# The role of the Indian Ocean in climate forecasting with a particular emphasis on summer conditions in East Asia


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## Abstract

The Indian Ocean Dipole (IOD) is a natural ocean-atmosphere coupled mode that plays important roles in seasonal and interannual climate variations. In the present article, we describe the IOD event locked to a seasonal cycle and then discuss its close relation with the Asian summer climate variations. In particular, we demonstrate that the extremely hot and dry summer condition in 1994 was due to the positive IOD event.


## 1. Introduction

It is well known that the summer climate condition over East Asia is dominated by activities of the East Asian summer monsoon system. Since the East Asian summer monsoon system is one subsystem of the Asian Monsoon (Wang and Fan, 1999), it interacts with another subsystem, the Indian summer monsoon, via variations of the Tibetan high and the Asian jet (Rodwell and Hoskins, 1996; Enomoto et al., 2002).

Inspired by the anomalous summer conditions in East Asia during 1994 (e.g. Behera et al., 1999; Vinayachandran et al., 1999), Saji et al. (1999) discovered the existence of an east-west SST dipole in the tropical Indian. They showed that this dipole is coupled to zonal wind anomalies in the central Indian Ocean, suggesting the Bjerknes-type of air-sea interaction as in the tropical Pacific. The term Indian Ocean Dipole (IOD) was introduced to denote this basin-wide ocean-atmosphere coupled mode in the Indian Ocean. The positive IOD event is characterized by the strong positive Sea Surface Temperature Anomalies (SSTA) in the tropical western Indian Ocean (50ºE-70ºE, 10ºS-10ºN, denoted as Box A) and the negative SSTA in the southeastern Indian Ocean (90ºE- 110ºE, 10ºS- Eq., denoted as Box B). Thus the Indian Ocean Dipole Mode Index (IODMI) is defined as the zonal difference of SST anomaly of Box A (western pole) from that of Box B (eastern pole).

After introducing the IODMI, Saji et al.(1999) identified six major positive IOD events during the period of 1958 -1999. The composite pictures of those six events (1961, 1967, 1972, 1982, 1994, and 1997) demonstrated that the air-sea coupled IOD event evolves during boreal spring, matures in fall and decays in winter (cf. Figs 2 and 3 in Saji et al., 1999). The dipole pattern related to IOD is identified in heat content/sea level anomalies (Rao et al., 2002a), OLR anomalies (Behera et al., 1999, 2002; Yamagata et al., 2002; Saji and Yamagata, 2002a) and sea level pressure anomalies (Behera and Yamagata, 2002). We show all those indices in Fig.1. Several other authors also discussed this Indian Ocean coupled phenomenon (Webster et al., 1999, Murtugudde et al., 2000, Iizuka et al., 2000; Vinayachandran et al., 2002; Feng et al., 2002; Ashok et al., 2002; Gualdi et al., 2002) using observed data and/or model simulations.

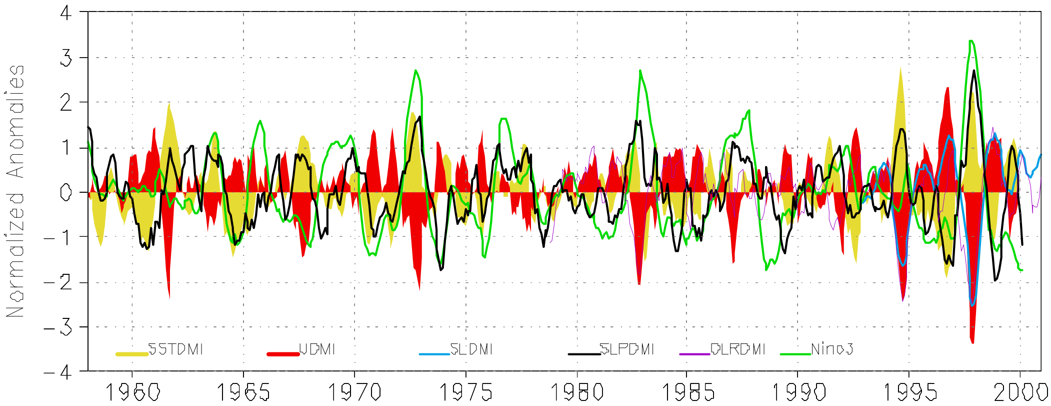


*Figure 1: Normalized indices of IOD, based on the anomalies of SST (SSTDMI) , zonal wind (UDMI), TOPEX/POSEIDON sea level (SLDMI), OLR (OLRDMI) and sea level pressure (SLPDMI). Niño-3 index from the eastern Pacific is shown for reference. SSTDMI is a difference between western (50ºE-70ºE, 10ºS-10ºN) and eastern (90ºE-120ºE, 10ºS-Eq) Indian Ocean. Similarly, SLPDMI is a difference between (96ºE-100ºE, 13ºS-9ºS) and (52ºE-56ºE, 9ºS-5ºS) and OLRDMI is a difference between (70ºE-80ºE, 5ºS-5ºN) and (90ºE-100ºE, 10ºS -Eq). The UDMI is obtained by taking area-average in the central equatorial region (70ºE-90ºE, 5ºS-5ºN). SLDMI is the sea level anomalies from the eastern box of the SSTDMI.*

We believe that the IOD has raised a new possibility to make a real advance in the predictability of seasonal and interannual climate variations originating in the tropics. Here we first describe that the IOD is a physical mode and then discuss its teleconnections.

