Decadal Modulations of the Indian Ocean Dipole in the SINTEX-F1 Coupled GCM

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(Manuscript received 6 April 2005, in final form 30 August 2005)

ABSTRACT

The decadal variation in the tropical Indian Ocean is investigated using outputs from a 200-yr integration of the Scale Interaction Experiment-Frontier Research Center for Global Change (SINTEX-F1) ocean–atmosphere coupled model. The first EOF mode of the decadal bandpass- (9–35 yr) filtered sea surface temperature anomaly (SSTA) represents a basinwide mode and is closely related with the Pacific ENSO-like decadal variability. The second EOF mode shows a clear east–west SSTA dipole pattern similar to that of the interannual Indian Ocean dipole (IOD) and may be termed the decadal IOD. However, it is demonstrated that the decadal air–sea interaction in the Tropics can be a statistical artifact; it should be interpreted more correctly as decadal modulation of interannual IOD events (i.e., asymmetric or skewed occurrence of positive and negative events). Heat budget analysis has revealed that the occurrence of IOD events is governed by variations in the southward Ekman heat transport across 15°S and variations in the Indonesian Throughflow associated with the ENSO. The variations in the southward Ekman heat transport are related to the Mascarene high activities.

1. Introduction

The Indian Ocean dipole (IOD) is an air–sea coupled phenomenon associated with a positive sea surface temperature anomaly (SSTA) to the west and a negative SSTA to the east (Saji et al. 1999; Webster et al. 1999). It has turned out that the IOD has a large impact on the climate of both the surrounding and remote regions such as east Asia, Europe, and South America (Guan and Yamagata 2003; Saji and Yamagata 2003; Yamagata et al. 2004). Thus, understanding, as well as simulating, the IOD are crucial to global climate research.

Recently, Ashok et al. (2001) have shown that the Indian summer monsoon rainfall (ISMR) is not correlated with the Southern Oscillation index (SOI) in recent decades owing to the frequent occurrences of the IOD. Behera and Yamagata (2003) have revealed that the IOD even weakens the link between El Niño and the Southern Oscillation by influencing the pressure at Darwin. Since the seminal work by Nitta and Yamada (1989), the decadal variability in the tropical Pacific has been studied extensively in recent decades. However, the decadal phenomenon in the tropical Indian Ocean has received little attention, partly owing to a lack of available observational data. Among a few studies, Meehl et al. (1998) and Allan et al. (2003) discussed a link with the decadal ENSO-like variability in the Pacific. Cole et al. (2000) suggested that the decadal variability in the proxy records of SST from the western Indian Ocean is related to the decadal variation in the tropical Pacific. Also, Allan et al. (1995) and Reason et al. (1996a) suggested the importance of decadal variability in the subtropical anticyclone in the southern Indian Ocean, while Reason et al. (1996b) showed that the decadal variability in the winds over the Pacific may...
introduce decadal variations in the southern Indian Ocean via the Indonesian Throughflow. More recently, Ashok et al. (2004) have revealed, using both data and coupled model results, the existence of a strong “decadal IOD.” They claimed that it is not correlated with the decadal ENSO events and suggested that ocean dynamics are involved in the decadal IOD. However, their analysis was limited because of the coarse resolution of their coupled model and the sparseness of observational data.

In the present article, we use outputs from a 200-yr integration of a coupled general circulation model (CGCM) with a horizontal high-resolution atmospheric component, and focus on the decadal climate variability in the tropical Indian Ocean. The content is organized as follows. A brief description of the CGCM along with its validity is given in the next section. In section 3, two modes of decadal variability in the tropical Indian Ocean are presented. In particular, a detailed discussion on the real nature of the decadal IOD is given there. The final section summarizes the main results.

