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Processes that Determine the Quasi-Biennial and Lower-Frequency Variability of the South Asian Monsoon

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Abstract

The spatial and temporal structures of atmospheric circulation and SST associated with the Indian monsoon rainfall variability on quasi-biennial (2–3 year) and lower-frequency (3–7 year) time scales were investigated, using the domain-averaged Indian rainfall, the NCAR/NCEP reanalysis and Reynolds SST data. We took both time-filtering and composite analysis approaches. The results indicate that physical processes that determine the monsoon rainfall variation on the 2–3 year and 3–7 year time scales are different. The quasi-biennial variability of the monsoon is primarily determined by local processes in the Indian Ocean. Both local SST and moisture flux convergence anomalies are highly correlated with the monsoon at a lagged time of 3–6 months. It is argued that a positive SST anomaly in the Indian Ocean increases local moisture due to enhanced surface evaporation. The accumulation of these moistures leads to a strong monsoon through anomalous moisture advection by summer mean flows.

The lower-frequency variability of the monsoon is primarily attributed to remote forcing mechanisms. Three possible processes may contribute to the monsoon variability on the 3–7 year time scale. The first is through the change of large-scale, east-west circulation induced by the eastern Pacific SST anomaly. The second is attributed to the effect of the SST anomaly in the Northwestern Pacific through enhanced (or suppressed) convective activity along the monsoon trough. The third is attributed to the tropical-midlatitude teleconnection—a strong north-south, land-ocean thermal contrast occurs six months prior to a wet monsoon, and it persists from the preceding winter to summer and is responsible for the monsoon intensity change.

1. Introduction

The power spectrum of the time series of the domain-averaged India rainfall reveals that on interannual time scales there are two distinctive peaks (Fig. 1). One is on the quasi-biennial time scale (2–3 year, hereafter referred to as the monsoon Tropospheric Biennial Oscillation (TBO) mode). The other has a longer time period (3–7 year, hereafter referred to as the monsoon Lower Frequency Variability (LFV) mode). The similar two power-spectrum peaks also appeared in the first EOF mode of the global precipitation field (Lau and Sheu 1988) and in the meridional wind component over the South China Sea and sea-level pressure difference between the Asian continent and Northwestern Pacific (Tomita and Yasunari 1996). Barnett (1991) investigated the spatial distribution and evolution of variability of near-global sea surface temperature (SST) and sea-level pressure (SLP) data in the quasi-biennial and lower-frequency period bands, and found that the near equatorial characteristics of the TBO mode are those of a quasi-progressive wave, while the characteristics of the LFV mode are those of a standing wave. However, Barnett's work did not specifically address the physical origin of quasi-biennial and lower-frequency variability of the Indian monsoon. Webster et al. (1998) investigated the temporal

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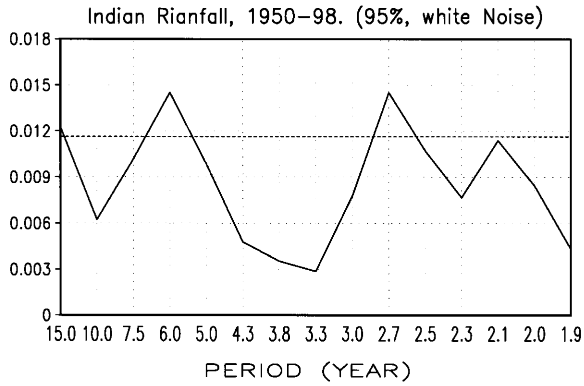

Fig. 1. Power spectrums of the domain-averaged Indian rainfall time series for a period of 1949–1998. The dashed line shows the 95% significance level.

characteristics of the Indian monsoon rainfall variability by use of a wavelet analysis and found the intermittent recurrence of the two power-spectrum peaks. Natural questions to be addressed are why the Indian monsoon has the two significant power-spectrum peaks, and what physical mechanisms are responsible for the monsoon variability on the two time scales.

El Niño-Southern Oscillation (ENSO) has been considered as a major factor that affects the Indian monsoon rainfall variability. Walker (1923, 1924) first recognized the effect of the Southern Oscillation on the Indian monsoon. Since then, a number of studies have been conducted to elucidate the monsoon-ENSO relationship (e.g., Yasunari 1990; Webster and Yang 1992; Ju and Slingo 1995; Lau and Yang 1996; see Webster et al. 1998 for a review). The Indian monsoon in general tends to have a simultaneous negative correlation with the eastern Pacific SST (e.g., Rasmusson and Carpenter 1983), although this simultaneous negative correlation has been broken in the recent two decades. Slingo and Annamalai (2000) hypothesized that the abnormal monsoon-ENSO relationship in 1997 resulted from the change of intensity and location of convection over the western Pacific. Kumar et al. (1999) suggested that it was attributed to the eastward shift of anomalous Walker circulation. Chang et al. (2001) proposed that the interdecadal change of snow cover over the Eurasia continent might disrupt the monsoon-ENSO relationship.

