The Asian Monsoon, the Tropospheric Biennial Oscillation and the Indian Ocean Zonal Mode in the NCAR CSM

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Abstract

The interaction of the Indian Ocean dynamics and the tropospheric biennial oscillation (TBO) is analyzed in the 300-year control run of the National Center for Atmospheric Research (NCAR) Climate System Model (CSM). Sea-surface temperature (SST) anomalies and equatorial ocean dynamics in the Indian Ocean are associated with the TBO and interannual variability of Asian-Australian monsoon in observations. We analyze the air-sea interactions involved in these processes in the coupled ocean-atmosphere model, so as to diagnose the causes of the SST anomalies and their role in the development of a biennial cycle in the Indian-Pacific Ocean region.

We first diagnose the ability of the model to reproduce the observed characteristics of the TBO by analyzing the local and remote forcings as well as the boreal springtime transition conditions associated with the TBO. By using SVD analysis, it is found that the model reproduces the dominant mechanisms which are involved in the development of the TBO’s influence on the south Asian monsoon: Large-scale forcing from the tropical Pacific, and regional forcing associated with both the meridional temperature gradient between the Asian continent and the Indian Ocean, as well as Indian Ocean SST anomalies. Using cumulative anomaly pattern correlation, the strength of each of these processes in affecting the interannual variability of both Asian and Australian monsoon rainfall is assessed.

The role of ocean dynamics in the Indian Ocean in the influencing interannual variability of the Asian Summer monsoon and the TBO is also explored. Specifically, we analyze the role that the Indian Ocean Zonal Mode (IOZM) of SST anomalies, known as the “Dipole”, as well as meridional oceanic heat transport in the interannual variability of monsoon and the TBO. It is found that the IOZM is an inherent feature of the Asian summer monsoon, and is also correlated to ENSO-related SST anomalies in the Pacific Ocean, and is thus a part of the biennial nature of the Indian-Pacific Ocean region. The coupled ocean-atmosphere dynamics and cross-equatorial heat transport contribute to the interannual variability and biennial nature of the ENSO-monsoon system, by affecting the heat content of the Indian Ocean and resulting SST anomalies over multiple seasons, which is a key factor in the TBO.
1. Introduction

The interannual variability of the Asian-Australian monsoon system has been a focus of substantial research in the recent decades because of the influence of the monsoon on global climate, its interaction with the El Niño-Southern Oscillation (ENSO) phenomenon, and the magnitude of its influence on the livelihood of more than 50% of the global population. Recently, the tropospheric biennial oscillation (TBO) has been shown to contribute to climate variations in the Indian and Pacific Ocean regions on interannual timescales. The relationship of ocean dynamics in the tropical Indian ocean region to these large-scale climate variations in the ENSO-monsoon system associated with the TBO in the National Center for Atmospheric Research (NCAR) coupled Climate System Model (CSM) will be the primary focus of this paper.

The TBO has been defined as the tendency for the Asian monsoon to alternate between “strong” and “weak” years. Thus the TBO encompasses ENSO years which also have a similar biennial relationship to the Asian-Australian monsoon. The transitions between the years in this cycle are characterized by large-scale coupled land-ocean-atmosphere interactions in the Indian and Pacific Ocean regions (Meehl 1997). Both observational (Yasunari 1990; Yasunari 1991; Ropelewski et al. 1992; Yasunari and Seki 1992; Yang et al. 1996; Tomita and Yasunari 1996; Meehl 1987; Meehl 1997; Webster et al. 1998) and modeling studies (Ogasawara et al. 1999; Kitoh et al. 1999; Chang and Li 2000; Li et al. 2001) have attempted to describe the basic processes involved in the biennial nature of the tropical climate in this region, and various theories have been given to assess the causes of the TBO (Meehl 1987; Meehl 1993; Meehl 1994; Goswami 1995; Webster et al. 1998; Kawamura et al. 1999; Clarke et al. 1998; Kim and Lau 2001). Recent studies have attempted to quantify the strength of the contributions
of various local and remote forcings on the TBO (Meehl and Arblaster 2001b), as well isolating
the coupled features involved in the transition of the TBO from a year of one sign anomalies to the
other (Meehl and Arblaster 2001c). It has also been suggested that decadal variations exist in the
strength of the ENSO-monsoon relationship (Torrence and Webster 1998), and also that the recent
breakdown in the ENSO-monsoon correlation for the Indian-Pacific Ocean region may be due to
global-scale warming due to anthropogenic factors (Krishna Kumar et al. 1999).

In this paper we will focus on how the ocean dynamics in the tropical Indian Ocean contribute
to the development of sea-surface temperature (SST) anomalies on interannual timescales, and thus
contribute to the biennial variability in the Indian-Pacific Ocean climate. Because of the heat stor-
age and consequent memory of the ocean, various oceanic dynamical processes associated with
the TBO help develop and maintain SST anomalies over multiple seasons, a key aspect of the bi-
ennial cycle. Loschnigg and Webster (2000), Loschnigg and Webster (2002) and Webster et al.
(2002) have shown that the variability of the Indian Ocean is in the form of distinct and slowly
varying dynamic modes and not just localized mixing as originally argued by Meehl (1987). The
presence of Indian Ocean SST anomalies has been central to many of the theories of the TBO
(e.g. Meehl 1997; Chang and Li 2000), and the development of an equatorial SST gradient in
the boreal fall season, along with the existence of equatorial wave dynamics that contribute to
thermocline depth anomalies and enhance SST anomalies, has been associated with the strong and
weak monsoon cycle and ENSO variability in descriptions of the TBO cycle (Meehl et al. 2001;
Webster et al. 2002). The existence of a strong Indian Ocean equatorial SST zonal mode (IOZM)
has been given much attention after the particularly strong event in 1997-98, concurrent with the
strong El Niño event of that year (Webster et al. 1999; Saji et al. 1999). It has been sug-
gested that this mode of SST anomalies, which has also been termed the “dipole mode”, is an
independent mode of oscillation and not associated with ENSO (Iizuka et al. 2000). But recent studies suggest that the mode of SST variability is more of a southern Indian Ocean phenomena (Behera et al. 2000), that there is no such independent mode (Baquero-Bernal and Latif 2001; Nicholls 2001) or that the IOZM is an integral part of the TBO (Meehl and Arblaster 2001b). Even in the studies that question the existence of a independent mode, the results suggest an equatorial SST anomaly gradient in the Indian Ocean during the boreal fall season, which will be discussed below as being connected to the ENSO-monsoon system. In observational studies (Meehl and Arblaster 2001b), ocean dynamics in the Indian Ocean are associated with TBO springtime transition as well as SST anomalies occurring in the equatorial regions. Thus ocean dynamics are likely associated with extremes of Indian Ocean dipole events that are a natural part of the TBO evolution.