2. Model

a. Model description

The model data used in this study are obtained from an atmosphere–ocean–land CGCM run on the Earth Simulator. The model is the Scale Interaction Experiment-Frontier Research Center for Global Change SINTEX-F1 model (Luo et al. 2003), which is an upgraded version of the SINTEX model (Gualdi et al. 2003; Guilyardi et al. 2003). The atmospheric component T106L19 of the European Community–Hamburg (ECHAM-4; Roeckner et al. 1996) model is coupled to the oceanic component of the Ocean Parallelise ocean general circulation model (OPA-8.2; Madec et al. 1998) using the Ocean Atmosphere Sea Ice Soil (OASIS 2.4; Valcke et al. 2000) coupling software package. No measures for flux adjustments are taken in the model. For the AGCM, a semi-Lagrangian transport method (Rasch and Williamson 1990) is used for the advection of cloud water and water vapor, while the parameterization of Tiedtke (1989) is used to represent convection and that of Morcrette (1991) is used for radiation. The horizontal resolution of the OGCM is $2^\circ \times 2^\circ$ cosine (latitude) with an increased meridional resolution of up to 0.5° near the equator. There are 31 levels in the vertical with 19 levels in the upper 400 m. The horizontal mixing of momentum is of the Laplacian type, while the “quasi-pure” isopycnal mixing (Guilyardi et al. 2003) is used for the horizontal mixing of tracers. Vertical eddy diffusivity and viscosity are computed from a 1.5-order turbulent closure scheme (Blanke and Delecluse 1993). The monthly mean output from the last 200 yr of the total 220-yr model integration is used for analysis after removing a linear trend using a least squares fit. Details of the CGCM can be found in the above references.

b. Model validation: Interannual variability

To check the model performance, we first analyze how the IOD events are resolved in the model. Since the nature of IOD events locked to the annual cycle is well reproduced (Yamagata et al. 2004), composite diagrams are constructed by averaging the 30 strongest positive events. To define the event strength, we use the dipole mode index (DMI) introduced by Saji et al. (1999), which is defined by the difference in SSTA between the tropical western Indian Ocean ($10^\circ$S–$10^\circ$N, $50^\circ$–$70^\circ$E) and the tropical southeastern Indian Ocean ($10^\circ$S–equator, $90^\circ$–$110^\circ$E). For observation, six positive events (1961, 1967, 1972, 1982, 1994, and 1997) are used following Saji et al. (1999). We have used the extended reconstructed SST (ERSST) data of Smith and Reynolds (2003) for SST, the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) for wind stress and outgoing longwave radiation (OLR), and the Simple Ocean Data Assimilation (SODA; Carton et al. 2000) for sea surface height (SSH).

Figure 1 compares the model results with the observations for October of year 0, when positive IOD events reach their maximum strength. The reversal in sign of the SST across the basin with the easterly wind stress anomaly along the equator is well reproduced in the model, but the eastern pole is too cold and extending too far to the west, compared to the observations. This is due to the flatter equatorial Indian Ocean thermocline and the shallower thermocline in the eastern equatorial Indian Ocean owing to the weaker simulated westerly wind stress along the equator. This results in excessive westward expansion of the positive OLR anomaly and a precipitation deficit in comparison to the NCEP–NCAR reanalysis data (cf. Yamagata et al. 2004). All of these biases remind us of the case for El Niño, in which the simulated positive SST also extends too far west in all of the CGCMs (Latif et al. 2001). The subsurface dipole mode (e.g., Rao et al. 2002) is also resolved in the model with the negative sea surface height anomaly (SSHA) to the east and the positive SSHA to the west.

Next, we examine the relationship between IOD events and the simulated ENSO events. The correlation between the model DMI and Niño-3 ($5^\circ$S–$5^\circ$N, $90^\circ$–$150^\circ$W) index is 0.40 for the whole year and 0.51 for the
boreal fall. This is in agreement with the observations; the correlation is 0.33 for the whole year and 0.53 for the boreal fall (Yamagata et al. 2004). In addition, a conventional EOF analysis of the SSTA captures a monopole ENSO-related mode as the leading mode with a variance contribution of 32%. The dipole mode appears as the second mode, explaining 23% of the total variance. This larger variance contribution of the

![Composite diagrams for (top to bottom) SSTA, zonal wind stress anomaly (TAUXA), OLR anomaly (OLRA), and SSHA in October (year 0) for (left) the observations and (right) the SINTEX-F1 model. Units are in °C, N m⁻², W m⁻², and cm, respectively. The statistical significance of the anomalies is estimated by the two-tailed t test. Shading indicates anomalies exceeding 95% significance.](image-url)
second mode compared to 14% in the observation is due to the shorter period and the larger amplitude of the simulated IOD events (Fig. 1b).

In summary, many features during the positive IOD events are relatively well reproduced. For further comparison, readers are referred to Behera et al. (2005) and Yamagata et al. (2004) for the tropical Indian Ocean, and Luo et al. (2003) and Tozuka et al. (2005) for the tropical Pacific.

