean SST anomaly may play a role and that should be studied separately. A power spectrum of the monthly Dipole Mode Index shows significant peaks in the 2–3-year range, and smaller peaks in the 3–5 year range.

Moreover according to Loschinger et al. (2003) ocean dynamics in the Indian Ocean is associated with springtime transitions as well as SST anomalies in the equatorial regions. Thus ocean dynamics which are associated with extremes of IOD is natural parts of TBO evolution. Large-scale meridional heat transport in the Indian Ocean modulates the north Indian Ocean SST on seasonal time scales and may be involved in interannual variability of Asian monsoon by affecting the heat content and SST anomalies as part of biennial cycle of SST modulation. Loschinger et al. (2003) hypothesized that Indian Ocean Dipole Zonal Mode and TBO as integral parts of self regulating systems of monsoons that are acted as negative feedbacks to keep the variability of monsoon both on annual and interannual scales. Thus TBO includes both ENSO and IOD years. But how these dynamic features of Pacific and Indian Ocean regulate TBO and interannual variability of Indian summer monsoon is not fully understood.

4.2 Objectives of the study

Almost all the earlier studies were looking at the biennial oscillation of Indian summer monsoon on the basis of its interaction with the Pacific by ENSO. The Indian and Pacific Ocean processes are found to influence TBO and interannual variability of Indian summer monsoon. The present study analyses the modification of biennial cycle of Indian summer monsoon with ENSO and IOD, those represents the dynamic processes in the Pacific Ocean and Indian Ocean respectively.
4.3 Data and methodology

The present study uses zonal and meridional wind, 200 hPa velocity potential and vertical velocity obtained from National Center for Environmental Prediction /National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al., 1996) for the period 1950-2005. NOAA outgoing longwave radiation (OLR) data set for the period 1974-2005 with a one year data gap at 1978 is used for studying the movement of atmospheric convection zones. Sea surface temperature for the study period is also obtained from NCEP. ISMR index, which is the area averaged June to September rainfall of 306 stations well distributed over India, (Parthasarathy et al., 1994) and updated for making it to period 1950-2005 is used to represent the strength of Indian summer monsoon in particular years.

By using ISMR value, we defined years as strong (weak) TBO year, if it has more (less) ISMR anomaly than the previous and following year. ENSO years are then identified from area averaged SST anomalies of \textit{nino3.4} region ($5^\circ$N-$5^\circ$S, $170^\circ$-120$^\circ$W), whose five month running mean should be at least $\pm 0.5^\circ$C for two consecutive seasons after March- April- May season over the following one year period. In order to assess the effect of ENSO in interannual variability of Indian summer monsoon, TBO years are analysed separately for those associated with TBO (ENSO-TBO) and independent of TBO (normal TBO). Similarly the SST, wind and circulation features of TBO years in presence of IOD are also analysed. The analyses were carried out from present strong monsoon (JJA0), following autumn (SON0), winter (DJF0), and spring (MAM1) and next monsoon (JJA1).

4.4 Results

The ENSO-TBO years and IOD-TBO years are studied separately to investigate the role of ENSO and IOD on TBO. The years in which IOD and ENSO co-occurs are also analyzed here.

4.4.1 Effect of ENSO on TBO

In order to investigate the effect of ENSO on TBO composite analysis of strong minus weak TBO years compared for TBO years both in presence and absence of ENSO. As
all the ENSO years are not TBO years, first the peculiarities of ENSO years are compared for all ENSO years using SST anomalies of $Nino3.4$ region.

### 4.4.1.1 Evolution of ENSO anomalies in the Pacific

From the above method of defining TBO years we have 35 TBO years in the 56 year study period. In this 14 are associated with ENSO in the Pacific. Thus all the ENSO years are not TBO years. In order to investigate the peculiarities of ENSO years, which are TBO years we have made analysis of SST anomaly formation of ENSO years of 1950-2005 period in figure 4.1

![Figure 4.1: Time series of Nino3.4 region SST anomalies for years (a) TBO years associated with El Nina , (b) associated with La Nina and (c) ENSO years independent of TBO.](image)
In the analysis onset time of ENSO is defined as the month in which the anomaly of the region is above ± 0.5°C. In the ENSO-TBO years the onset of ENSO (both El Nino and La Nina) is in any of the month (April or May) before the onset of Indian summer monsoon in June (figure 4.1 a and b). In other ENSO years the onset is at the time of onset of summer monsoon or later (see figure 4.1c). In all ENSO years the SST anomaly attains maximum by the next winter. The spring season is most important for the TBO associated with Indian summer monsoon. Thus onset of the ENSO must be in the spring season to influence the interannual variability of Indian summer monsoon through TBO mechanism.