In this study, we will be analyzing the influence of the ocean dynamics in a coupled model in this region, and its relation Indian Ocean SST dipole, the Asian monsoon, ENSO and the TBO.

The large scale meridional oceanic heat transport in the Indian Ocean has been studied on both seasonal (Hsiung et al. 1989; Hastenrath and Greischar 1993; Wacongne and Pacanowski 1996; Loschnigg and Webster 2000) and interannual (Garternicht and Schott 1997; Lee and Marotzke 1997; Loschnigg and Webster 2002) timescales. It has been shown that the heat transport modulates the northern Indian Ocean SST on seasonal timescales (Loschnigg and Webster 2000), and also may be involved in the interannual variability of the Asian monsoon by affecting heat content and SST anomalies as part of a biennial cycle of SST modulation (Loschnigg and Webster 2002). We will examine the cross-equatorial heat transport in a coupled model in this study to determine the relationship on interannual timescales to the monsoon, the IOZM and the TBO.

Webster et al. (2002) attempted to understand the Indian Ocean and monsoon phenomena de-
scribed above, and the influence of ENSO, in terms of a single phenomena. They hypothesized that the IOZM and the TBO were integral parts of a self-regulation process of the monsoon which acted as negative feedbacks to keep the variability of the monsoon, both on annual and interannual time scales, within rather narrow limits. Their hypothesis resulted from the observation that the interannual variability of South Asian monsoon rainfall had a very small amplitude and that prolonged periods of drought or flood (multi-year) were extremely rare. The elements of the theory can be described as follows:

(i) A strong (weak) monsoon year is accompanied by stronger southward (northward) meridional oceanic heat fluxes that tend to decrease (increase) the cross-equatorial SST gradient. These variations in meridional heat flux result from enhancements (decreases) in cross-equatorial Ekman transports of heat induced by the anomalous wind (Loschnigg and Webster 2000; Loschnigg and Webster 2002). It is noted that the anomalous ocean fluxes of heat are in the opposite sense to the anomalous divergent wind component and the anomalous meridional atmospheric heat flux (Webster et al. 2002). These anomalous oceanic heat transports are central to the biennial nature of Indian Ocean SSTs and the monsoon by producing for the following year a cooler or warmer North Indian Ocean regime, as originally suggested by Meehl (1994) and Meehl (1997).

(ii) The anomalous low tropospheric monsoon wind induces anomalous upwelling in the eastern and the western parts of the ocean basin. A strong monsoon increases upwelling along the Somalia coast and Arabian Sea but reduces the upwelling along the coast of Sumatra. A weak monsoon, on the other hand, reduces the upwelling along the Somalia coast but increases it off Sumatra. Thus, a strong monsoon is associated with west to east anomalous
SST gradient, while a weak monsoon is associated with a reversed SST gradient. Webster (2002) notes that the creation of the anomalous SST gradient occurs in early summer at the time of minimal east-west SST gradient along the equator. It is hypothesized that the slack climatological SST gradient allows the positive feedback between the atmosphere and ocean and the growth of the IOZM, as suggested by Webster et al. (1999). Thus, the positive phase of the IOZM (warm western Indian Ocean) follows a weak monsoon while the negative phase follows a strong monsoon. It should be noted that these observations are in keeping with Saji et al. (1999) who noted that the IOZM was phase locked to the annual cycle. Webster et al. (2002) suggest that the anomalous heat flux associated with the IOZM emphasizes the biennial heat transports noted in (i) above.

(iii) The combined effects of the biennial Ekman meridional and zonal IOZM heat transports was to repress large scale variations in monsoon magnitude. Furthermore, they suggested that these phenomena, in combination, would reduce the impact of external forcing of the monsoon such as by ENSO. In fact, these regulation feedback processes occur relative to any major perturbation of the monsoon which is why the development of the IOZM occurs sometimes in the absence of an extreme in ENSO. In this manner, anomalous meridional oceanic heat transports and the IOZM are natural responses of the perturbed monsoon system to external forcing and act to reduce the impact of such forcing.

On the the problems with the theory outlined above, is that it is the result of a combination of interpretation of reanalysis results and other observations and experiments with stand-alone ocean models. A major purpose of this study is to examine the hypothesis and test its consistency from the context of a coupled ocean-atmosphere model.
In section 2 we will discuss the coupled model and the observational datasets used for comparison. Section 3 will assess the ability of the coupled model to simulate the biennial cycle in the Indian-Pacific Ocean region, and will discuss the role of the Indian Ocean SST dipole and the oceanic meridional heat transports in the Asian-monsoon ENSO system. Discussion and conclusions will be given in section 4.

2. Model description and comparisons with observational climatology

The model used is the NCAR CSM global coupled ocean-atmosphere-land surface-sea ice model, described in Boville and Gent (1998). The atmosphere used is the NCAR CCM3 with T42 resolution and 18 vertical levels (Kiehl et al. 1998). The ocean component is a global GCM with a 2.4° by 2.4° resolution which is reduced to 1.2° latitude in the equatorial tropics (Gent et al. 1998). The model also includes active land surface (Bonan 1998) and sea ice (Weatherly et al. 1998) components. The coupled model is run without flux adjustments. For this study we analyze monthly mean fields of the first 100 years of the 300 year control run. The oceanic fields are available only from year 27 onward, as the implementation of the model code for the calculation of the monthly mean field was not available before that period of the model control run.