3. Decadal variation

a. Simulated decadal variability

The model DMI shows the existence of decadal variations in addition to interannual variations (Fig. 2). A wavelet analysis captures clearly the lower-frequency variations as well as the IOD events as major variations with a period of about 4 yr (Fig. 3a). Variations with a period of 9 yr or longer are significant during, for example, years 40–80 and 110–140. A similar decadal signal is also found in the observed DMI constructed from the ERSST data (Smith and Reynolds 2003) from 1854 to 2002 (Fig. 3b).

To capture the dominant modes that determine the decadal variability, we have applied the EOF analysis to the bandpass-filtered (9–35 yr) SSTA. The bandpass filter we used is that of the complex Morlet wavelet transform (Torrence and Compo 1998). The first EOF mode, which explains 37% of the total decadal variance, represents a basinwide mode (Figs. 4a and 4c). Its principal component is correlated well with the decadal Niño-3 SSTA index, suggesting a close connection with the Pacific ENSO-like decadal variability; the simultaneous correlation coefficient is 0.76. In particular, the surface heat flux plays an important role; the correlation between the principal component and the area-averaged surface heat flux north of 15°S amounts to 0.53, with the latter leading the former by 2 yr. Also, a composite analysis of the decadal net heat flux anomaly based on the principal component shows a positive (negative) anomaly in the Tropics for the positive
Fig. 4. (a),(c) Principal components and (b),(d) spatial amplitudes of the first and second EOF modes of the simulated decadal (9–35 yr) SSTA. Also shown are (e),(g) principal components and (f),(h) spatial amplitudes of the first and second EOF modes of the observed decadal (9–35 yr) SSTA. Contour interval is 0.01°C and negative anomalies are shaded for the spatial amplitude.
Thus, we may conclude that the surface heat flux variation related to the decadal ENSO is responsible for the first EOF mode.

On the other hand, the second EOF mode, explaining 14% of the variance in the decadal band, shows a clear east--west dipole pattern (Figs. 4b and 4d). In contrast with the first mode, this second mode shows almost no correlation with the decadal ENSO; the simultaneous correlation coefficient is only 0.02 and the correlation coefficient is 0.19 when this mode leads the decadal Niño-3 SSTA by 4 yr. Since the spatial pattern resembles that of the interannual IOD (Saji et al. 1999; Webster et al. 1999), we may call this mode the decadal IOD mode hereafter. For comparison, we have also applied the EOF analysis to the observed SSTA (Figs. 4e–h). We obtain a basinwide pattern in the first mode (58% of the variance) and a dipole pattern in the second mode (12% of the variance).

Figure 5 shows the typical evolution of the decadal SST, wind, and SSH anomalies from years −6 to +6. Here, composites are calculated from nine positive decadal events with amplitudes that are larger than one standard deviation of the decadal DMI. The negative

Figure 5. Composite diagrams of the decadal (a) SST (contour interval, 0.05°C), (b) wind stress (unit, N m$^{-2}$), and (c) SSH (contour interval, 1 cm) anomalies for years −6 to +6. Shading indicates anomalies exceeding 95% significance.
SSTA along the Indonesian coast is first generated by the alongshore southeasterly wind anomaly (year −3), and the warm SSTA gradually develops in the western equatorial Indian Ocean (year 0) as a result of the zonal advection anomaly and a positive latent heat flux anomaly (figure not shown). This, in turn, leads to intensification of the easterly wind stress anomaly along the equator. The correlation coefficient between the decadal DMI and the decadal zonal wind index is −0.75. Here, we have defined the decadal zonal wind index based on the principal component of the first EOF mode in the decadal zonal wind stress anomaly. The off-equatorial positive SSHAs generated by this easterly wind stress anomaly appear to propagate westward in the off-equatorial region. This further enhances the warm SSTA in the western equatorial Indian Ocean. After reflecting at the western boundary, they propagate eastward along the equator, terminating the positive “decadal IOD” event by year +3. Interestingly, the above subsurface variability, except for the time scale, resembles that of the interannual subsurface dipole mode revealed by Rao et al. (2002). However, the propagation speed is much slower compared with that of the interannual IOD, which is in agreement with Ashok et al. (2004). In addition, the shortwave radiation acts as a negative feedback as positive (negative) anomalies are seen in the eastern (western) equatorial Indian Ocean. Then, the negative decadal IOD event starts to emerge in year +6. We note that a negative decadal IOD event is almost a mirror image of the positive event described here, although the amplitude is somewhat weaker compared with that of the positive decadal IOD.

