NOTES AND CORRESPONDENCE

Influence of the Indian Ocean Dipole on the Southern Oscillation

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Abstract

The influence of the Indian Ocean Dipole (IOD) on the interannual atmospheric pressure variability of the Indo-Pacific sector is investigated. Statistical correlation between the IOD index and the global sea level pressure anomalies demonstrates that loadings of opposite polarity occupy the western and the eastern parts of the Indian Ocean. The area of positive correlation coefficient in the eastern part even extends to the Australian region, and the IOD index has a peak correlation coefficient of about 0.4 with the Darwin pressure index, i.e. the western pole of the Southern Oscillation, when the former leads the latter by one month. The correlation analysis with seasonally stratified data further confirms the lead role of the IOD. The IOD-Darwin relation has undergone interdecadal changes; in the last 50 years the correlation is highest during the most recent decade of 1990–99, and weakest during 1980–89.

1. Introduction

The Southern Oscillation (SO) is traditionally known as a seesaw in the atmospheric sea level pressure between the region from the central to the eastern tropical Pacific and the region covering the tropical Asia-Pacific. Historically, the SO study has its root in the predictability of the Indian summer monsoon (Walker 1924). The interesting seesaw in the atmospheric pressure field was later linked to the predictability of the Indian summer monsoon (Walker 1924). The interesting seesaw in the atmospheric pressure field was later linked to the well-known SST variability in the eastern tropical Pacific Ocean by Bjerknes (1969). This combined phenomenon, commonly known as the El Niño-Southern Oscillation (ENSO), is identified as the most dominant climate signal that affects the weather, and climate worldwide. For further understanding of the ENSO theory and observations, readers may refer to Philander (1990) and the recent review articles of Neelin et al. (1998) and Wallace et al. (1998).

The Bjerknes hypothesis, in light of these recent studies, is that ENSO evolves as a self-sustained coupled oscillation in which warm/cold anomalies of SST in the tropical Pacific cause the anomalous strengthening/weakening of the trade winds that, in turn, drive the ocean currents associated with the SST anomalies. The oceanic process provides a key to understand the observed 3–7 years periodicity of
ENSO, which is rather slow as compared to the fast evolving atmospheric processes. Although the oceanic component of ENSO is a key element that determines the SO, influences from other climate variations outside of the Pacific cannot be neglected in this regard (Zhang et al. 1997). In a previous study, Deser and Wallace (1987) showed that the major negative swings of the Southern Oscillation that occurred in 1963 and 1977 were not accompanied by strong El Niño events. Hence, they concluded that the El Niño and Southern Oscillation are more loosely coupled than numerous past studies would suggest. The motivation of the present short article is due to the fact that the Indian Ocean Dipole (IOD) discovered recently (Saji et al. 1999; Webster et al. 1999; Behera et al. 1999; Iizuka et al. 2000; Yamagata et al. 2002), has suggested a new role of the Indian Ocean in determining the sea level pressure oscillation over the Asia-Pacific sector. Since its existence has come to the light very recently, a brief description of the Indian Ocean phenomenon is provided here.

Saji et al. (1999) found that the coupled ocean-atmosphere phenomenon evolves with an east-west dipole in the SST anomaly, and named it the Indian Ocean Dipole. The Dipole Mode Index (DMI) is thus defined as the SST anomaly difference between the eastern and the western tropical Indian Ocean (see insets in Fig. 3a for the regions used to compute the DMI). The changes in the SST during the IOD events are found to be associated with the changes in the surface wind of the central equatorial Indian Ocean. In fact, winds reverse direction from westerlies to easterlies during the peak phase of the positive IOD events when SST is cool in the east and warm in the west. The effect of the wind is even more significant at the thermocline depths through the oceanic adjustment process (Rao et al. 2002); the thermocline rises in the east and deepens in the central and western parts. Since the seasonal southeasterly winds along the Sumatra coast are also strengthened during the positive IOD events, the induced coastal upwelling causes strong SST cooling in the east (Behera et al. 1999).

These changes in surface winds and oceanic conditions are coupled with the changes in the atmospheric convection through either convergence or divergence of the moist air. The convection in the eastern Indian Ocean is suppressed during the positive IOD events, whereas the convection in the central and western Indian Ocean is enhanced. Behera et al. (1999) have shown that the coupling between the SST cooling and the suppressed convection in the eastern Indian Ocean gave rise to further easterly wind anomalies during the typical IOD event of 1994. Using a simple Matsuno-Gill type model, Saji and Yamagata (2001) have also shown that the atmospheric linear response to such changes in the heat source of the eastern Indian Ocean strengthens the easterly wind anomalies, which in turn are responsible for the evolution of the SST dipole. Ashok et al. (2001) have further confirmed this using a more sophisticated general circulation model.