## 2. IOD as a natural coupled mode in the tropical Indian Ocean

### 2.1. IOD's appearance in the statistical analyses

Despite the remarkable presence of IOD events, the dipole pattern appears as the second dominant mode in the SST anomalies in conventional statistical analysis such as the EOF method. It is rare for climate dynamists to discuss the second mode of variability. This is why some felt difficulties in understanding the new concept of the IOD. The dominant EOF mode is a basin-wide SST monopole that has a high correlation (~0.85) with the Nino-3 index, when the latter leads the former by 4 months. Thus, the dominant Indian Ocean SST variability is caused by the external forcing related to ENSO. The wavelet spectra of the SST anomalies in the eastern (10ºS-Eq., 90ºE-110ºE) and western (10ºS-10ºN, 50ºE-70ºE) poles show different behavior because of the masking effect of the dominant EOF mode (Fig. 2). However, we recover a remarkable coherence in the variability of the two boxes after removing the external ENSO effect (Fig 2 lower panels). This shows quite a contrast to other major oscillatory modes such as the Southern Oscillation and the north Atlantic Oscillation that appear as the first dominant modes; the statistical dominance allows a negative correlation between the poles for those two modes in the raw data. Since IOD is the second mode in SST variability, we need to remove the dominant mode to detect its sea-saw statistically as demonstrated in Behera et al. (2002). This subtle aspect of the tropical Indian Ocean variability as discovered by Saji et al. (1999) was missed unfortunately in earlier studies (e.g. Hastenrath et al., 1993).

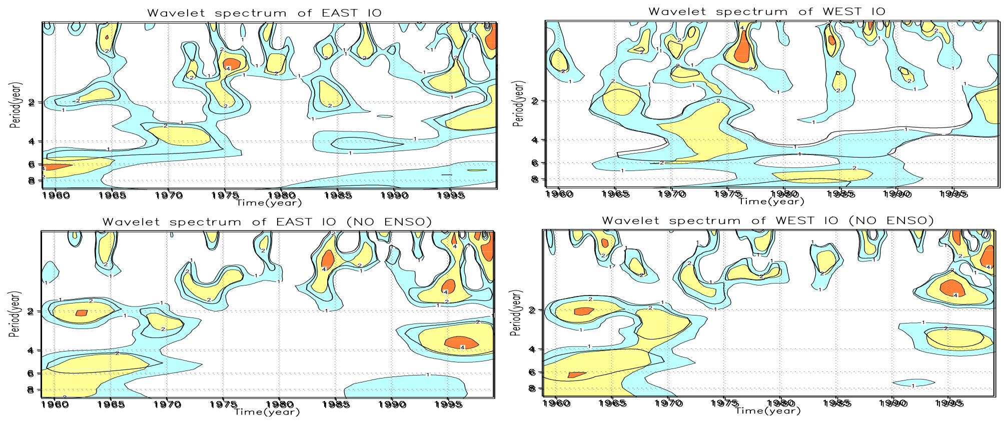


*Figure 2 Wavelet power spectrum (using the Morlet wavelet) of the SST anomalies (derived from GISST, Rayener et al. 1996) in eastern (left panels) and western (right panels) poles of the IOD. Upper two panels show the spectrum for the whole data and lower two panels show the spectrum when ENSO effect is removed from the data through a 4-month lagged regression of the Niño-3 index. Shaded is the wavelet power at each period being normalized by the global wavelet spectrum and the thick black contour is the 95% significance level*

## 2.2. IOD's relation with ENSO

The IOD evolution is locked to seasons. Thus it is important to introduce the seasonal stratification in the statistical analysis (cf. Nicholls and Drosdowsky, 2000; Allan et al., 2001). During the peak season (September-November) of the IOD, the first two dominant EOF modes (for the Indian Ocean north of 15ºS) show the east-west dipole patterns (figure not shown). Interestingly, the first EOF mode has a stronger correlation (~0.65) with the IOD as compared to its correlation with the Niño-3 (~ 0.58). The latter correlation corresponds to a similar correlation (~0.54) between IODMI and Niño-3 during this season; one is apt to conclude that IOD events occur as a part of ENSO events because of this high correlation (Allan et al., 2001; Baquero-Bernal et al., 2002). Rather, we claim that it reflects the fact that one third of the positive IOD events co-occur with El Niño events. The non-orthogonality of two time series does not necessarily mean that the two phenomena are always connected in a physical space.

We note that the second EOF mode, which also shows a dipole, has a significant correlation coefficient with IODMI (~0.69) but an insignificant value with the Niño-3 (~0.28). This difference in statistical correlation confirms the visual examination of the principal component that the second dipole mode is related to independent occurrences of IOD in certain years such as 1961, 1967 and 1994 (figure not shown).

| | Years of Positive IOD | Years of Negative IOD |
|---|---|---|
| 1 | **1961*** | **1958*** |
| 2 | 1963 | **1960*** |
| 3 | **1967*** | 1964 |
| 4 | 1972 | 1970 |
| 5 | **1977*** | **1989*** |
| 6 | 1982 | **1992*** |
| 7 | **1994*** | **1996*** |
| 8 | 1997 | - |

*Table 1 Years of IOD events. The asterisk denotes pure events, i.e. no El Niño (La Niña) during a positive (negative) IOD event.*

To investigate such a complex relation, Yamagata et al. (2002) have analyzed the Walker circulation that may connect the Indian Ocean with the Pacific through the atmospheric bridge. As 30% of the positive IOD co-occur with El Niño, a simple correlation analysis is misleading. Therefore, they used appropriate statistical tools like the composite technique and the partial correlation method to extract the distinct nature of the IOD. A positive (negative) IOD event is identified as a pure event when it is not accompanied simultaneously by El Niño (La Niña) (see Table 1). The presence of the anomalous Walker cell operating only in the Indian Ocean is clearly seen in the pure IOD composite (Fig. 3), thereby confirming the independent occurrence of the pure IOD. To avoid misunderstanding, we repeat that this analysis does not exclude the possibility that some IODs may be physically linked with some ENSO events.