4.4.1.2 TBO mechanism in presence and absence of ENSO

Strong minus weak composite of various oceans –atmosphere parameters for ENSO-TBO and normal TBO years are carried out from JJA0 to JJA1 and the results are presented in the form of comparison.

4.4.1.2.1 SST pattern

Figure 4.2 shows the TBO anomalies from JJA0 to JJA1 for ENSO-TBO (left panels) and normal TBO years (right panel). During the strong summer monsoon season (JJA0) entire north Indian Ocean is cooled in ENSO TBO years along with equatorial east and central Pacific and western Pacific is warm. But in the absence of ENSO, monsoon cooling is confined to Arabian Sea, Bay of Bengal and equatorial east Indian Ocean. In Pacific Ocean anomalies are reverse in absence of ENSO with cool west Pacific and warm east Pacific. In the post monsoon season (SON0), anomalies strengthens in both the cases.

By the boreal winter (DJF0) entire Indian Ocean is cooled and cooling from equatorial east Pacific reaches extreme west if TBO is associated with ENSO in the Pacific. But in the absence of ENSO, cooling in the north Indian Ocean decreases and confines to west side only and warming in the east Pacific reaches extreme west. Cooling continues to next spring (MAM1) in the Indian Ocean and extreme east Pacific starts warming in presence of ENSO, while entire east and central Indian Ocean is warm in normal TBO
years and equatorial west Pacific and extreme east Pacific starts cooling. Indian Ocean remains cool for next year summer (JJA1) of TBO year in presence of ENSO and warming in the east Pacific extends west. In the absence of ENSO the north and southeast Indian Ocean is warm and cooling from the east Pacific extends to central Pacific.

Thus the evolution and movement of TBO anomalies are different in both these cases. In TBO scale, Pacific Ocean anomalies are opposite in both the cases and the effect of monsoon cooling is confines to oceanic regions close to land region and upto winter only. The ENSO-TBO pattern resembles all TBO cases.

4.4.1.2.2 850 hPa wind pattern

The strong minus weak composite of 850 hPa wind for ENSO-TBO and normal TBO for the present summer (JJA0), following winter (DJF0) and next summer (JJA1) is given in figure 4.3. During JJA0 westerlies in the north Indian Ocean are stronger in the absence of ENSO. In these case the westerlies doesn’t extend to north of Australia.
during boreal winter. Thus southeast movement of anomalies is absent in non-ENSO TBO years. But by next summer wetsrelies reverse to easterlies in north Indian Ocean and Indian subcontinent.

Figure 4.3: strong minus weak TBO year composite of wind at 850 hPa for TBO year summer (JJA0), coming winter (DJF0) and next year summer (JJA1).

4.4.1.2.3 Sea level pressure (SLP) pattern

Figure 4.4 illustrates the TBO cycle of SLP in presence and absence of ENSO from JJA0 to JJA1. During the summer season of strong TBO year (JJA0), SLP is low over India and entire Indian Ocean extending to Indonesia and Australia in the presence of ENSO. In normal TBO years SLP lowers in these regions, but the low pressure center is northwest of India. By the boreal winter (DJF0) the minimum SLP anomaly moves to Australia and northwest Pacific making Indian to Australian monsoon transition in ENSO-TBO years. In the absence of ENSO, the low pressure region is in the western Indian Ocean and northwest Pacific Ocean and SLP is high over Australian monsoon region. By the next year summer (JJA1) SLP is high over India and western Indian Ocean and is low in the east Pacific in the presence of ENSO. But in the absence of ENSO SLP remains low over India and western Indian Ocean in JJA1.

Thus the transition and reversal of SLP anomalies are possible in presence of ENSO only for TBO years.
4.4.1.2.4. 200 hPa velocity potential and convergence

200 hPa velocity potential and convergence pattern for seasons from JJA0 to JJA1 is shown in figure 4.5. Negative anomaly of velocity potential in the upper level is associated with convergence and upward motion in the lower levels.