It is clear from the above discussions that the positive IOD event produces anomalous atmospheric circulation, wherein an anomalous downdraft (updraft) branch of the Walker circulation is seen in the region that covers from the eastern tropical Indian Ocean to the western tropical Pacific (from the western tropical Indian Ocean to the east African region). The conditions are just opposite during a negative IOD event (e.g., Ansell et al. 2000). Therefore, the air-sea coupled phenomenon in the tropical Indian Ocean may affect the atmospheric pressure fluctuation in the region around Indonesia-Australia. Here, the existence of such an interesting relation is reported by analyzing the observed data.

2. Data and methods

We use mean monthly data for the period from 1958 to 1999 in this study. Monthly anomalies of sea level pressure and surface wind fields are derived from the NCEP-NCAR reanalysis data (Kalnay et al. 1996). SST anomalies are computed from the GISST 2.3b dataset (Rayner et al. 1996). The low frequency variability, comprising of the periodicity of above 7 years is removed from all the data sets using the Fourier transformation. The values of Darwin pressure index and SOI are obtained from the NOAA Climate Prediction Center web site (http://www.cpc.ncep.noaa.gov/data/indices). Following Saji et al. (1999) the DMI is defined as the SST anomaly difference between western (50°E–70°E, 10°S–10°N) and
eastern (90°E–110°E, 10°S-Eq) tropical Indian Ocean (Fig. 3a shows the regions). Niño-3 index is derived from the GISST data.

Simple statistical tools, such as composite technique and correlation method, are used to address the inverse relation. Besides the simple correlation method, a partial correlation technique is also used to show exclusive relationship between two variables while excluding influence arising from another independent variable. For example, the partial correlation between DMI and global sea level pressure anomalies, while excluding the relation arrived because of the correlation between Niño-3 and global sea level pressure anomalies, is defined as follows.

\[
r_{13,2} = \frac{(r_{13} - r_{12} \cdot r_{23})}{\sqrt{(1 - r_{12}^2) \sqrt{(1 - r_{23}^2)}},
\]

where \(r_{13}\) is the correlation between DMI and global sea level pressure anomalies, \(r_{12}\) is the correlation between DMI and Niño-3 index and \(r_{23}\) is the correlation between Niño-3 and global sea level pressure anomalies. Similarly, the partial correlation can also be obtained for Niño-3 and sea level pressure anomalies while excluding the influence arising due to the IOD. Statistical significance of the correlation coefficients is determined by a 2-tailed “t-test”.

In section 4, the wavelet power spectrum analysis (Torrence and Compo 1998) is used to discuss the dominant time-frequency variability of the DMI. Unfiltered monthly anomalies are used for this analysis, and zero padding is applied to reduce wraparound effects. The Morlet wavelet is considered here as the mother wavelet, and the transform is performed in Fourier space using the method described in Torrence and Compo (1998). The Morlet wavelet function is given as;

\[
\psi_{\omega_0}(t) = \pi^{-1/4} e^{i \omega_0 t/s} e^{-t^2/2s^2},
\]

where \(t\) is the time, \(s\) is the wavelet scale, and \(\omega_0\) is a nondimensional frequency. The wavelet power spectrum is then defined as the absolute value squared of the wavelet transform. The null hypothesis for the wavelet power assumes that the time series has a mean power spectrum, viz., the global wavelet spectrum. When a peak in the wavelet power spectrum is significantly above this background spectrum, it is assumed to be a true feature with a certain degree of significance. For example, the 95% confidence level (significant at 5%) means that it is the 95th percentile of the background spectrum. Details of the wavelet method, and the measures of significance are discussed in Torrence and Compo (1998).

Recent data sets from 1958 to 1999 are used in most of the analyses. Besides, extended data sets for the period from 1880 to 1999 are also used in section 4 for understanding the inter-decadal nature of the IOD-Darwin relationship.

3. The tropical Indian Ocean Dipole and the Darwin pressure

A partial correlation (detail in section 2) between the DMI and the sea level pressure anomalies clearly shows a dipole pattern in the sea level pressure anomalies of the tropical Indian Ocean. The positive (negative) correlation coefficient peaks at 0.7 (∼0.35) in the eastern (western) Indian Ocean (Fig. 1). Further, the positive correlation coefficients are seen over a wider region in the eastern Indian Ocean as compared to that of the negative correlation coefficients in the western Indian Ocean. Thus, anomalous conditions associated with the IOD have a greater impact on the eastern side as compared to the western side. For a comparison between this pressure dipole and the SST dipole as discussed in Saji et al. (1999), the time series of both indices are shown in Fig. 2. The pressure index is obtained by taking the anomaly difference between the sea level pressure in the eastern (96°E–100°E, 13°S–9°S) and the western (52°E–56°E, 9°S–5°S) tropical Indian Ocean. As seen in the figure, time series of the pressure index corresponds well with that of the SST index and the correlation coefficient between the two indices is 0.65 (0.74 for June–November), when the latter leads the former by one month.