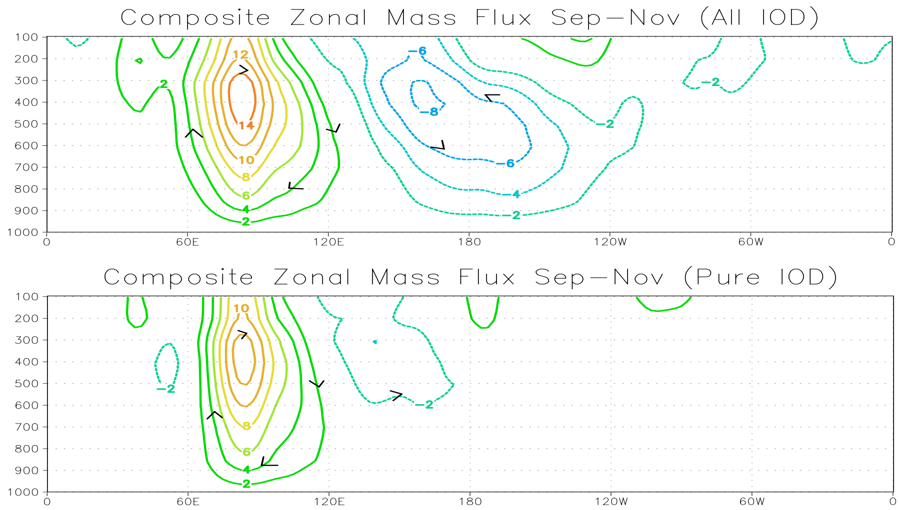


*Figure 3 September through November IOD composite ((positive events-negative events)/2) of zonal mass flux in the equatorial band (5°N-5°S) for pure IOD events (bottom). Contour interval is $4*10^9$ kg $s^{-1}$.*

## 3. IOD teleconnections in East Asia

The societal benefit of the IOD can be appreciated by analyzing its impact on the global climate system. Several recent studies (Ashok et al., 2001; Li and Mu, 2001; Behera and Yamagata 2002; Saji and Yamagata 2002b; Guan et al., 2002; Lareef et al., 2002) have shown IOD influences on many parts of the globe such as India, Australia, East Africa, and East Asia. However, we here focus our attention only on East Asia because of limitation of space.

The activities of the East Asian summer monsoon have profound economical and societal impacts on the East Asian countries. The anomalous changes in the summer monsoon circulation can lead to either abnormally hot (and dry) or cold (and humid) summer in this region. As mentioned in Introduction, East Asian countries suffered from the record-breaking hot and dry summer climate in 1994. Park and Schubert (1997) examined the nature of this year using some assimilated data from 1985 through 1994. Their conclusion is that “the anomalous circulation is primarily the result of an orographic forcing associated with zonal wind changes over Tibet”. However, we here show that the abnormal 1994 East Asian summer conditions are actually related to an ocean-atmosphere coupled signal in the tropical Indian Ocean, which is now called IOD.

Using the SST data (GISST2.3b) from 1979 through 1999 (Parker et al., 1995), we calculated the SSTA for June-July-August (JJA) and its standard deviation (σ) for three different tropical regions and the IODMI (Table 2, lower line). The IOD event in 1994 shows the variance that reached about 2.6 , indicating a very strong positive IOD event in the summer of 1994. We also note that the NINO3 region (5ºS-5ºN, 150ºW-

90ºW) showed the weak negative SST anomaly during the same period despite the negative Southern Oscillation Index (cf. Behera and Yamagata, 2002).

| Regions | IODMI | Box A | Box B | NINO3 |
|---|---|---|---|---|
| SSTA | 0.90 | 0.24 | -0.65 | -0.21 |
| σ | 0.35 | 0.32 | 0.31 | 0.85 |

*Table 2 JJA mean SSTA (1994) and its standard deviation for different tropical regions*

The Indian summer monsoon is expected to be significantly influenced by the IOD. Using all Indian rainfall data derived from the *in situ* observations (Parthasarathy et al., 1995), we actually found that India received good monsoon rainfall during June-July-August of 1994; it amounts to 265mm per month, which is 19% above the mean climatological value. This is consistent with our earlier study using both the observational data and an atmospheric general circulation model (AGCM) which suggests that the well-known negative correlation between Indian summer monsoon rainfall and El Niño can be interfered by the IOD during some decades (Ashok, et al., 2001).

The Indian summer monsoon system interacts with the tropical Indian Ocean. The East Asian summer monsoon interacts with the Indian summer monsoon via the tropospheric jets, Tibetan high, and even the westerly jet stream at about 40ºN in the upper troposphere (e.g. Lau and Li, 1984; Liang and Wang, 1998; Wang and Fan, 1999; Wang et al, 2001; Enomoto et al, 2002; Lu et al, 2002). When the circulation over South Asia changes anomalously, it is reasonable to expect that the summer monsoon circulation over East Asia will also change anomalously. We here show using the reanalysis data how the atmospheric circulation was influenced by the IOD during the summer in 1994.