Upper level divergence is over India and west Indian Ocean in JJA0 indicating upward motion over Indian monsoon region and downward motion over east Pacific when ENSO is present along with TBO. In normal TBO years, the upward motion center in the equatorial Pacific over dateline is stronger than that over the Indian monsoon region. Downward motion is over the extreme east Pacific and equatorial east Indian Ocean. In the post monsoon season (SON0) upward motion zone moves to Indonesian region and downward motion center is over the equatorial east Pacific for ENSO-TBO years. The southeast movement of anomalies are absent for TBO in the absence of ENSO onset years and convergence/divergence centers remains as such in the post monsoon. The convergence center extends to northwest Pacific from Australian monsoon region making Australian monsoon strong in DJF0 for ENSO-TBO composites. But absence of ENSO brings divergence over Australia, making in phase Indian to Australian monsoon transition weak in the absence of ENSO.

By the next spring (MAM1) convergence zone moves further east in Pacific and divergence appears over equatorial Indian Ocean for ENSO-TBO years. In the absence of ENSO convergence is over African region and divergence is over extreme east Pacific. During the next summer monsoon (JJA1) convergence associated with TBO-ENSO is in the east Pacific and divergence is in the western Indian Ocean. If ENSO years are removed, pattern remains as MAM1 with convergence in the African region and divergence in the east Pacific.
Figure 4.4: Strong minus weak TBO year composite of SLP for seasons from JJA0 to JJA1. Left panel for TBO years, which are ENSO onset years also and right panel for non-ENSO TBO years.

Figure 4.5: Strong minus weak TBO year composite of 200 hPa velocity potential (shaded) and convergence(vector) for seasons from JJA0 to JJA1. Left panel for TBO years, which are ENSO onset years also and right panel for non-ENSO TBO years.
4.4.1.2.5 Convection anomalies and its propagation

Figure 4.6 shows the Hovmuller diagram showing the movement of north Indian Ocean OLR anomalies in TBO cycle. During ENSO only TBO years, convection starts in the western Indian Ocean by the beginning of strong TBO year and strengthens by onset of monsoon in June. The convection anomalies propagates eastward and by the next year weak monsoon the convection reaches close to dateline and convection is absent in the Indian Ocean region. Thus the TBO cycle is completed and convection moves from Indian region to central Pacific from strong TBO to weak TBO year. But in the absence of ENSO, convection starts in the extreme west Indian Ocean by onset of strong monsoon and it also moves east and by next year summer it is at 120°E, i.e., in the Indian Ocean region itself. Thus in both the presence and absence of ENSO, convection anomalies associated with Indian monsoon reverses, but its movement from one TBO to next year is different in both the cases.

Figure 4.6: Time-longitude diagram for strong-minus weak TBO year composite of OLR in the north Indian Ocean (equator to 25°N) for ENSO-TBO (left) and non-ENSO TBO (right) years, Months with zero indicates previous year, 1 current TBO year and 2 next year for both strong and weak TBO years.
4.4.1.2.6 Effect of ENSO on summer season atmospheric circulation cells

The summer season equatorial (10°S-10°N) zonal circulation for TBO year and year after the TBO is shown in figure 4.7. Equatorial Walker circulation has upward motion anomalies in the Indian Ocean and west Pacific in the positive phase of TBO associated with strong Indian summer monsoon (JJA0) in the presence of ENSO in the Pacific Ocean. In the absence of ENSO equatorial region over Indian Ocean has downward motion and upward motion at dateline and downward motion is seen in the extreme east Pacific.

During the next summer season (JJA1) ENSO-TBO years has downward motion in the western Indian Ocean and west Pacific and downward motion in the eastern Indian Ocean and east Pacific. But in the absence of ENSO, Indian region between 60°E and 100°E has upward motion anomalies and downward motion in the western and central Pacific (lower panels of figure 4.7). Thus the upward motion in the equatorial Indian Ocean in strong phase of TBO is the effect of ENSO.

Figure 4.7: height-longitude profile of zonal wind and negative of vertical velocity averaged over 10°S-10°N for JJA0 and JJA1, indicating equatorial zonal circulation.

Local Hadley circulation over monsoon region (60°-95°E) for three monsoon seasons, JJA-1, JJA0 and JJA1 is illustrated in figure 4.8 from the vertical profile of meridional wind and vertical velocity. The Local Hadley circulation over the Indian monsoon longitudes has upward motion in the tropics between 5°S and 30°N and downward
motion south of 5°S in presence of ENSO for strong TBO year. In the absence of ENSO in the Pacific, downward motion exists between 5°S and 10°N and upward movement is observed from 10°N to 40°N. In the preceding year summer anomalies reverses in both the cases with downward motion from 5°S to 30°N in presence of ENSO and upward motion close to equator and downward motion in the region between 10°N and 40°N in absence of ENSO.