Meehl and Arblaster (1998) discuss the ability of the CSM to reproduce characteristics of ENSO and the Asian-Australian Monsoon. They find that the CSM is able to reproduce the major features of the monsoon system, including the mean climatology, interannual variability and the connections of the Asian-Australian monsoon to the tropical Pacific. The Nino3 SST anomaly index in the CSM produces about 60% of the observed variability, within the range simulated by other coupled models. As in observations, indices of Asian-Australian monsoon strength are negatively correlated with Nino3 SST, and both indices show amplitude peaks in the ENSO and
TBO frequencies (3-6 yr and 2.3 yr, respectively). Meehl and Arblaster (1998) also note that in correlation maps of the Nino3 SST index with sea-level pressure (SLP), the zero-line separating positive and negative correlations with SLP (their Fig. 10) is shifted further west in the CSM as compared to observations, causing a weaker association of central and eastern Pacific SST anomalies with the Australian monsoon, and a stronger association with the Indian monsoon. Meehl and Arblaster (1998) also note that the model has a tendency to produce multi-year ENSO events. Although these type of protracted events have been noted in observations (Trenberth and Hoar 1996; Allan and D’Arrigo 1999; Reason et al. 2000), their existence in the model solutions influences the biennial tendency of the results, and will be discussed further in Section 3.

Comparisons of model climatology for the annual cycle with the Xie and Arkin (1996) CMAP precipitation dataset and the NCEP/NCAR reanalysis dataset (Kalnay et al. 1996) for surface skin temperature ($T_s$) and 850mb winds are shown in Figs. 1 and 2, respectively. The precipitation in the CSM (Fig. 1) is higher than in the observation for the regions in the equatorial Indian Ocean and in the Pacific warm pool region near the Phillipines by up to 6 mm/day. The precipitation errors in the CSM are related to errors in the SST fields via differences in latent heat flux, evaporation and low-level winds, leading to differences in the low-level moisture convergence, especially in the Phillipine region in JJAS (Fig. 1c). Latent heat flux in the North Indian Ocean and western Pacific ITCZ regions is about 30-60 $Wm^{-2}$ too great, resulting in evaporation errors of between 1-2 mm/day too high. The comparison of the CSM $T_s$ and surface winds to observations (Fig. 2) shows surface temperatures in the CSM in the Indian Ocean and western Pacific warm pool regions about 1°C cooler than NCEP reanalysis values in many regions throughout the year, and a monsoon inflow in JJAS that is weaker than in observation (Fig. 2c), resulting in reduced summer monsoon rainfall in the Indian region, particularly in the Bay of Bengal region (Fig. 1c). There also exist
significant errors in precipitation in the CSM when compared to observations for the boreal fall and winter seasons. In Figs. 1e (model SON) and 1g (model DJF), the CSM shows precipitation values of a greater sign than observations in the western equatorial Indian Ocean (Figs. 1f and 1h). These errors in convection could affect the equatorial response of the atmosphere and thus the mechanisms associated with the IOZM. Aside from these differences as compared to observations, the model does a reasonable job in simulating the large-scale aspects of the annual cycle of the Asian-Pacific region, and is therefore sufficient for the purposes of this paper.

3. Results

a. Analysis of local and remote TBO forcings

In this section we analyze the ability of the CSM to simulate the various local and remote forcings during the boreal spring transition period of the TBO, as well as the ability of the CSM to transition from the boreal summer Asian monsoon to the boreal winter Australian monsoon. We first perform lagged singular value decomposition (SVD) analysis to define the maximum covariability of the spatial associations between various fields in the boreal spring season (MAM) and the resulting Indian summer (JJAS) monsoon precipitation. The analysis is similar to the observational study of Meehl and Arblaster (2001a), to which we will be comparing our results. We focus on three main local and remote processes that are known to influence the interannual variability of the Asian summer monsoon. First, SST anomalies in the Indian Ocean can affect the Indian summer monsoon moisture source and produce a more vigorous monsoon circulation. Second, SST anomalies in the central and eastern Pacific Ocean can affect the monsoon through changes in the large-scale east-west Walker circulation. And third, the 500 hPa height anomalies over Asia cause changes in the large-scale atmospheric circulation, which results in changes in Asian land temper-
atures that enhance the land-sea temperature gradient that drives the Indian summer monsoon. The association of each of these three forcings are shown in Fig. 3.

For positive SST anomalies in the central and eastern Indian Ocean in MAM (Fig. 3a) there are positive precipitation anomalies located mainly in the Bay of Bengal, but also over the central and northern Indian subcontinent in JJAS (Fig. 3b). Increased SSTs in these areas provides the Indian summer monsoon with an added moisture source and help influence the increased strength of the monsoon circulation. Compared to observational results (Meehl and Arblaster 2001a), the model produces a similar strong response of SST over the central Indian Ocean but fails to produce the precipitation response in JJAS over western India, producing increased values only over eastern India and the Bay of Bengal. This result is likely related to similar systematic errors in the CSM climatology as noted in section 2. Negative SST anomalies in the central Pacific Ocean in MAM (Fig. 3c) are associated with positive rainfall anomalies much of India and the Bay of Bengal in JJAS (Fig. 3d). The relationship between the precipitation anomalies over India and the SST anomalies in the Pacific is the result of changes in the large scale east-west circulation spanning the Pacific and Indian Oceans, which affect the monsoon through either enhancing or suppressing the vertical motion of the monsoon circulation over the Indian subcontinent. In observations, during the biennial cycle the sign of the SST anomaly in the central and eastern Pacific Ocean transitions from positive anomalies in MAM to negative anomalies in JJAS, leading to increased precipitation over the Indian region in JJAS. The model result shows persistent negative anomalies, starting in MAM and continuing through the boreal summer months (not shown). The continuous La Niña-like conditions still produce positive rainfall anomalies in JJAS. The persistent SST anomalies of one sign is an illustration of the tendency for the model to produce multi-year ENSO events. For positive 500 hPa height anomalies over central and south Asia in MAM (Fig. 3e), there are
small positive precipitation anomalies in eastern India and the Bay of Bengal. Meehl (1997) suggests that this pattern is a Rossby wave response to convective heating anomalies over the equatorial Indian Ocean and western Pacific in the form of an anomalous ridge in central Asia. The results here show a slightly weaker precipitation response over India and the Bay of Bengal as compared to observations, but otherwise a similar response over India and the monsoon to the Asian temperature anomalies.