Several theories have been proposed to understand the origin of the TBO (see Li et al. 2001a for a review). Nicholls (1978) proposed that the TBO resulted from local air-sea interactions modulated by annual-cycle basic states. Meehl (1987) hypothesized that anomalous convection associated with a strong Indian monsoon propagated southeastward and changed the eastern Pacific SST in such a way that it suppressed the monsoon convection in the subsequent year. Meehl (1997) further proposed a teleconnection mechanism through which anomalous tropical heating affects the midlatitude atmospheric circulation and land temperatures. By excluding the tropical-midlatitude teleconnection, Chang and Li (2000) and Li et al. (2001b) demonstrated, in a simple coupled atmosphere-ocean model, that the TBO is an inherent monsoon mode, resulting from multi-region interactions among the Asian and Australian monsoons and adjacent tropical oceans.

Since the monsoon is an annual event while the typical life cycle of El Niño/La Niña events spans about two years, the ENSO-monsoon relationship may vary with the growing and decaying phase of ENSO. Many of the previous studies focused on the simultaneous relationship. It is practically more useful to study the lagged monsoon-ENSO relationship. Yang and Lau (1998) pointed out that while the summer SST anomaly (SSTA) had a direct impact through the attenuated Walker and local Hadley circulation, the SSTA in the preceding winter and spring might have an indirect impact through the change of land soil moistures. Kawamura (1998) suggested that enhanced convection over the maritime continent generated anomalous low-level cyclonic circulation as a result of the Rossby wave response, leading to increased rainfall and soil moisture, and thus a weak Indian monsoon in the following summer. There has been a controversy with regard to the lagged monsoon-ENSO relationship. Lau and Yang (1996) showed that prior to a strong Indian monsoon a positive SSTA appears in the preceding winter. Kawamura (1998), on the other hand, noted that a negative SSTA appears in the preceding winter prior to a strong monsoon. Given the fact that the TBO and LFV modes are dominant in different periods (Webster et al. 1998; An and Wang 2000), it is conceivable that the controversial results may arise from the different-period data sets they used for their analyses.

The objective of this study is to investigate the physical mechanisms responsible for the quasi-biennial and low-frequency variability of the Indian monsoon rainfall by revealing the spatial and temporal structures of atmospheric circulation and SST associated with the monsoon TBO and LFV modes. Our strategy is first to apply a time-filtering technique to separate rainfall data into 2–3 year and 3–7 year bands respectively. Then, by analyzing the spatial and temporal patterns of atmospheric circulation associated with the two bands, we intend to investigate physical processes responsible for the rainfall variability on the two time scales. To compare with results from the time-filtering analysis, we will also conduct a composite analysis using the original unfiltered data.

The rest of this paper is organized as follows. In section 2, we describe the data and the time-filtering technique. In sections 3 and 4, we present results from the time-filtering and the composite analyses, respectively. The roles of local and remote SST anomalies and the relative contribution of the TBO and LFV modes are discussed in section 5. In section 6 we summarize the results.

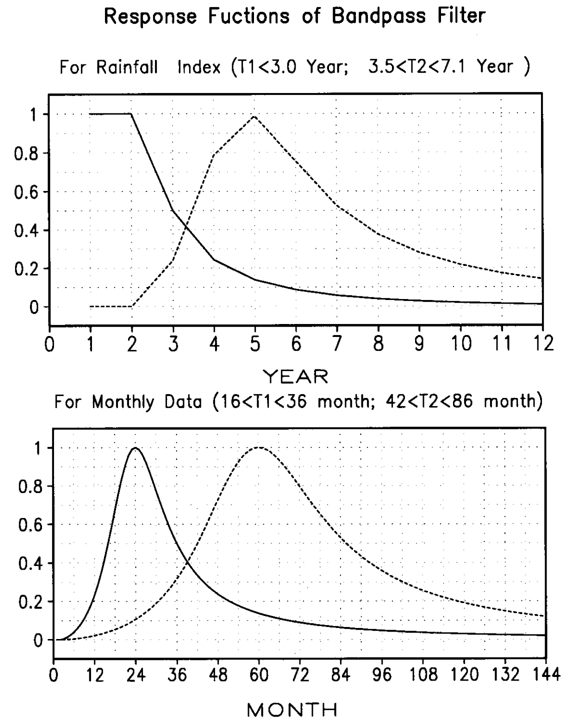

Fig. 2. Response functions of the band-passed filters for the seasonal mean (top panel) and monthly mean (bottom panel) data.