Figure 4.8: height-longitude profile of meridional wind and negative of vertical velocity averaged over 60°E-95°E, indicating local Hadley circulation for three consecutive summer seasons. Left panel for ENSO-TBO years and (a), (b) and (c) for JJA-1, JJA0 and JJA1 and (d), (e) and (f) are their counter parts in the absence of ENSO

Thus the local Hadley circulation between the equatorial Indian Ocean and Indian monsoon region plays a role in TBO in the absence of ENSO. When ENSO is present, it brings additional upward/downward motion in the equatorial Indian Ocean and that over the monsoon region is similar in both the situations.

4.4.2 Role of IOD in TBO

In the 56 year period from 1950-2005 we have 10 positive dipole years and 9 negative dipole years. But in the 10 positive dipole years, four of them (1961, 1967, 1994, 1997) are associated with strong TBO years and another three (1972, 1982, 2002) are associated with weak TBO and 1963, 1986 and 1991 are independent of TBO. Similarly for negative IOD also we have 1960, 1974, 1992 are associated with negative TBO and
1956, 1975 and 1988 are associated with strong TBO years and 1971, 1984 and 1996 are detached from TBO. Thus the interaction of TBO and IOD are two-fold. IOD years are associated with both ENSO and non ENSO TBO years. In presence of ENSO, both TBO and IOD are in opposite phase and in absence of ENSO IOD and TBO are in same phase.

4.4.2.1 In-phase and out of phase association of TBO and IOD

As the influence of IOD is two fold in TBO, TBO cycle is analysed separately for strong minus weak TBO composites, in which strong (weak) TBO is associated with strong (weak) IOD and strong (weak) TBO is associated with weak (strong) IOD years. The later case is associated with La Niña/El Niño years also. The results are presented in the form of comparison of these two cases for individual season from the TBO year summer to next year summer for SST, wind and 200 hPa vertical velocity.

During the monsoon season when TBO and IOD are strong, monsoon cooling confines to ocean region close to India and Indonesia and other regions are warm along with east Pacific. Upward motion is over east Pacific and west Indian Ocean regions. Westerlies in the north Indian Ocean extends to equatorial Pacific and easterlies are strong in the equatorial Indian Ocean (see left panels in figure 4.9). During the summer monsoon season when strong TBO is associated with weak IOD, entire Indian Ocean SST is cool and convergence is over India and Indian Ocean regions. Westerlies are strong in the north Indian Ocean and easterlies are present in the south Indian Ocean and maximum easterlies are over the equatorial Pacific (right panel of figure 4.9).

After the strong monsoon, cooling in north Indian Ocean decreases and convergence in the western Indian Ocean moves to southwest Indian Ocean when both TBO and IOD are in same phase as seen in figure 4.10. Easterlies are strong in the equatorial east Indian Ocean and westerly maximum moves to east Pacific. Entire Indian Ocean and east and central Pacific regions are warm and convergence is over Indonesia in the TBO years with IOD and ENSO. Westerly maximum moves southeast and easterlies in the central and east Pacific.
Figure 4.9: JJA0 strong-weak anomalies for TBO cycle in presence of IOD. Left side of figure for in phase IOD and TBO (a) SST, (c) 200 hPa velocity potential (e) 850 hPa wind and right panel for out of phase relationship of same parameters.

Figure 4.10: SON0 strong-weak anomalies for TBO cycle in presence of IOD. Left side of figure for in phase IOD and TBO (a) SST, (c) 200 hPa velocity potential (e) 850 hPa wind and right panel for out of phase relationship of same parameters.
The in-phase and out-of-phase association of TBO and IOD for the boreal winter season DJF0 is shown in figure 4.11. During the boreal winter following the strong monsoon (DJF0), Indian and Pacific Oceans are warm and convergence is over southwest Indian Ocean in the inphase transition of both TBO and IOD. Easterlies remains in the southeast Indian Ocean and north of Australia and westerlies are in the extreme east Pacific (figure 4.11, left panels). Entire Indian Ocean is cool along with equatorial Pacific and convergence is over Australia and west Pacific in the boreal winter after strong monsoon when TBO and IOD are in opposite phase (figure 4.11b). Westerly anomalies reach north of Australia and easterlies are in the eastern Indian Ocean.