It is possible to quantify the impact of each of these local and remote forcing mechanisms on the monsoon precipitation so as to assess the relative strength of each of the mechanisms in each year, and whether they are functioning independently from one another or whether they are acting in unison. We analyze spatial anomaly pattern correlations (Lau and Wu 1999; Meehl and Arblaster 2001a) between the model produced summer monsoon rainfall and the SVD projections produced individually and cumulatively. The cumulative pattern correlation for the kth SVD for i transition conditions to the model produced precipitation anomaly for the jth year is:

\[ A_{i,j} = \langle O_j, \sum_{k=1}^{i} S(k) \ast C(k) \rangle \]

where \( \langle \rangle \) is the spatial anomaly pattern correlation over the monsoon region, \( S \) is a normalized precipitation SVD pattern for JJAS associated with a MAM forcing mechanism, \( C \) is the precipitation SVD expansion coefficient for that year, and \( O \) is the model produced Indian rainfall anomaly. Each anomaly pattern correlation is first calculated separately (Fig. 4a), and then in a cumulative form (Fig. 4b). From Fig. 4a it is seen that in many years all three mechanisms are acting in unison to produce variations in Indian summer monsoon rainfall, and in some years all three mechanisms are failing to have much of an effect on the precipitation. For a majority of the years, the three conditions are very similar in their strength of correlations, and then fall out of synchronous im-
pacts during certain decades when all three are continuously weak (e.g. years 30-40 and 60-70). This is indicative of the decadal variability in the relationships, and is also apparent in observations (Torrence and Webster 1998). The cumulative pattern correlation shown in Fig. 4b also periods of especially strong relationships (e.g. years 10-15, 20-30 and 50-60). The mean cumulative anomaly pattern correlation for the 100 years of model output is 4.6. When we use the reconstructed SVD patterns to calculate the Indian summer monsoon rainfall index, that index correlates to the actual precipitation index at 0.8, which is slightly lower than that of the observational study (0.9), but high enough to illustrate the model's ability to reproduce the correct forcings for this region.

Between the boreal summer and winter monsoons in the Asian-Australian monsoon region, there tends to be a persistence between the transition from a strong Indian monsoon to a strong Australian monsoon, with a convective maximum progressing over south-east Asia in SON towards north Australia in DJF (Webster et al. 1998; Meehl and Arblaster 2001b). For this reason we undertake a similar analysis for the boreal winter Australian monsoon, using the SON conditions as the forcing mechanisms for the DJF precipitation over northern Australia. Three processes thought to be associated with the Australian monsoon are: first, SST over the equatorial and southern Indian Ocean; second SST anomalies over the central and eastern Pacific Ocean; and third, rainfall over southeast Asia, which is a signal of the transition of the area of maximum convection from south Asia during JJAS to northern Australia in DJF.

Results for the SVD analysis of SON forcings on DJF rainfall over northern Australia are shown in Fig. 5. The general pattern of SST anomalies is a dipole pattern of eastern equatorial and southeast Indian Ocean SST, and cooler SST in the western equatorial and Arabian Sea regions in SON (Fig. 5a) associated with stronger precipitation over northern Australia and the maritime continent in DJF (Fig. 5b). The persistence of the dipole pattern of SST anomalies of the same
sign from MAM to SON is another example of the tendency of the model to produce multi-year ENSO events, which also have an influence on the Asian monsoon by producing prolonged years of strong or weak monsoons, and thus producing some-sign SST anomalies in a dipole pattern over subsequent years. In the Pacific Ocean, cooler central SSTs in the central equatorial basin (Fig. 5c) are associated with slightly higher precipitation for similar regions over northern Australia and the maritime continent in DJF (Fig. 5d), but of slightly lower strength. Finally, the transition of the convective maximum in SON over southeast Asia and the maritime continent (Fig. 5e) are associated with increased precipitation over the regions near northern Australia (Fig. 5f). Except for some slight differences in the strength of the correlations, these patterns are very similar to those of observations (Meehl and Arblaster 2001b). The pattern correlations for DJF rainfall are shown in Fig. 6. The striking feature is how all three of the individual pattern correlations follow each other very closely (Fig. 6a), showing how the different forcing mechanisms are closely connected to each other in their ability to determine boreal winter Australian monsoon precipitation. As in the observations the SON transition conditions are closely related to those in MAM. This is indicative of the seasonal persistence from the Indian to the Australian monsoon in observations and model. The mean cumulative anomaly pattern correlation for the Australian monsoon transition conditions in Fig. 6 is 0.43.

From the above analysis we can conclude that the coupled model sufficiently reproduces the observed local and remote forcing mechanisms that affect anomalous rainfall in both the summer and winter phases of the Asian-Australian monsoon, as well as the transition from a strong summer to winter monsoon. For this reason, we will use the model to diagnose the relationship of the coupled ocean dynamics in the Indian Ocean with the large-scale anomalies of the Asian-Australian monsoon, how they play a role in the TBO and the Indian-Pacific Ocean climate.
b. Indices relating to Monsoon strength

In this section we will analyze two indices relating to Asian summer monsoon strength that will be used in following sections to assess ocean dynamical features as they relate to the interannual variability of the Asian monsoon. The first is a boreal summer (JJAS) rainfall index which is similar to the All-India rainfall index of Parthasarathy et al. (1991), except that it includes rainfall over oceanic regions as well as land, which is important for determination of large-scale strength of the monsoon. The boundaries for the rainfall index are shown in Fig. 1c (10° to 25°N and 70° to 100°E). This index area is similar to the boundaries of the convective index used by Wang and Fan (1999), and we will use the similar term CI1 to identify it.