## 3.1. Anomalous circulation features

Using the NCEP/NCAR reanalysis data (Kalnay, et al., 1996) from 1979 through 2001 and the CMAP precipitation data from 1979 through 1999 (Xie and Arkin, 1996), we have plotted the circulation anomalies during the summer months (JJA) of 1994 (Figs. 4 and 5). Large positive air temperature anomalies are found over the northeastern and eastern China, Korea, and Japan in 1994 summer (Fig. 4a). Some positive anomalies are also found above the Kuroshio Extension in the Northwestern Pacific. The anomalies of thickness between 200hPa and 850hPa isobaric surfaces are also positive (not shown), indicating the temperature of the air column is anomalously high. Those regions are associated with the strong negative precipitation anomalies (Fig. 4b). The water vapor anomalously diverges from this region, leading to a severe drought condition. These results agree well with those in Park and Schubert (1997). It is known that this northeastern part of Asia was covered during the summer of 1994 by an anomalous anticyclonic circulation in the lower troposphere. This anomalous circulation is found in the upper troposphere over this region (Fig. 5a), showing its equivalent barotropic structure. On the other hand, we find an anomalous cyclonic circulation elongating westward from the tropical western Pacific to the southern part of China (Fig. 4a); this circulation facilitates the surplus rainfall in this region (Fig. 4b). It prevents the moist monsoonal southerly wind from blowing northward from the Bay of Bengal and the South China Sea to the eastern part of China, Korea and Japan.

The above anomalous cyclonic circulation along with the intensified monsoon trough over India appears to be linked directly with the tropical IOD event. As seen in Fig. 4b, the distinctive 1994 IOD structure over the tropical Indian Ocean is manifested in rainfall anomalies and also in the velocity potential field (Fig. 5b). The water vapor converges into the western Indian Ocean (Fig. 4b), while it diverges in the southeastern Indian Ocean. An anomalous meridional circulation associated with the IOD connects the anomalous decent branch over the southeastern Indian Ocean and the anomalous ascent branch at about 20ºN, as simulated by

Ashok et al., (2001). More precisely, the anomalous northwestward low-level winds from the eastern pole of IOD reaches the Peninsula of India and then turns eastward (Fig. 4a). Since just the opposite winds are seen in the upper troposphere (Fig. 5a,b), the wind field in the tropics has a baroclinic structure. These results are in agreement with other results obtained from both data analysis and AGCM studies (Behera, et al., 1999; Ashok, et al., 2001; Guan et al., 2002).

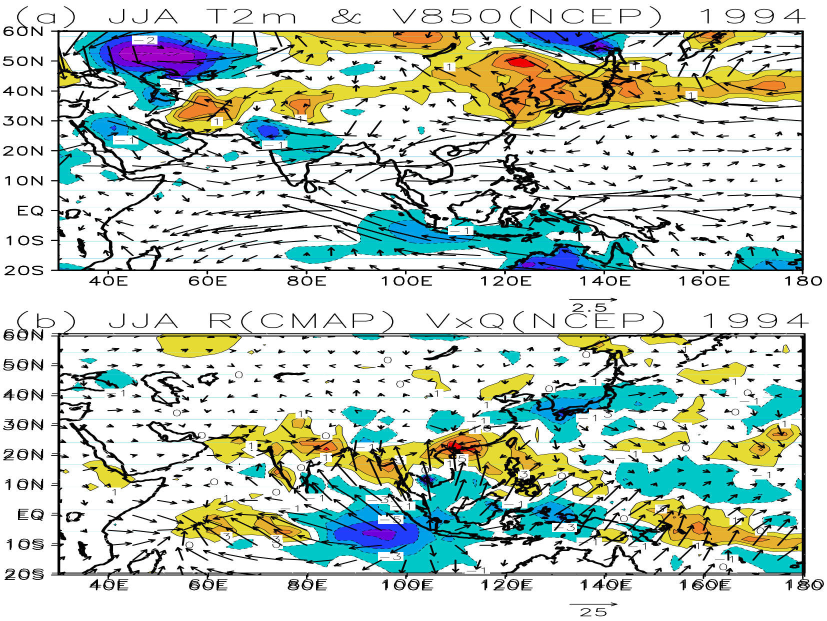


*Figure 4 (a)The JJA mean anomalous air temperature at 2m above the earth surface (in °C) along with the wind at 850hPa (in $m{\cdot}s^{-1}$) during 1994. (b) The anomalous precipitation (in $mm{\cdot}d^{-1}$) and the anomalous water vapor flux (in $Kg{\cdot}m^{-1}{\cdot}s^{-1}$) which is vertically integrated from the earth surface up to 300hPa (shown with vectors).*