It is also found that the dipole pattern in the correlation analysis could be reproduced using the composites of sea level pressure anomalies (figure not shown). The close resemblance found in different analyses clearly emphasizes the existence of the air-sea coupled phenomenon in the Indian Ocean. The pattern is different when the sea level pressure anomalies are partially correlated with the Niño-3 index by removing the IOD influence (Fig. 1b); we don’t find a dipole pattern over the tropical Indian Ocean region in this case.
Fig. 1. Partial correlation during the peak IOD phase (July–November) between the time series of a) DMI and global sea level pressure anomalies when linear effects of Niño-3 is removed, b) Niño-3 and global pressure anomalies when linear effects of DMI is removed. Values exceeding 95% confidence limit using a 2-tailed t-test are shaded.

Fig. 2. The normalized dipole mode indices for the anomalies of sea level pressure (thick line) and SST (thin line). The time series are normalized by the respective standard deviation values. The correlation between them when SST index lead the pressure index by one month is shown in the inset.
Figure 3 shows lag correlation functions between the DMI, and each of the pressure indices from the Pacific. Besides the three well-known pressure indices of Darwin, Tahiti and Southern Oscillation Index (SOI), three additional pressure indices derived from anomalies of sea level pressure are used here; the central-west Pacific index ($160^\circ$E–$180^\circ$E, $6^\circ$S–$6^\circ$N), the eastern Pacific index ($110^\circ$W–$90^\circ$W, $6^\circ$S–$6^\circ$N), and the tropical Pacific index (difference of the previous two indices) (see Fig. 3a for the detail). The lag-correlation functions between the DMI and each of these indices are calculated by starting DMI time series 6 months prior, and then sliding it up to 6 months after the start of the year. This is expected to provide the time evolution of the relationship between the IOD and the pressure fluctuation of the tropical Pacific Ocean. As seen in Fig. 3b, the DMI has a peak correlation of 0.4 with the Darwin pressure index when the former leads the latter by one month. When the Darwin index leads by 3 months, the correlation (0.2) drops below the significant level. The central-west Pacific index has a peak correlation of about 0.3 when DMI leads by 4–5 months (Fig. 3b). Similar value is also seen for the DMI-SOI correlation. However, the correlation improves to −0.4 when we consider a correlation between the tropical Pacific index and the DMI. These results may explain the reason why some earlier studies found the progressive patterns from the tropical Indian Ocean to the tropical Pacific Ocean (e.g., Tomita and Yasunari 1996). In contrast, the correlation between the DMI and the pressure index of Tahiti/eastern Pacific is insignificant at all lead/lag periods. Even if the analysis is repeated by removing the strong El Niño years from the time series of all the indices, the same conclusion is reached (Fig. 3c).

The seasonal evolution of the DMI-Darwin correlation is further explored using seasonally stratified cross-correlation matrix (Table 1). As found in the table, a peak correlation coefficient of 0.71 is achieved for the simultaneous correlation during the mature IOD season of September–November. The coefficient continues to remain quite strong at 0.58 one season after (December–February) the peak correlation, i.e., when DMI leads the Darwin index by one season. More importantly, the simultaneous correlation coefficient of 0.49 for
June–August peaks at 0.55 when the same season DMI is correlated with the next season (September–November) Darwin index. The lead role of the DMI remains quite strong even after removal of the eastern Pacific influence (Table 2). In this case, the correlation coefficient is above the significant level from the boreal spring. On the other hand, the Darwin index has no significant lead correlation with the DMI. Thus, it is clear that the DMI leads the Darwin index in the cross-correlation matrix consistent with the previous analysis as shown in Fig. 3. Furthermore, the DMI also has a lead role in its correlation with the central-west Pacific index (Table 3). Interestingly to note here is that the correlation between the September–November DMI and the December–February central-west Pacific index is stronger than the corresponding DMI-Darwin correlation. The seasonal evolution is consistent with the view as discussed earlier that IOD influence progressively advances from the eastern Indian Ocean to the western Pacific Ocean.