Another method of determining the strength of the Asian summer monsoon is by the westerly wind shear (Webster and Yang 1992). Figure 7a shows the climatological U850-U200 wind shear for JJAS, which is used for the calculation of the Westerly Shear Index (WSI1). Figure 7b shows the correlation of U850-U200 wind shear for JJAS with the CI1 for 100 years of the model. The region used for the calculation of the WSI1 (5° to 15°N and 35° to 75°E) is in the region of the highest positive correlation with the CI1 (+0.68) ————-. These results are very similar to the observational analysis of Wang and Fan (1999) (see their Fig. 4a), with strong positive values over the western Indian Ocean and eastern Africa between 20°N and 20°S. The bounds used for the calculation of the WSI here are very similar to those used in (Wang and Fan 1999). Also interesting to note in Fig. 7b is the large region of negative over the western Pacific ocean, also similar to observations, which is an illustration of the Walker circulation strength associated with ENSO-monsoon relationships.

The timeseries for the WSI1 and the CI1 are shown in Fig. 8a for the first 100 years of the
model control run. The dots representing the “strong” and “weak” TBO years will be discussed in the next section. Power spectra of the monthly values for the CI1 and WSI1 are shown in Figs. 8b and 8c, respectively. Both spectra show small peaks in the 2-3 year range, signifying a signal in the TBO timescale. The WSI has some peaks in the 3-5 year range, signifying some relationships to more longer scale variability typically seen of ENSO-like Pacific Ocean variability. Although both spectra also show significant power in the decadal range, variability on this time scale will not be discussed in this paper.

c. Indian Ocean dynamics and the TBO

To analyze the relationship of Indian Ocean dynamics to the TBO and to the interannual variability of the Monsoon, we construct composites from the model based on years where both of the indices described in the previous section (CI1 and WSI1) are relatively greater or less than the previous or following years. That is, if $P_i$ is the value of the index (both CI1 and WSI1 in this case) for a given year $i$, then a relatively strong year is defined as

$$P_{i-1} < P_i > P_{i+1}$$

and a relatively weak monsoon is defined as

$$P_{i-1} > P_i < P_{i+1}$$

Such “strong” and “weak” years are defined as “biennial” years (Meehl and Arblaster 2001b). These years are shown in Fig. 8a as the solid red and blue dots, respectively. All composites shown in this section will be for “strong” minus “weak” years.

Figure 9 shows the composites for $T_s$ and 850mb winds for six seasons from the model during part of a biennial cycle (MAM(0), JJA(0), SON(0), DJF(+1), MAM(+1) and JJA(+1); JJA(0) is
the season for which the composites are based). The strong monsoon circulation is seen in JJA(0) (Fig. 9b), with strong southeasterlies in the southwestern Indian Ocean and strong along-shore winds near Somalia. There are also northwesterlies near Sumatra, which are an indication of reduced along-shore and offshore winds there. Surface temperatures over the Indian subcontinent are anomalously cool in JJA(0) and SON(0) (Fig. 9c) from increased monsoon precipitation and wetter land surfaces. During the period near JJA(0), the central and eastern Pacific Ocean transitions to cool surface temperatures. These cool anomalies, along with the strong easterly winds near the western equatorial regions in the Pacific Ocean area in JJA(0) and SON(0) have the sign of La Niña conditions. This pattern of SST anomalies in the Pacific Ocean is consistent with the SVD analysis in Fig. 5c. The development of a negative (cool west, warm east) dipole of surface temperature in the equatorial Indian Ocean region begins in SON(0), which is consistent with the SVD analysis in Fig. 5a. The cool western Indian Ocean has contributions from anomalously strong upwelling from the strong along-shore monsoon winds. The warm anomalies in the eastern equatorial regions are partly a consequence of the reduced upwelling there due to the weaker along-shore and offshore winds near Sumatra. The surface winds are westerly along the equator in the Indian Ocean during SON(0) and DJF(+1) (Fig. 9d). This situation can be thought of as a “negative” SST dipole, and is the reversed sign of the “positive” SST dipole event of 1997-98 (Webster et al. 1999), which had an anomalously warm western equatorial Indian Ocean and cool eastern equatorial regions. The cool temperatures persist into DJF(+1) and MAM(+1) in the western Indian Ocean (Figs. 9d and 9e), which is a remnant of the previous strong monsoon. In the Pacific Ocean, there is the development of warm surface temperature anomalies in the central and eastern part of the equatorial regions in JJA(+1) (Fig. 9f), such as that seen in El Niño events. There are also signs of a weaker monsoon circulation in JJA(+1), with with anomalous northeasterly winds near Somalia, and the
beginning of slightly increased along-shore winds off Sumatra (Fig. 9f). During this time, slight westerlies are seen in the western and central equatorial Pacific Ocean. These transitions of surface temperature anomalies and wind circulation features show that there is a biennial component to the large-scale climate in this region. It is interesting to note the similarity of Fig. 9c to Figs. 3a and 5a.