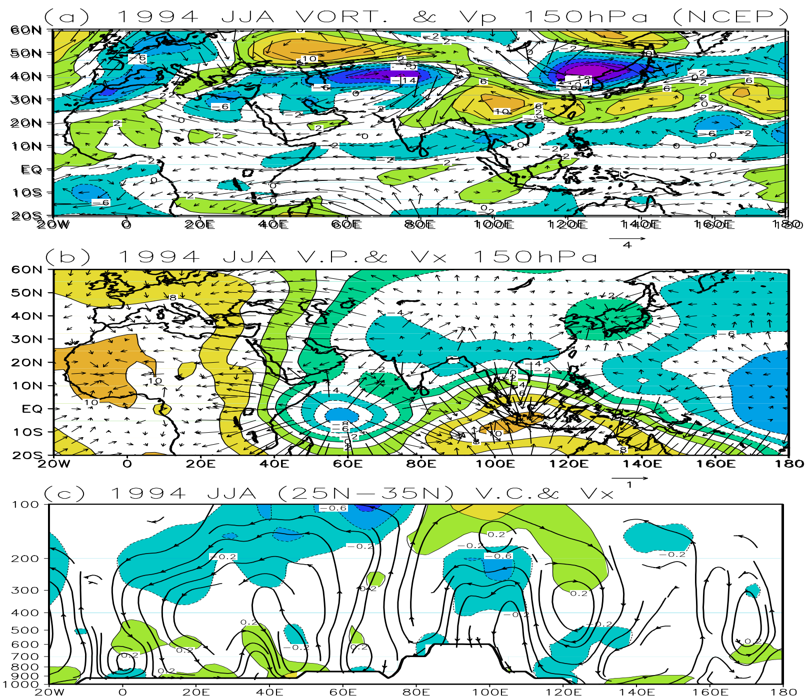


*Figure 5(a) JJA mean anomalous vorticity (to be multiplied by $1{\times}10^{-6}\ s^{-1}$) along with the rotational wind ($m{\cdot}s^{-1}$) at 150hPa in 1994. (b) JJA mean velocity potential along with the divergent wind ($m{\cdot}s^{-1}$) at 150hPa in 1994. The contour interval is $4{\times}10^{-5}m^{2}{\cdot}s^{-1}$. (c) JJA mean zonal-vertical circulation averaged over (25°N-25°N). The contours denote the zonal component of the divergent wind with contour interval of $0.2m{\cdot}s^{-1}$.*

## 3.2. Teleconnection mechanisms

The precipitation over India and the southern part of China is enhanced during the positive IOD event (Saji and Yamagata, 2002b). The northward branch of the meridional circulation excited by the eastern pole of the positive IOD leads to the anomalous updraft and the associated divergent flow in the upper troposphere over the Tibetan Plateau (Fig. 5b). As discussed by Sardeshmukh and Hoskins (1988) using a simple model, we observe the anticyclonic circulation at 150 hPa west of the vorticity source region, i.e., the Tibetan Plateau (Fig. 5a). A cyclonic circulation is simultaneously generated east of the vorticity source region. A Rossby wave train is also excited, propagating northeastward from the monsoon region.

The IOD-induced divergent flow in the upper troposphere near India also progresses westward and converges over Mediterranean/Sahara region (Fig. 5b). The zonal section averaged between 25°N and 35°N at 150 hPa captures the vertical circulation (Fig. 5c); the anomalous convection over India, which is induced by the IOD SSTA as explained, is amazingly linked to the anomalous decent in the Mediterranean/Sahara region, as discussed by Rodwell and Hoskins (1996) in a somewhat different context.

To examine mechanisms behind the above circulation changes in more detail, we show in Fig .6 the heat budget anomalies. Over the northern as well as the eastern part of China, the anomalous diabatic heating is dominant in the thermodynamic equation (Fig. 6c). Over Japan and Korea, the dynamic heating due to the anomalous descent of air is dominant, which cancels the anomalous negative horizontal advection of temperature. Around the Sea of Okhotsk, however, the anomalous positive horizontal advection of temperature balances the dynamic cooling (Fig. 6b). These differences imply, from the viewpoint of the heat budget, that there are different mechanisms of the hot summer over land and sea. Furthermore, these results indicate that the strong positive SSTA around Japan in 1994 (not shown) is not the cause of the hot summer. Rather, it is the result of the hot and dry summer condition.

Over India and the Bay of Bengal, the net anomalous diabatic heating is found (Fig. 6c), which is balanced by the negative anomalies of dynamic cooling due to the anomalous upward motion (Fig. 6b). On the other hand, the net diabatic cooling is found over the Mediterranean Sea/Sahara region (Fig. 6c). The negative anomalies of the horizontal advection of temperature are also found in this region (Fig. 6a). The anomalous dynamic heating due to the decent of air compensates both the diabatic and dynamic cooling.

Based on this heat budget diagnosis along with the vertical circulation shown in Fig. 5c, the relationship between the IOD/Monsoon and the anomalous circulation changes over the Mediterranean Sea/Sahara region can be established. The present view confirms the monsoon-desert mechanism put forward by Rodwell and Hoskins (1996); they suggested that the diabatic heating due to convective activities in the Indian region could induce an anticyclonic Rossby wave pattern that covers west Asia and northern part of Africa. The adiabatic decent thus induced by the remote thermal forcing from the Asian summer monsoon may intensify the decent induced by radiative cooling over the Mediterranean Sea/Sahara region.

The IOD-induced dynamic warming due to the decent of the air over the Mediterranean Sea/Sahara region and its vicinity must steadily perturb the mid-latitude westerly. Since the mid-latitude westerly acts as a Rossby wave-guide (Hoskins and Ambrizzi, 1993), the wave energy could propagate along the westerly eastward to East Asia, resulting in the summer circulation variations around East Asia and the Western Pacific.