4. Interdecadal change in the IOD influence on the Darwin pressure

In this section the decadal modulation of the IOD-Darwin relationship is examined. For this purpose, longer time series (starting from January 1880 to December 1999) SST data and Darwin pressure index are used. The correlation coefficients between the DMI and the Darwin pressure index are separately computed for each decade; the whole analysis period of 120 years is equally divided into 12 decades. The correlation is highest (0.44) during the most recent decade of 1990–99 (Fig. 4); when two of the strongest IOD events of 1994 and 1997 occurred. During the last 50 years, which is considered as the most reliable period for the SST data, the correlation is above 99% significance level during 1950–59, 1960–69, 1970–79. This signifies the role of IOD in the pressure swings of the Asia-Pacific sector. Interestingly, such a relation drops below the significant level during 1980–89. During this decade, three ENSO events evolved against one IOD event of 1982. Thus, it is natural to expect that such a situation lessened the IOD effect on the Darwin pressure. Similar decadal modulation of the IOD effect on the Indian monsoon rainfall is recently discussed by Ashok et al. (2001).

The DMI correlation with the Darwin pressure anomaly, prior to 1950, remained below the statistical significance level during several decades (Fig. 4). Especially the relation is very weak during 1930–1949, when we find only
a few IOD events. During the period when the IOD occurrences are more frequent, IOD-Darwin correlation is above the significant level. The wavelet power spectrum analysis of the DMI time series (Fig. 5) shows significant amount of power at the quasi-biennial period during the decades of high correlation. We also found that the zonal wind anomaly index (Saji et al. 1999) of the central Indian Ocean (70°E–90°E and 5°S–5°N) has similar quasi-biennial periodicity (figure not shown) concurrent with the DMI in the last 50 years period. This emphasizes the presence of the ocean-atmosphere coupling during the active decades of the IOD. The absence of such a coupling process in the Indian Ocean may be a key to understand the weak IOD-Darwin correlation during certain decades. The ocean dynamics that lead to the quasi-biennial behavior during the active and break phase of the IOD events are discussed in Rao et al. (2002). As shown in that study, near-equatorial Rossby waves carry the signal from the eastern to the western Indian Ocean, and then reflect back to the equatorial wave-guide as Kelvin wave to complete a cycle in about 2 years. This behavior of the equatorial Indian Ocean may also give rise to the quasi-biennial oscillation of the Indo-Pacific sector, as discussed by Meehl (1987) and needs further analysis.
5. Summary and discussion

It is shown that the SST variability in the Indian Ocean, associated with the IOD, influences the evolution of sea level pressure anomaly in the region that spans from the western tropical Indian Ocean to the western tropical Pacific region. During the positive IOD events, we observe positive (negative) sea level pressure anomalies in the Indonesia-Australia (western Indian Ocean) region. The dipole pattern in sea level pressure anomalies that is associated with the IOD is very different from a basin-wide pattern associated with the ENSO. A broad 3–6 year period spectral peak is found in the Niño-3 index, whereas the spectral peak in the IOD index is at around 2 years. Thus the inherent periodicity associated with the IOD events, which is different from that of the Pacific ENSO events, may provide co-variability between the IOD and the pressure fluctuation at Darwin, i.e., one pole for the Southern Oscillation. The correlation analysis in the present study clearly supports this hypothesis; the IOD index shows a significant correlation with the sea level pressure anomaly at Darwin. The correlation analysis further shows the lead role of IOD in determining the evolution of such a correlation with the Darwin and central-west Pacific indices.

Though the IOD correlation with the pressure variations in the eastern Pacific is insignificant, the former has a significant correlation with the SOI, and the pressure difference index of the tropical Pacific. It can be explained by the fact that pressure variations in the western Pacific will set up anomalous winds that force the oceanic Kelvin waves initiating changes in the eastern Pacific. The above novel relationship, however, undergoes decadal modulation. A significant reduction of the IOD impact on the Darwin pressure is observed for the recent decade; 1980 through 1989. Understanding the physical mechanism that determines such decadal modulations of the ocean-atmosphere coupled system in the Indo-Pacific sector is underway using sophisticated coupled ocean-atmosphere models.

The analysis here has brought out an interesting aspect of the climate variation in the tropical oceans. Although either of the independent modes of the Pacific and the Indian Oceans, i.e., the ENSO and the IOD, has a distinct air-sea coupled nature in the respective basins, two basins interact through atmospheric conditions. Therefore, it is interesting to construct first the real nature that is uniquely associated to the air-sea coupled phenomenon in each of the tropical oceans, and then to clarify the interactions among these phenomena. This work is currently underway.

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References