Composites of precipitation anomalies from the model for the seasons starting with JJA(0) and going to SON(+1) are shown in Fig. 10. Strong positive anomalies of rainfall exist over Indian subcontinent, Bay of Bengal and the eastern equatorial Indian Ocean in JJA(0) (Fig. 10a), along with negative anomalies in the central Pacific Ocean coinciding with the cool SST anomalies there. Negative anomalies develop in the western Indian Ocean north of the equator in SON(0) (Fig. 10b), and become stronger in DJF(+1) (Fig. 10c), with opposite sign anomalies of precipitation developing in the central and southeastern equatorial region. This pattern of precipitation is consistent with the patterns of surface temperature and wind anomalies (i.e. an increased Walker-type circulation over the equatorial Indian Ocean, with increased surface westerlies). The strong positive precipitation anomalies transition through into the regions over the maritime continent and northern Australia during SON(0) and DJF(+1), consistent with the SVD analysis in Figs. 5e and 5f. The following boreal summer monsoon over India in JJA(+1) (Fig. 10e) shows negative precipitation anomalies, and the Pacific Ocean displays slight positive precipitation anomalies in the equatorial region in the following season (SON(+1), Fig. 10f), both opposite in sign to anomalies the previous year, showing the biennial nature to the Indian-Pacific region.

Composites for oceanic temperature as a function of longitude and depth along the equator in the Indian Ocean are shown in Fig. 11. During the strong monsoon season of JJA(0) (Fig. 11b), warm sub-surface temperature anomalies develop in the eastern equatorial areas near the coast of
Sumatra, in association with the anomalously weak along- and off-shore wind anomalies. The beginning of cool surface anomalies in the western equatorial region described in Fig. 9b can also be seen here. In the following season (SON(0), Fig. 11c), the warm sub-surface anomalies in the east reach peak levels of near 1°C between 50 and 100m of depth, while the cool western anomalies reach levels near 0.5°C. This pattern is comparable to observations (Meehl et al. 2001). In DJF(+1) (Fig. 11d), the anomalous sub-surface temperature gradient persists, with slight reductions in the magnitudes of the warm eastern regions, but cool western anomalies persisting in magnitude and reaching depths of 100m or more. In observations, the cool anomalies begin to cover the entire equatorial basin in DJF (+1) (Meehl et al. 2001). One possible reason that the model fails to reproduce this feature is that this surface transition in the eastern Indian Ocean may be due to surface heat flux anomalies in DJF, related to wind and enhanced latent heat fluxes, causing the surface temperatures to become negative even though there are positive temperature anomalies just below the surface in the observations (Meehl et al. 2001). This could be why the model does not “shut off” the dipole signature in the surface temperatures in the boreal winter season. In SON(0) (Fig. 11c), the temperature gradient at the sub-surface may also be thought of as a large-scale tilting of the thermocline, with a shallow thermocline in the west and a deeper thermocline in the east. During the following boreal summer season (JJA(+1), Fig. 11f), large negative anomalies are seen at sub-surface levels, displaying the biennial character of this process.

Composites of oceanic vertical velocity in the model at a depth of 50m for the Indian Ocean are shown in Fig. 12. In JJA(0) (Fig. 12a), areas of strong negative vertical velocity anomalies (negative values denote downwelling) are seen in the eastern equatorial Indian Ocean, and persist into the following season (SON(0), Fig. 12b). These downwelling anomalies area associated with the wind anomalies discussed for Figs. 9b and 9c). Anomalies of positive vertical velocity (up-
welling, and thus cooling) exist near the western equatorial region in JJA(0) and in the western off-equatorial regions in SON(0). During the following boreal summer monsoon season (JJA(+1), Fig. 12e), anomalies of positive vertical velocity exist in the eastern equatorial regions, further illustrating the biennial nature of the ocean-atmosphere system in this region. These vertical velocity results from the model confirm that wind forced upwelling anomalies hypothesized earlier play a role the SST anomalies in the TBO.

To illustrate the relationship of the Indian Ocean equatorial surface temperature gradient in the boreal fall season to the Indian summer monsoon, ENSO, and thus to the TBO, we produce an index to measure the IOZM, which is the temperature difference between the western and eastern equatorial areas in the model for SON, and compare it to the WSI1 as well as the Nino 3.4 surface temperature anomaly. The index, similar to the “Dipole Mode Index” (DMI) of Saji et al. (1999), is calculated as the temperature anomaly difference in SON for the western equatorial Indian Ocean region (5°S to 10°N and 40°E to 55°E) minus the temperature anomaly in the eastern equatorial region (10°S to 5°N, 90°E to 102°E). Figure 13a shows the WSI1 and the DMI for 100 years of model output. Table 1 list the correlations of the DMI to WSI1 and to various other indices to be discussed later. The WSI1 and DMI have a fairly strong negative correlation (-0.66), showing that the cooling in the western equatorial region along with the warming in the east during the boreal fall season is related to the anomalously strong monsoon. In years when the monsoon is anomalously weak, the reverse is true, and the western regions are cool while the eastern regions are warm. The WSI1 also correlates strongly to the SON SST anomaly in just the eastern region off the coast of Sumatra (-0.59), illustrating how that region area is a key area in these processes (Annamalai et al. 2001). Figure 13b shows the WSI and the index for the annual mean anomalous surface temperature in the Nino3.4 region. The two indices are negatively correlated (-0.83), showing
a strong relationship between the Indian summer monsoon and the tropical Pacific in the CSM. Strong summer monsoon years are associated with negative SST anomalies (La Niña in extreme years) in the Pacific, and vice versa for weak monsoon years. Note that the incidences of protracted multi-year El Niño or La Niña events also coincide with multi-year anomalies of the same sign of the WSI1 (e.g. years 42-44, 49-51). Contrary to other modeling studies (Iizuka et al. 2000), the results from the CSM also suggest that the IOZM is also strongly linked to the tropical Pacific. Figure 13c shows the Nino3.4 index along with the DMI, which correlate at +0.56. This result is similar to observations (Annamalai et al. 2001). The relatively strong correlations of all three indices to each other illustrates that the DMI is a part of the large scale Indian-Pacific ocean climate and is an inherent feature of the TBO-ENSO-monsoon system.

d. Cross-equatorial oceanic meridional heat transport

Loschnigg and Webster (2002) examined the interannual variability of the meridional cross-equatorial oceanic heat transport of the Indian Ocean. In their paper, it was hypothesized that the heat transport may be part of a coupled ocean-atmosphere system of SST modulation that may affect the interannual variability of the Indian summer monsoon by decreasing (increasing) the mean heat content in the northern Indian Ocean regions during strong (weak) monsoons. This anomalous heat content would affect the SST in the following seasons and thus influence the strength of the subsequent summer monsoon through Indian Ocean SST-monsoon relationships (Clark et al. 2000).