Figure 4.11: DJF0 strong-weak anomalies for TBO cycle in presence of IOD. Left side of figure for in phase IOD and TBO (a) SST, (c) 200 hPa velocity potential (e) 850 hPa wind and right panel for out of phase relationship of same parameters

In the next year spring season (MAM1), when TBO and IOD are in same phase, Indian Ocean is warm along with equatorial Pacific and convergence is at extreme southeast Pacific. Easterlies are seen over equatorial Indian Ocean and west Pacific (see figure 4.12, left panels). In opposite phase TBO and IOD, cooling continue in both the Indian and Pacific Oceans and convergence moves to north central Pacific. Westerlies also
move to west Pacific shifting easterlies further east and easterlies are seen over Indian subcontinent (figure 4.12, right panels).

Figure 4.12: MAM1 strong-weak anomalies for TBO cycle in presence of IOD. Left side of figure for in phase IOD and TBO a) SST, c) 200 hPa velocity potential e) 850 hPa wind and right panel for out of phase relationship of same parameters

Figure 4.13: JJA1 strong-weak anomalies for TBO cycle in presence of IOD. Left side of figure for in phase IOD and TBO (a) SST, (c) 200 hPa velocity potential (e) 850 hPa wind and right panel for out of phase relationship of same parameters
In the next year monsoon of in phase TBO and IOD, Indian Ocean is warm along with west Pacific and east Pacific and convergence is over equatorial east Indian Ocean (figure 4.13 a and c). Easterlies are seen over India and in the west Pacific Ocean. In out of phase JJA1 season, Indian Ocean is still cool and east Pacific is warm and convergence is over central Pacific in presence of both IOD and ENSO. Westerlies are seen in the equatorial Indian Ocean and west Pacific.

Thus the cyclic evolution and movement of TBO scale anomalies are different according to the phase of IOD. When IOD and TBO is in same phase ENSO is also absent and in reverse association, ENSO is absent. Pacific Ocean SST and Walker circulation anomalies are opposite in both the cases, while lower level wind and SST in the north Indian Ocean has slight changes only.

4.4.3 TBO cycle in the absence of both ENSO and IOD (pure TBO years)

In this section the possibility of TBO cycle in the absence of both these Indian Ocean and Pacific Ocean phenomena are investigated for SST, velocity potential and wind for JJA0 to JJA1 with the same method used for above analysis and is presented in figure 4.14 and 4.15.

Along with the onset of strong monsoon Indian Ocean cools close to land and southeast region. Westerlies strengthen over western Indian Ocean and India and easterlies are in the equatorial Indian Ocean. Convergent anomalies are seen over India with center in the northwest Pacific and divergence is over southeast Indian Ocean and northeast Pacific. Central and southwestern Indian Ocean is still warm in the SON0 and warm center in the east Pacific moves to central region. Westerly maximum moves to northwest Pacific and southeast Indian Ocean and convergence is in the northwest Indian Ocean. Entire Indian Ocean is not cooled in the preceding winter. Westerly maximum is over the Australian region, though weak westerlies are seen in western Indian Ocean. But downward motion is seen over the north of Australia. East Pacific also cool in the MAM1 season along with the north Indian Ocean. South Indian Ocean has westerly anomalies and north equatorial region has easterly anomalies and
equatorial Indian Ocean is convergent. North Indian Ocean is cool in the next monsoon season and easterlies are strong in the north Indian Ocean and upward motion center is over southeast Pacific, though Indian region is convergent.

Figure 4.14: strong minus weak composite of SST (left panel) and 200 hPa velocity potential (right panel) for pure TBO (both ENSO and IOD are absent) years

Figure 4.15: strong minus weak composite of 850 hPa wind over Indian Ocean region for pure TBO years

4.5 Discussion and summary

The present chapter identifies the difference in evolution of anomalies associated with TBO in presence of ENSO and IOD. ENSO years with spring onset only can affect the
interannual variability of Indian summer monsoon associated with TBO. The theory of TBO by Meehl and Arblaster (2002a) stresses the spring season processes give rise to monsoon transition in the next summer. The condition in Pacific during spring season affects TBO than the maximum anomaly of ENSO. Monsoon cooling will not distribute to the entire north Indian Ocean in absence of ENSO and its southeast movement to entire Indian Ocean by next winter is controlled by ENSO. Pacific Ocean SST anomalies are opposite in ENSO-TBO and normal TBO years and warms by weak TBO in absence of ENSO (Pillai and Mohankumar 2007). Southeast movement of westerly anomaly to Australian monsoon region from north Indian Ocean is controlled by ENSO. Biennial Oscillation of SLP over India and Indian Ocean in the interannual variability of monsoon is possible only with the presence of ENSO. Convergence and upward motion strengthens over equatorial Pacific close to dateline in summer and post monsoon season in the absence of ENSO and divergence appears over Australia in the winter season. Equatorial Walker circulation has downward (upward) motion in the equatorial Indian Ocean in strong (weak) monsoon season in the absence of ENSO. Local Hadley circulation between the Indian monsoon region and adjacent ocean is present in absence of ENSO only for strong/weak monsoon. Eastward movement of equatorial convection anomalies is limited to Indian Ocean region only in the absence of ENSO as illustrated in Pillai and Mohankumar 2007.