It is hard to imagine, however, that tropical ocean dynamics permit such a slow propagation of ocean waves (cf. Gill 1982). We, therefore, suspect that the variations in the decadal frequency band are generated by an asymmetric (or skewed) occurrence of positive and negative events as shown in the appendix. For example, the off-equatorial Rossby wave propagation around 7°N/S cannot account for such a decadal time scale. A closer look at the time series of the DMI shown in Fig. 2 and the wavelet power spectrum in Fig. 3 supports this view. The positive decadal IOD event of years 126–133, for example, is associated with four successive interannual positive IOD events without a single negative event, whereas the negative decadal IOD event of years 213–220 is associated with three successive negative IOD events without a single positive event. This suggests that the above decadal IOD may be just a statistical/mathematical artifact. Therefore, it will be more reasonable to interpret the phenomenon as the *decadal modulation of interannual IOD events*.

Then, what causes these modulations? Can the IOD itself introduce its decadal variation? Or does some external forcing play an important role? In Fig. 5, large decadal anomalies in wind anomalies and SSHAs are seen in the subtropical southern Indian Ocean. It appears that those anomalies are involved in the evolution of the decadal variation in the Tropics. Since the first law of thermodynamics is the basic principle in understanding the change in energy of a climate system, we will address the variations of these heat budgets in the next subsection (cf. Boccaletti et al. 2004).

*b. Mechanism of decadal IOD*

We have divided the Indian Ocean into 27 boxes as shown in Fig. 6, and calculated the heat budget (cf. Philander and Pacanowski 1986). Here, we used a reference temperature of 0°C in calculating the heat transport. The heat transport below 440 m is much smaller compared with that of the upper 440 m and will not be discussed in the present study. The annual mean surface heat flux shows the well-known north–south antisymmetric pattern (e.g., Hsiung et al. 1989); the ocean gains heat mostly to the north of 15°S and loses heat to the south of 15°S (Fig. 6a). In particular, the largest heat gain occurs along the equator, the African coast, and the Indonesian coast, while the largest heat loss occurs in the Agulhas Current region and along the west coast of Australia. As a whole, the Indian Ocean does not gain or lose heat through the sea surface in an annual mean basis in our coupled model. This is consistent with the estimate of 0.1 ± 0.2 PW from the observational study of Ganachaud et al. (2000). Since regions of heat gain and those of heat loss are connected to keep the ocean basically in a steady state, heat must be transferred to the south across 15°S (Figs. 6a and 6b). In the upper layer (0–50 m), the most dominant heat transport is associated with the southward Ekman heat transport owing to the summer monsoon north of 25°S. This southward transport in summer dominates the northward heat transport in winter and contributes to the heat loss from the Tropics on an annual mean. Also, the South Equatorial Current (SEC) and western boundary currents (East Madagascar Current, Mozambique Current, and Agulhas Current) transport a large amount of heat. In particular, the large heat transport in the lower layer (50–440 m) is associated with the western boundary currents and the SEC. The direction of the net meridional transport is northward in the lower layer so that it compensates the heat loss in the lower layer to the north of 15°S. The upwelling north of 15°S is responsible for this meridi-
onal circulation. The strong downwelling associated with an anticyclonic wind stress curl of the Mascarene high transports heat downward across the depth of 50 m and drives this cross-equatorial shallow cell (cf. Schott et al. 2004).

The above annual mean heat budget may be summarized with the schematic diagram in Fig. 7. The Indian Ocean of the present interest gains 0.35 PW from the surface heat flux, 1.09 PW from the Indonesian Throughflow, and 0.09 PW from the vertical heat flux, while it loses 1.53 PW by the southward heat transport across 15°S. The amount of heat transport across 15°S and that by the Indonesian Throughflow are within the range of estimates from past studies (Godfrey 1996; Ganachaud et al. 2000; Talley 2003). About 70% of the meridional heat flux across 15°S is associated with the Indonesian Throughflow, whereas only about 30% has its origin in the accumulation of the surface heat flux in the northern Indian Ocean. This ratio is in agreement with the OCMC results of Garternicht and Schott (1997).