Figure 14 shows a time series of the heat transport \(Q_v\) across the equator in the Indian Ocean in the CSM during the peak boreal summer monsoon months (JJAS) correlated with the WSI1 index. (Results for the heat transport are displayed only from year 27 onwards, due to model
output availability described in section 2.) The two timeseries are have a fairly strong negative correlation (-0.56, see Table 1), showing that in years of strong monsoon circulation, there is anomalously strong southward (negative) heat transport in the Indian Ocean. This anomalous heat transport affects heat content in the northern Indian Ocean and contributes to patterns of anomalous SST that can persist through multiple seasons to enhance the TBO transition of the monsoon the following boreal spring season. The basic process is that in the “weak” monsoon cycle, there would be less southward oceanic heat transport in JJAS from the reduced Ekman transports, leaving positive heat content anomalies in the northwest Indian Ocean (Loschnigg and Webster 2000) and thus contribute to a pattern of positive SST anomalies. The correlation in the CSM between the JJAS $Q_v$ and the DMI is -0.41, indicating that the heat transport plays a role in the development of the equatorial temperature gradient in the following boreal fall season. SST anomalies then persist in this region through several seasons and help develop a “strong” monsoon cycle the next year, as warm SST in the northwest Indian Ocean has been shown to lead to a strong monsoon (Clark et al. 2000; Meehl and Arblaster 2001a; Meehl and Arblaster 2001b). Thus the cross-equatorial oceanic heat transport is another mechanism whereby coupled dynamics in the Indian Ocean contribute to the TBO that also encompasses the ENSO-monsoon system.

4. Discussion and Conclusion

The NCAR CSM is shown to adequately simulate the seasonal cycle of temperature, precipitation and winds associated with the Asian-Australian monsoon. SVD analyses show the model also captures the salient features of the TBO transition conditions for MAM 500 hpa height, Indian and Pacific Ocean SST and subsequent patterns of Indian monsoon rainfall. The model also captures the patterns of SON Indian Ocean SST (in the form of the IOZM), Pacific SST and southeast
Asia rainfall and subsequent patterns of DJF Australian monsoon rainfall. Single and cumulative anomaly pattern correlations calculated from the first component SVDs are also comparable to observations in that the contribution of the transition conditions vary more from year to year in the Indian monsoon compared to the Australian monsoon. This indicates that, for the model as in observations, contributions of varying strength from year to year can ultimately end up influencing the Indian monsoon. But as those conditions set the strength of the Indian monsoon, there is a consistency of the coupled interaction in the system that carries over to the Australian monsoon. That is, once set, a strong Indian monsoon often leads to a strong Australian monsoon, and vice versa for a weak Indian monsoon.

The coupled ocean dynamics that contribute to the IOZM are shown to be an integral part of the TBO-monsoon system in the Indian Ocean. Through anomalies in upwelling at both the eastern and western sides of the equatorial Indian Ocean, the strength of the monsoon wind circulation affects the ocean dynamics and enhances SST anomalies in this region. The strength of the monsoon winds also affects the magnitudes of the southward summertime oceanic heat transport, which contribute to developing ocean heat content and SST anomalies that persist and affect the following monsoon season. To a large degree, the interrelationship of the components of the Indian Ocean and the monsoon, as suggested by Webster et al. (2002); Webster (2002), are corroborated by associations found in the NCAR CSM. The interaction of these processes are displayed in Fig. 15 (adapted from Webster et al. (2002)), which shows a schematic of a biennial cycle for the Indian Ocean involving the ocean dynamics in the equatorial region. As the starting period is arbitrary, we begin in the boreal summer season during a “weak” monsoon (JJA 0). The reduced monsoon surface wind circulation causes reduced upwelling in the western equatorial regions, and enhanced upwelling in the eastern equatorial regions near Sumatra,
due to the strong along- and off-shore winds there. This begins the development of SST anomalies that are characteristic of a “positive” dipole. Also during this period, the weaker monsoon circulation produces lower than average southward heat transport, leaving larger heat content anomalies in the northern oceanic areas. In the boreal fall months (SON 0), the SST dipole reaches its maximum magnitude with the anomalous equatorial surface easterlies enhancing the SST anomalies on either side of the basin, and tilting the thermocline along the equator, raising it in the east and depressing it in the west. This situation is similar to the sign of anomalies during the 1997-98 period in the Indian Ocean. During this time (SON 0 to DJF +1), the Pacific develops warm SST anomalies in the central and eastern part of the basin, with El Niño conditions prevailing, and an opposite tilt of the equatorial thermocline to that of the Indian Ocean. The SST anomalies that develop through the dipole and which are enhanced through the anomalous heat transport persist through the winter season and assist in developing a “strong” monsoon the following boreal summer (JJA +1). Stronger than normal monsoon surface wind circulation causes upwelling and downwelling anomalies at either sides of the equator in the Indian Ocean that are reversed as compared to those of the previous year. Also, the increased monsoon winds drive an increased southward heat transport, reducing heat content in the northern Indian Ocean. These processes begin to develop a “negative” dipole in the summer months, which then reaches peak magnitude in the boreal fall with the influence of the equatorial westerlies (SON +1). These westerlies also help tilt the thermocline in an opposite direction to the previous summer, with a raised thermocline in the west and depressed in the east. In the Pacific, La Niña conditions are present, with cool SST anomalies in the central equatorial basin and an opposite tilt to the thermocline to that of the Indian Ocean (DJF +2).