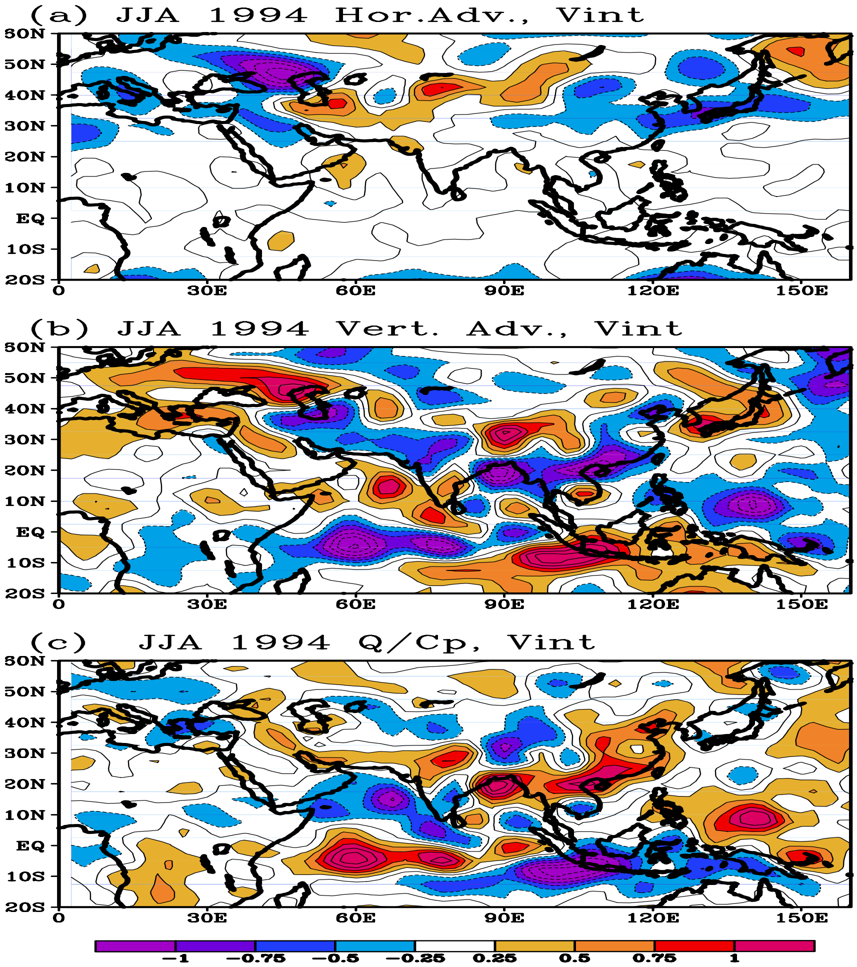


*Figure 6 JJA mean vertically integrated quantities for 1994. (a) The anomalous horizontal advection of temperature, (b) the anomalous vertical advection of potential temperature, and (c) the anomalous diabatic heating rate. All these quantities are vertically averaged over pressure from surface to 100hPa. The unit is °C·d* $^{-1}$.

This scenario can be examined by calculating the wave activity flux (WAF) (Plumb, 1986; Takaya and Nakamura, 2001). Fig. 7a clearly shows that the wave activity flux at 200hPa are much larger along the Asian westerly jet than those over other regions. The longitude-height cross-section (Fig. 7b) shows that the anomalous wave energy propagates upward into the upper troposphere around the regions of the Mediterranean Sea, the Caspian Sea, and the East Asia along the westerly jet stream. To the north of the Asian jet, the wave propagation is very weak; this suggests that the 1994 East Asian summer climate is not directly related to variations in higher latitudes. The upward propagating wave energy in the eastern flank of the Tibetan Plateau suggests that orographic forcing also plays an important role in 1994, as suggested by Park and Schubert (1997).

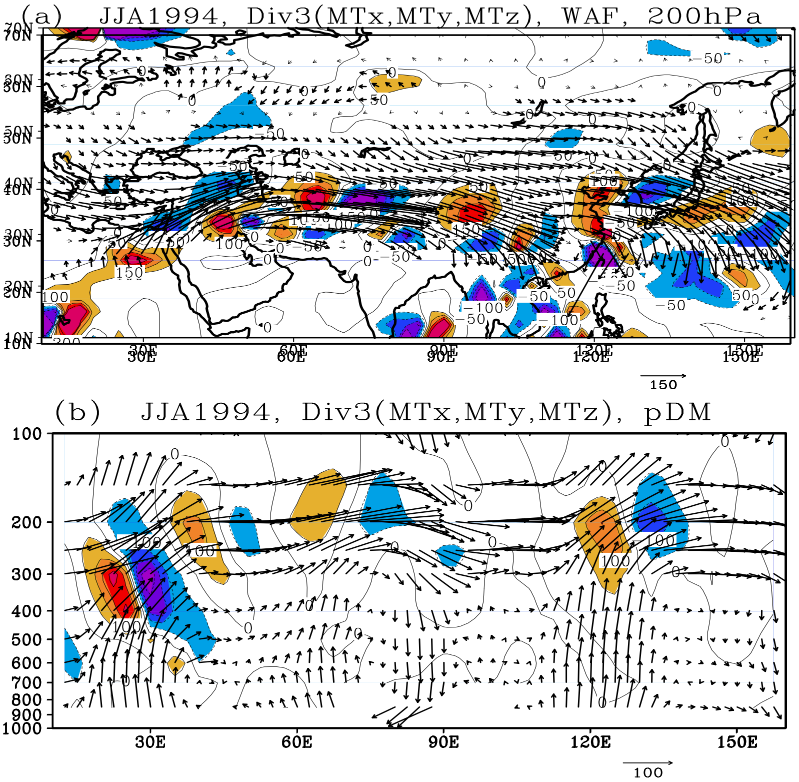


*Figure 7 (a)The wave-activity flux* $\mathbf{M}_T$ *(in* $m^2 \cdot s^{-2}$*) along with the 3-dimensional divergence* $\nabla_3 \mathbf{M}_T$ *(to be multiplied by* $1.0\times10^{-6}$ $m\cdot s^{-2}$*) at 200hPa. (b) The wave-activity flux(*$\mathbf{M}_{Tx}$*,* $M_{Tz}$ *) and the 3-dimensional divergence in zonal-vertical section. The high frequency components in the time-series have been removed by using a 5-day running mean. The wave-activity fluxes and their divergences in (b) have been averaged over (35ºN-45ºN). The vertical component of wave-activity flux* $M_{Tz}$ *is enlarged by 10 before plotting.*

## 4. Summary

Using various ocean and atmosphere data, we have demonstrated that the IOD is a natural ocean-atmosphere coupled mode in the Indian Ocean. Although the IOD emerges statistically as the second major mode in the SST anomalies, it shows up as a remarkable event in some years and induces climate variations in many places of the world. The year of 1994 is such a case and the dramatic impact on summer conditions in East Asia actually led authors to shed light on this important climate signal. We have discussed here how the IOD event influences summer conditions in East Asia.