To examine how the annual mean picture of the heat budget is altered by the IOD, we have first constructed composite diagrams for biennial IOD events as a canonical IOD event is associated with a biennial tendency in both the observation (Saji et al. 1999; Rao et al. 2002; Meehl et al. 2003) and the present model (Fig. 8). Here, we have only chosen 24 biennial IOD events, where a positive event is followed by a negative event. Although there are 16 biennial IOD events with a negative event followed by a positive event, they are qualitatively similar to the mirror image of those discussed.

FIG. 6. The annual mean heat flux (PW) across various surfaces in the model: (a) in the upper 50 m, (b) between 50 and 440 m, (c) the surface heat flux, and (d) the heat flux across the interface between the upper and lower layers.

FIG. 7. Schematic diagram of the annual mean heat budget for the upper 440 m.
here and will not be described in this paper. Given that the simulated and observed IOD events show clear similarity, we assume that the model can also provide some insight into the interannual variation of the heat budget in the Indian Ocean. As shown in Fig. 8, the total heat content anomaly of the tropical band increases steadily from January of year 0 to January of year +1. The meridional flux, the Indonesian Throughflow flux, and the surface flux contribute to this increase, while the vertical flux tends to dampen this tendency. Both the meridional flux and the surface flux have their maxima in November. This maximum in surface flux is associated with an increase in downward shortwave radiation at the surface as the negative SSTA in the eastern equatorial Indian Ocean suppresses convective activity (Fig. 1f). The heat transport associated with the Indonesian Throughflow interestingly shows two positive peaks during year 0. The first peak is a result of the weaker spring Wyrtki jet at the initial stage of the IOD as discussed by Yamagata et al. (1996) and R. Suzuki et al. (2004, personal communication). The weaker spring Wyrtki jet results in the lower sea level in the eastern Indian Ocean, which enlarges the pressure gradient between the eastern Indian Ocean and the western Pacific, and thus increases the Indonesian Throughflow heat transport.

The second peak, on the other hand, is associated with the peak phase of the IOD, where the easterly wind stress anomaly lowers the sea level in the eastern Indian Ocean. Then, the total heat content anomaly starts to decrease continuously after February of year +1. The largest contribution to this comes from the Indonesian Throughflow; the reduced heat transport is explained by the fact that the downwelling Kelvin wave, which plays an important role in the decay of the positive IOD, increases the sea level height in the eastern Indian Ocean (Rao and Yamagata 2004) and decreases the heat transport associated with the Indonesian Throughflow (cf. Wyrtki 1987). Furthermore, the surface heat flux anomaly becomes negative from February (year +1), while the meridional heat flux anomaly becomes negative from July (year +1). It is interesting to note that the negative (positive) IOD event is similar to an El Niño (La Niña) event in the Pacific. This is because the negative IOD event and the El Niño event export heat accumulated in the equatorial region, whereas the positive IOD event and a La Niña event import heat to the equatorial region from the off-equatorial region (cf. Wyrtki 1985; Philander and Hurlin 1988; Jin 1997). In fact, the correlation coefficient between the DMI and the heat content anomaly in the Tropics north of 15°S

![Composite diagrams of (a) the upper-ocean heat content, (b) the meridional heat flux across 15°S, (c) the heat flux in the Indonesian Throughflow, (d) the surface heat flux, and (e) the vertical heat flux at 440-m depth for the biennial IOD event. Positive values signify that the tropical band of the Indian Ocean gains heat.](image-url)
reaches 0.79 for the whole 200-yr integration, when the former leads the latter by 3 months.

The above description is schematically summarized in Fig. 9. In year 0 of the positive IOD during the biennial case, the tropical band gains heat by 0.19 PW (Fig. 9a). The surface flux, the Indonesian Throughflow, and the meridional flux across 15°S contribute constructively to this tendency. However, the largest contribution comes from the meridional flux; heat loss to the south decreases by 0.12 PW. In contrast, the tropical band loses 0.22 PW of heat in year +1. Again, the surface flux, the Indonesian Throughflow, and the meridional flux across 15°S contribute to this constructively; the largest contribution comes from the decrease in the heat transport by the Indonesian Throughflow (−0.14 PW).