Thus local SST in the Indian Ocean close to Indian monsoon region, monsoon convection, and local Hadley circulation has biennial cycle in the absence of ENSO and forces interannual variability of Indian summer monsoon. At the same time large-scale features affecting Indian monsoon like Pacific Ocean SST anomalies, SLP, Walker circulation etc is modulated by ENSO making its biennial transition and in phase Indian to Australian monsoon transition. By Meehl et al (2003), Indian to Australian monsoon linkage is through large scale east-west circulation with eastern Walker cell links Indian and Australian monsoon to dynamics in the Pacific and western Walker cell connects Indian and Australian monsoon to dynamical processes in the Indian Ocean. Thus the convection movement from Asian to Australian monsoon connects Indo-Pacific Oceans. Similarly the persistence of anomalies in the Indian Ocean is not observed in the
absence of ENSO. Our present observation of TBO cycle for Indian Ocean SST and local processes supports Wu and Kirtman (2007) who through the modeling study produced TBO after suppression of Pacific Ocean. Thus the biennial cycle observed for Indian Ocean parameters can produce interannual variability of Indian summer monsoon in absence of ENSO. But many properties of TBO like in-phase Indian to Australian monsoon transition is not observed.

The interaction of IOD with TBO is twofold for interannual variability of Indian summer monsoon. In presence of El Nino (La Nina) in the Pacific, negative (positive) TBO is associated with positive (negative) IOD and in absence of El Nino/La Nina TBO and IOD are in same phase. The in-phase TBO and IOD SST anomalies look like non ENSO TBO composites discussed earlier in this chapter. The SST, convergence and wind anomalies of out of phase TBO-IOD composite looks like ENSO-TBO cases, indicating that in both these cases ENSO is the dominant factor. Walker circulation and Pacific Ocean SST anomalies are opposite in TBO-IOD in phase and out of phase association.

According to Ashok et al (2004), when a strong IOD is associated with El Nino, its influence on monsoon is reduced by both poles of IOD. An anomalous divergence is induced in east Indian Ocean and divergence over west Pacific and convergence over India by positive IOD. But in the case of monsoon variability governed by TBO, associated with positive (negative) IOD and El Nina (La Nina), SST pattern looks similar to TBO-ENSO composite, which include both IOD-ENSO and non-IOD ENSO years. Convergence/divergence pattern is also similar in JJA0 and SON0 of La Nina (El Nina) is associated with negative (positive) IOD in strong (weak) TBO years. But in DJF0, upper level convergence of equatorial Indian Ocean is strengthened in IOD only ENSO-TBO years and the upper level divergence in the west Pacific shifts northward in next two seasons.

The SST composite of IOD only TBO and non-ENSO TBO cases is comparable, though the anomaly is strong in the IOD only TBO cases. But the convergence/divergence pattern is different in both the cases. In IOD only TBO years convergence is in the east Pacific from JJA0 to MAM1 and is in the equatorial east Indian Ocean by JJA1. In IOD
only TBO case, only one Walker circulation cell with one convergent center and one divergent center is present. But in the case of non-ENSO TBO, the central Pacific always remains convergent and more than one pair of convergent/divergent centers is obtained. Thus in the case of IOD only TBO years, the convergence/divergence associated with IOD dominates the TBO and in when associated with ENSO, it dominates over IOD.

In the absence of both these dynamic processes in both the Indian and Pacific Ocean, SST and wind pattern has biennial cycle (pure TBO years). SST pattern of JJA0 to DJF0 resembles the TBO pattern in the absence of ENSO. But when IOD is present the warming of the Indian Ocean in the next phase starts in the MAM1 itself. But in pure dipole years Indian Ocean is cool in the MAM1 season and starts warm by JJA1 only. But convergent/divergent centers have no biennial cycle in absence of ENSO and IOD.