The abnormally hot and dry summer in 1994 was associated with the anomalous anticyclonic circulation over Japan, Korea and the eastern and northeastern part of China. The anomalous cyclonic circulation over the southern part of China and the western Pacific weakened the monsoonal northward wind from the Bay of Bengal, the South China Sea, and the tropical Western Pacific, preventing the subtropical East Asia from receiving the normal water vapor from the tropical regions. The anomalously hot summer climate over East Asia is explained as a result of the anomalous dynamic heating around Japan, and diabatic heating over the northeastern and eastern part of China.

The IOD induced the summer circulation changes over East Asia in 1994 are at least in two ways. One is that a Rossby wave train is excited in the upper troposphere by the IOD-induced divergent flow in the upper troposphere over the Tibetan Plateau. The wave train propagates northeastward from the southern part of China. Another is that the IOD-induced diabatic heating around India excites a long Rossby wave pattern to the west of the heating. Through the monsoon-desert mechanism proposed by Rodwell and Hoskins (1996), the circulation changes over the Mediterranean Sea /Sahara region can be linked to the IOD/Monsoon variations. The westerly Asian jet acts as a waveguide for eastward propagating tropospheric disturbances to connect the circulation change around the Mediterranean Sea with the anomalous circulation changes over East Asia. This process may contribute to strengthening the equivalent barotropic structure in East Asia as suggested by Enomoto et al.(2002).

The study of teleconnections of IOD has just started. It is rather amazing that the monsoon-desert mechanism which plays a key role in understanding the hot and dry summer in East Asia was introduced by examining the summer of 1994 prior to the discovery of IOD (cf. Hoskins, 1996).

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## References

Allan, R., D. Chambers, W. Drosdowsky, H. Hendon, M. Latif, N. Nicholls, I. Smith, R. Stone,Y. Tourre (2001). Is there an Indian Ocean dipole, and is it independent of the El Niño - Southern Oscillation? *CLIVAR Exchanges* **6,** 18-22.

Ashok, K., Z. Guan, and T. Yamagata (2001). Impact of the Indian Ocean Dipole on the Decadal relationship between the Indian monsoon rainfall and ENSO, *Geophys. Res. Lett*. **28**, 4499-4502.

Ashok. K., Z. Guan, and T. Yamagata (2002). A further look at the ocean-atmosphere coupled mode in the Indian Ocean. *J. Met. Soc. Japan* (in press).

Baquero-Bernal, A., M. Latif, and S. Legutke (2002). On dipole-like variability in the tropical Indian Ocean. *J. Climate* **15**, 1358–1368.

Behera, S. K., R. Krishnan, and T. Yamagata (1999). Unusual ocean-atmosphere conditions in the tropical Indian Ocean during 1994. *Geophys. Res. Lett*. **26**, 3001-3004.

Behera, S. K., S.A. Rao, H.N. Saji, and T. Yamagata (2002). Comments on "A Cautionary Note on the Interpretation of EOFs". *J. Climate* (in press)

Behera, S. K., and T. Yamagata (2002). The Indian Ocean Dipole impact on the Southern Oscillation. *J. Met. Soc. Japan* (accepted).

Clarke, A. J., and X. Liu (1994). Interannual sea level in the Northern and Eastern Indian Ocean. *J. Phys. Oceanogr*. **24**, 1224–1235.

Enomoto, T., B.J.Hoskins, and Y. Matsuda (2002). The formation of the Bonin high in August. *Q. J. R. Meteor. Soc.(*in press).

Feng, M., G. Meyers, and S. Wijffels (2001). Interannual upper ocean variability in the tropical Indian Ocean. *Geophys. Res. Lett.* **28** , 4151-4154.

Gualdi, S., E. Guilyardi, A. Navarra, S. Masina, and P. Delecluse (2002). The interannual variability in the tropical Indian Ocean as simulated by a CGCM.. *Climate Dynamics* (in press).

Guan Z., K. Ashok, and T. Yamagata (2002a). Summer-time response of the tropical atmosphere to the Indian Ocean dipole sea surface temperature anomalies. *J. Meteor. Soc. Japan* (accepted).

Guan, Z., K. Ashok, and T. Yamagata (2002b). Influence of the Indian Ocean Dipole sea surface temperature on the East Asian summer climate. To be submitted.

Hastenrath, S, A. Nicklis, and L. Greischar (1993). Atmospheric-hydrospheric mechanisms of climate anomalies in the western equatorial Indian Ocean. *J. Geophys. Res* **98** (C11), 20 219–20 235.

Hastenrath, S. (2002). Dipoles, Temperature Gradient, and Tropical Climate Anomalies, *Bull. Amer. Met. Soc*. **83**, 735-738.

Hoskins, B. J. (1996). On the existence and strength of the summer subtropical anticyclones. *Bull. Amer. Met. Soc*. **77**, 1287-1292.

Hoskins, B. J. and T. Ambrizzi (1993). Rossby wave propagation on a realistic longitudinally varying flow. *J. Atmos. Sci*. **38**, 1179-1196.

Iizuka, S., T. Matsuura, and T. Yamagata (2000). The Indian Ocean SST dipole simulated in a coupled general circulation model. *Geophys. Res. Lett*. **27**, 3369-3372.

Kalnay, E., and coauthors, 1996: The NCEP/NCAR 40 year Reanalysis Project, *Bull. Amer. Meteor. Soc*. **77**, 437-471

Lareef, Z., S. A. Rao, T. Yamagata (2002). Modulation of Sri Lankan Maha rainfall by the Indian Ocean Dipole. *Geophys. Res. Lett*. (in revision).