The above is a standard viewpoint for explaining the canonical IOD events with the biennial nature. Variations in various fluxes cause modulations in the recharge–discharge of heat and lead to variations in IOD occurrences. Since the total heat gain during the positive IOD event of year 130 is smaller compared with the case for the canonical positive IOD, another positive event occurs in year 131 (Fig. 9c). During the positive IOD event of year 130, the southward heat transport across 15°S of 1.60 PW is larger by 0.19 PW compared with the canonical positive IOD event. This value is larger than the annual mean value by 0.07 PW. Therefore, the heat gain is small even though the tropical band gains more heat from the surface heat flux (+0.07 PW) and the Indonesian Throughflow (+0.05 PW). We note here that the change in southward heat transport across 15°S is mostly due to the change in the zonal wind stress; the correlation coefficient between the zonal wind stress and the southward heat transport at 15°S is 0.63. Therefore, we suggest that the decadal variability in the strength of the Mascarene high in the southern Indian Ocean (cf. Allan et al. 1995; Reason et al. 1996a) could be the major cause of the modulation in the IOD occurrences. The variation may be associated with the atmospheric teleconnection from the tropical Pacific (see Fig. 2 in Luo et al. 2003) and/or the Anti-
artic Circumpolar Wave phenomenon (Venegas 2003). Work in this direction is in progress.

Since year 166 corresponds to a La Niña year, the sea level in the western tropical Pacific is anomalously high. Therefore, the heat transport of the Indonesian Throughflow is larger when compared with the annual mean by 0.16 PW (Fig. 9d). The large (small) Indonesian Throughflow during La Niña (El Niño) is in accord with observational studies (Meyers 1996; Bray et al. 1996). As a result, the tropical band loses heat only by 0.01 PW, whose value is much smaller compared with that of the canonical negative event (−0.22 PW), and another negative event occurs in year 167. Therefore, ENSO influences the heat budget of the tropical band in the Indian Ocean through variations of the Indonesian Throughflow and, thus, the occurrence of IOD events. This explains why some coarse-resolution coupled GCMs with large openings in the Indonesian seaways simulate a positive IOD event after the mature phase of El Niño in the Pacific (cf. Cai et al. 2005). In addition, ENSO may also influence the Indian Ocean via changes in the surface heat flux (Klein et al. 1999) and Ekman pumping (Xie et al. 2002). Since the amplitude and frequency of ENSO changes decadally (Barnett 1991; Garreaud and Battisti 1999; Tozuka and Yamagata 2003), the effect of ENSO on the IOD may also vary decadally.

4. Conclusions

Using the output from a 200-yr integration of the SINTEX-F1 ocean–atmosphere coupled model, we have investigated the decadal variation in the tropical Indian Ocean.

The model captures the interannual IOD events locked to a seasonal cycle quite realistically. It also captures the lower-frequency variations. The first EOF mode of the bandpass- (9–35 yr) filtered SSTA represents a basinwide warming (or cooling) mode and explains 37% of the total variance in the decadal band. Its principal component is correlated well with the decadal Niño-3 SSTA index, suggesting the close connection with the Pacific ENSO-like decadal variability. The second EOF mode explains 14% of the variance in the decadal band and shows a clear east–west SSTA dipole pattern. Although the spatial structure resembles that of the interannual IOD, its time scale is much longer. Based on the examination of DMI and its wavelet analysis, we suggest that the decadal IOD-like variability is an artifact of the linear statistical analysis. It should be interpreted more precisely as a decadal modulation of interannual IOD events (i.e., asymmetric or skewed occurrence of positive and negative events).

Using heat budget analysis, we have demonstrated that four elements modulate the IOD occurrence. In particular, we have shown that variations of the southward heat transport associated with changes in the Mascarene high are important. Variations of the Indonesian Throughflow sometimes associated with ENSO also play a dominant role in generating the modulation. In addition, changes in the surface heat flux associated with the atmospheric condition (possibly related to ENSO/monsoon variability), the Indonesian Throughflow transport associated with changes in the Southern Hemisphere wind field (Reason et al. 1996b) and subsurface vertical heat transport associated with the cross equatorial cell affect the heat budget of the tropical Indian Ocean and, thus, the skewness in the IOD occurrence.