The results here show that, contrary to previous modeling study (Iizuka et al. 2000), the IOZM is not independent of the tropical Pacific and ENSO. Rather, it is an integral part of the TBO
and thus the ENSO-monsoon system. Also, consistent with several studies (Baquero-Bernal and Latif 2001; Nicholls 2001; Allan et al. 2001), the only season with a significant SST gradient in the equatorial Indian Ocean region is the boreal fall (SON). The IOZM is thus an inherent feature of the interannual variability of the large-scale Indian-Pacific Ocean region climate, and is not an independently oscillating mode of the Indian Ocean.

One feature of the TBO that may have a significant effect of the near-equatorial SST anomalies that has not been analyzed here is the existence of equatorial waves that may enhance the Indian Ocean zonal SST gradient. In the strong positive SST dipole event of 1997-98, it was hypothesized that equatorial waves contributed to a positive feedback which enhanced the SST anomalies, particularly on the western side of the Indian Ocean basin (Webster et al. 1999). The basic idea that the easterlies in the boreal fall would produce an upwelling Kelvin wave that would raise the thermocline in the east, maintaining the cooling off Sumatra. The equatorial easterlies would also develop an off-equatorial Rossby wave that would propagate westward and deepen the thermocline in the western Indian Ocean. This Rossby wave was very evident in TOPEX data during 1997-98, with most of the strength near 5° to 10° south (Webster et al. 1999). It is also possible that Kelvin waves reflected from the eastern coast to also produce Rossby waves that propagate westward and enhance the thermocline depth there. Meehl and Arblaster (2001b) consider the equatorial waves to be a part of the transitions of SST anomalies in the Indian Ocean (as well as the Pacific), and thus a part of the biennial cycle for this region. We intend to analyze this equatorial wave activity in the NCAR CSM in relation to the TBO and the IOZM in a future paper.

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References


Loschnigg, J., and P.J. Webster, 2002: Interannual variability in an Indian Ocean basin model. Submitted to J. Climate.


Table 1: Correlations of model indices: WSI1 (JJAS U850-U200 wind shear anomalies, 5° to 15°N and 35° to 75°E), DMI (SON SST anomaly difference between 5°S to 10°N, 40° to 55°E and 10°S to 5°N, 90° to 102°E), Nino 3.4 (Annual mean SST anomaly, 5°S to 5°N and 170°W to 120°E), Sumatra (SON SST anomaly, 10°S to 5°N, 90° to 102°E), $Q_v$ (JJAS oceanic meridional heat transport across the equator in the Indian Ocean). All values are significant at the 99.9% level.

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Figure 2: Model climatological mean surface temperature (°C) and 850mb winds (m/sec) for MAM, JJAS, SON and DJF.

Figure 3: First component SVD correlation patterns, relating A) MAM Indian Ocean SST (20°S to 20°N, 20°E to 110°E) to B) JJAS Indian rainfall (5°N to 40°N, 60°E to 110°E); C) MAM Pacific Ocean SST (20°S to 20°N, 150°E to 80°W) to D) JJAS Indian rainfall; E) Asian 500mb height (30°N to 70°N, 50°E to 110°E) to F) JJAS Indian rainfall. Note the correlations in parts a, c and e may extend beyond the areas used in the SVD calculation.

Figure 4: A) Timeseries of individual anomaly pattern correlations for the MAM SVD patterns of Indian Ocean SST, Pacific Ocean SST and Asian 500mb height with the JJAS Indian rainfall. B) Cumulative anomaly pattern correlation for all three patterns in (A). Dashed line shows the mean correlation of 0.46.

Figure 5: First component SVD correlation patterns, relating A) SON southern Indian Ocean SST (25°S to EQU, 90°E to 120°E) to B) DJF Australian rainfall (20°S to 5°N, 100°E to 150°E); C) SON Pacific Ocean SST (20°S to 20°N, 150°E to 80°W) to D) DJF Australian rainfall; E) SON southeast Asian rainfall (10°S to 10°N, 90°E to 130°E) to F) DJF Australian rainfall.

Figure 6: A) Timeseries of individual anomaly pattern correlations for the SON SVD patterns of Indian Ocean SST, Pacific Ocean SST and southeast Asian rainfall with the DJF Australian rainfall. B) Cumulative anomaly pattern correlation for all three patterns in (A). Dashed line shows the mean correlation of 0.42.

Figure 7: A) Model climatological 850mb-200mb U-shear for JJAS (m/sec). B) Correlation of Model CI1 to 850mb-200mb U-shear for JJAS. Boxed area shown in both A) and B) is the region used for the calculation of the WSI1 U-shear index (5°N to 15°N, 35°E to 75°E).
Figure 8: A) Timeseries of normalized CI1 and WSI1 anomalies. Red dots and blue dots signify strong and weak TBO years, respectively, where both the CI1 and WSI1 are relatively greater or less than preceding and following years. B) Power spectrum of the monthly values of CI1, with the 95% significance level shown as the dotted line. C) Power spectrum of the monthly values of WSI1.
Figure 9: Composite of anomalies of $T_s$ and 850mb winds for strong minus weak TBO years for A) MAM (0), B) JJA (0), C) SON (0), D) DJF (+1), E) MAM (+1) and F) JJA (+1). Only values above the 95% significance level are plotted.

Figure 10: Composite of anomalies of total precipitation winds for the strong minus weak TBO years for A) JJA (0), B) SON (0), C) DJF (+1), D) MAM (+1), E) JJA (+1) and F) SON (+1). Units are mm/day.

Figure 11: Composite anomalies of oceanic temperature vs. depth at 0.6°N in the Indian Ocean for strong minus weak TBO years for A) MAM (0), B) JJA (0) C) SON (0), D) DJF (+1), E) MAM (+1) and F) JJA (+1). Units are °C.

Figure 12: Composite anomalies of oceanic vertical velocity at 50 meters in the Indian Ocean for strong minus weak TBO years for A) JJA (0), B) SON (0), C) DJF (+1), D) MAM, (+1), E) JJA (+1) and F) SON (+1). Units are m/day. Positive values indicate upward motion.

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