Lau, K.-M. and M.-T. Li (1984). The monsoon of East Asia and its global associations-A survey. Bu*ll. Amer. Meteor. Soc*. **65**, 114-125.

Li, C., and M. Mu (2001). Influence of the Indian Ocean dipole on Asian monsoon circulation. *CLIVAR Exchanges*. **6**, 11–14.

Liang X.-Z. and W.-C Wang (1998). Associations between China monsoon rainfall and tropospheric jets. *Q. J. R. Meteor. Soc.* **124**, 2597-2623.

Lu, R.-Y., J.-H. OH, and B.-J. Kim (2002). A teleconnection pattern in upper-level meridional wind over the North African and Eurasian continent in summer. *Tellus* **54A**, 44-55.

Meyers, G. (1996). Variation of Indonesian throughflow and El Niño - Southern Oscillation. *J. Geophys. Res*. **101**, 12,255-12,263.

Murtugudde, R. G., J. P. McCreary, and A. J. Busalacchi (2000). Oceanic processes associated with anomalous events in the Indian Ocean with relevance to 1997-1998. *J. Geopys. Res*. **105**, 3295-3306.

Nicholls N., and W. Drosdowsky (2001). Is there an equatorial Indian Ocean SST dipole, independent of the El Nino-Southern Oscillation? *Preprints, Symp. on Climate Variability, the Oceans, and Societal Impacts*, Albuquerque, NM, Amer. Met. Soc. 17-18.

Park, C.-K., and S.D. Schubert (1997). On the nature of the 1994 East Asian summer drought. *J. Climate*.**10**, 1056-1070.

Parker, D.E., Folland,C.K. and M. Jackson (1995). Marine surface temperature: observed variations and data requirements. *Climate Change* **31**, 559-600.

Parthasarathy, B., A. A. Munot, and D. R. Kotawale (1995). Monthly and seasonal rainfall series for all-India homogeneous regions and Meteorological sub-divisions: 1871-1994. *Res.Rep. No. 65*, Indian Institute of Tropical Meteorology, Pune, India, 1995.

Plumb, R.A. (1986). Three-dimensional propagation of transient quasi-geostrophic eddies and its relationship with the eddy forcing of the time-mean flow. *J. Atmos. Sci*. **43**, 1657-1678.

Rao, S. A., S. K. Behera, Y. Masumoto, and T. Yamagata (2002a). Interannual variability in the subsurface tropical Indian Ocean with a special emphasis on the Indian Ocean Dipole. *Deep- Sea Res.* II, **49**, 1549-1572.

Rao, S. A, V. V. Gopalkrishnan, S. R. Shetye and T. Yamagata (2002b). Why were cool SST anomalies absent in the Bay of Bengal during the 1997 Indian Ocean Dipole Event? *Geophys. Res. Lett.,* **29**, 50.1, 50.4.

Rayner, N.A., E.B. Horton, D.E. Parker, C.K. Folland, and R.B. Hackett (1996). *Clim. Res. Tech. Note 74,* UK Met. Office, Bracknell.

Rodwell, M.J. and B.J. Hoskins (1996). Monsoons and the dynamics of deserts, *Q. J. R. Meteor. Soc*. **122**, 1385-1404.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata (1999). A dipole mode in the tropical Indian Ocean. *Nature* **401**, 360-363.

Saji, N. H., and T. Yamagata (2002a). Structure of SST and surface wind variability in COADS observations during IOD years. *J. Climate* (in revision).

Saji, N. H., and T. Yamagata (2002b). Interference of teleconnection patterns generated from the tropical Indian and Pacific Oceans. *Climate Res*. (in revision).

Sardeshmukh, P.D. and B.J. Hoskins (1988). The generation of global rotational flow by steady idealized tropical divergence. *J. Atmos.Sci*. **45**, 1228-1251.

Takaya, K. and H. Nakamura (2001). A formulation of a phase-independent wave-activity flux for stationary and migratory quasigeostrophic eddies on a zonally varying basic flow. *J. Atmos. Sci*. **58**, 608-627.

Vinayachandran, P.N., N.H. Saji, and T. Yamagata (1999). Response of the equatorial Indian Ocean to an anomalous wind event during 1994. *Geophys. Res. Lett*. **26**, 1613-1616.

Wang B. and Z. Fan (1999). Choice of South Asian summer monsoon indices, *Bull. Amer. Meteor. Soc*. **80**, 629-638.

Wang B., R. Wu, and K.-M. Lau (2001). Interannual variability of the Asian summer monsoon: contrasts between the Indian and the Western North Pacific-East Asian monsoons. *J. Climate* **14**, 4703-4090.

Webster, P.J., V.O. Magana, T.N. Palmer, J. Shukla, R.A. Tomas, M. Yanai and T. Yasunari (1998). Monsoons: Processes, predictability, and the prospects of prediction. J. Geophys. Res., 103(C7), 14451-14510.

Webster, P. J., A. Moore, J. Loschnigg, and M. Leban (1999). Coupled ocean-atmosphere dynamics in the Indian Ocean during 1997-98. *Nature* **40**, 356-360.

Xie, P. and P.A. Arkin (1996). Analyses of global monthly precipitation using gauge observations, satellite estimates and numerical model predictions, *J.Climate* **9**, 840-858.

Yamagata, T., S.K. Behera, S.A. Rao., Z. Guan, K. Ashok, and H.N. Saji (2002). The Indian Ocean dipole: a physical entity, *Clivar Exchanges* **24**, 1-6.