It will be interesting to test our hypothesis using paleoclimate data. Based on CGCM simulations, Liu et al. (2003) have suggested that the summer monsoon in the Indo-Pacific sector was stronger during the mid-Holocene. This appears to be due to the enhanced Northern Hemisphere solar radiation in the boreal summer associated with the change in the orbital forcing. The proxy data support this result (e.g., Staubwasser et al. 2002). According to the present result, a larger surface heat loss owing to an increase in evaporative cooling associated with the enhanced Indian summer monsoon may introduce a favorable condition for the positive IOD events. This is actually the case in Liu et al.’s (2003) CGCM simulation; the simulated seasonal SSTA shows a zonal SSTA dipole structure.

The above new view on the decadal variability in the Indian Ocean, that is, that the decadal IOD is a result of a decadal modulation of interannual IOD events, is obtained from analyzing the coupled GCM results; however, we need to be careful in applying this view to the real world. To check this hypothesis, more observational data need to be accumulated in the Indian Ocean.

Acknowledgments. We thank Prof. S. G. H. Philander and Drs. S. K. Behera, K. Ashok, and Mr. T. Miyasaka for stimulating discussions. Useful comments made by two reviewers, Drs. C. J. C. Reason and H. Spencer, helped us to improve our manuscript. We are indebted to Dr. R. Zhang for the data management of the SINTEX-F1 model. The SINTEX-F1 model was run on the Earth Simulator. Wavelet software was provided by C. Torrence and G. Compo, and is available online (http://paos.colorado.edu/research/wavelets/). The present research is supported by the twenty-first century COE grant for the Predictability of the Evolution and Variation of the Multi-scale Earth System: An In-
APPENDIX

Synthetic Dataset for the DMI

To examine whether the “decadal IOD” can be a result of the frequent occurrence of either positive or negative events, we have constructed a synthetic dataset for the DMI. First, we made a composite diagram for the 10 strongest positive IOD events, as the IOD is phase locked to the seasonal cycle. Then, the DMI is calculated for each month (DMI\textsuperscript{iod}) and is used to generate a 100-yr synthetic DMI time series following

\[
DMI(t) = DMI_{iod}(m) \quad \text{if} \quad 12(4k - 3) + 1 \leq t(n, m) \leq 12(4k - 3) + 12 \\
= -DMI_{iod}(m) \quad \text{if} \quad 12(4k - 1) + 1 \leq t(n, m) \leq 12(4k - 1) + 12 = 0.0 \quad \text{otherwise}
\]

Here, \(t\) is time (in month), \(m\) is month (1 \(\leq m \leq 12\)), \(n\) is year (1 \(\leq n \leq 100\)), and \(k\) is an integer (1 \(\leq k \leq 25\)). Furthermore, we allowed decadal modulation of these interannual IOD events as summarized in Table A1.

Figure A1 shows the time series of the synthetic DMI thus obtained for the raw and 9–35-yr bandpass-filtered data. The decadal signal seen during years 65–85 is associated with frequent occurrences of positive and negative interannual IOD events.

When wavelet analysis is applied to the synthetic DMI, statistically significant decadal peaks emerge for years 65–85 (Fig. A2). Similarly, synthetic SSTA, SSHA, or zonal wind stress anomaly fields are generated following the procedure used in creating the synthetic DMI. Then, the 9–35-yr bandpass filter is applied to the above anomaly fields. Images that are strikingly similar to those shown in Fig. 5 are obtained (figures not shown). Hence, the decadal IOD may be the result of a frequent occurrence of either positive or negative IOD events.

REFERENCES


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**TABLE A1. Modulations introduced into the synthetic DMI.**

<table>
<thead>
<tr>
<th>Years</th>
<th>Description of modulations</th>
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<td>68, 72</td>
<td>Positive IOD events occur instead of negative IOD events</td>
</tr>
<tr>
<td>74, 78</td>
<td>Negative IOD events occur instead of positive IOD events</td>
</tr>
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**Fig. A1.** Time series of the raw (solid line) and its bandpass-(9–35 yr) filtered (dashed line) synthetic DMI.

**Fig. A2.** Wavelet power spectrum (using the Morlet wavelet) of the synthetic DMI in Fig. A1. Shading represents the wavelet power at each period being normalized by the global wavelet spectrum. The thick solid line contour encloses regions of greater than the 95% confidence level for a red-noise process.