2. Data and methodology

2.1 Data

The primary data used in this study are the domain-averaged Indian rainfall, NCAR/NCEP reanalysis that includes wind, moisture, temperature and geopotential height fields, and the Reynolds SST (Reynolds and Smith 1994) for a period of 1949–1998. The Indian rainfall is represented by an area-averaged precipitation from 26 stations reasonably distributed over the Indian subcontinent. These gauge stations are picked up from the NOAA climatological baseline station data over land, documented at NCDC and updated from CAC global CEAS summary of day/month observations. This area-averaged rainfall has a correlation coefficient of 0.86 with the all-Indian-rainfall index (Mooley and Parthasarathy 1984).

2.2 Time-filtering technique

A band-pass filter (Murakami 1979) is used to separate the data into approximately 2–3 year and 3–7 year windows, respectively (see Fig. 2). These two bands represent the two significant power-spectrum peaks in the area-averaged Indian rainfall field. Since the rainfall data are seasonal means while other variables are monthly means, two types of filters are designed for seasonal and monthly mean data, respectively.

The amplitude response functions for these two filters are shown in Fig. 2. In Fig. 2a, the solid line indicates the response function of the high-pass filter with a half-power point at 3 year, while the dashed line indicates the band-pass filter with half-power points at 3.5 and 7.1 years, respectively, and a peak point at 5 years. This filter is specially designed for the Indian summer (JJAS) rainfall. Figure 2b shows the response functions for the monthly mean variables. It has half-power points at 16 and 36 months (solid line) for the 2–3 year window, and 42 and 86 months (dashed line) for the 3–7 year window. For both filters, the time frequencies of the peak-power point (denoted by W_0) and the two lateral half-power points (denoted by W_1, W_2) abide by the following relationship:

$$W_0^2 = W_1 \cdot W_2.$$

The resultant peak-power points for these two bands are 2 and 5 years, respectively. Thus, the band pass filters well separate the 2–3 year and 3–7 year windows. After the rainfall, SST and atmospheric variables are separated into two independent time series, they are normalized based on their standard deviations. A lagged correlation analysis is then performed for each data set. A composite analysis using the original unfiltered data is also carried out to crosscheck the lagged correlation analysis results obtained from the time-filtered data.

3. Results from the time-filtering analysis

3.1 SST and surface winds

Figure 3 shows the lagged correlation between the Indian summer rainfall and domain averaged SSTs in the Indian Ocean (IO, 60°–95°E and 0°–15°N), western Pacific (WP, 130°–150°E and 10°–20°N) and eastern Pacific (EP, 170°–120°W and 5°S–5°N). In the 2–3-year band (top panel), a significant positive correlation between the IO SSTA and the monsoon rainfall appears in the preceding winter and spring, with a maximum correlation coefficient exceeding 0.6, far above the 95% significance level. (The 95% level corresponds to a correlation coefficient of about 0.4 when taking into account the decrease of degrees of freedom due to the time filtering). The fact that a warm IO SSTA leads to a wet monsoon implies that the IO SSTA may play an active role in affecting the Indian monsoon.

The warm SSTA in the IO in the preceding winter may result from the decrease of the prevailing northerly associated with a weak Asian winter monsoon. It is seen from Fig. 4 that the meridional wind at the surface is positively, lagged correlated (with a maximum correlation coefficient greater than 0.4) with the summer monsoon. This implies that a southerly wind anomaly appears over the northern IO in the preceding winter. The southerly wind anomaly can contribute to the ocean warming through 1)

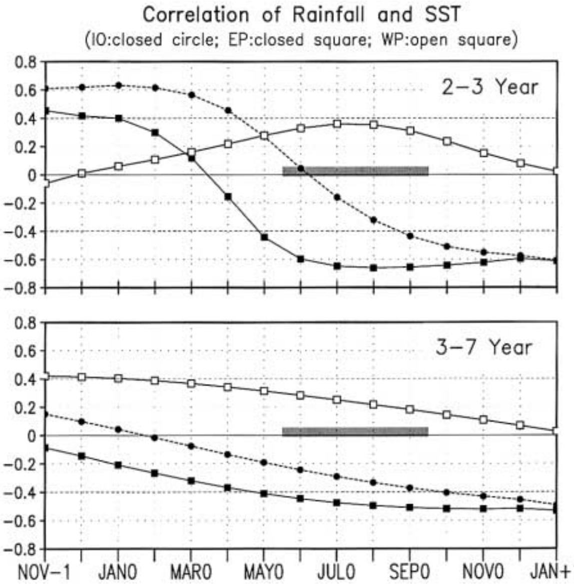

Fig. 3. Lagged correlation of the India monsoon rainfall with the SSTA in the central Indian Ocean (60–95°E and 0–15°N, circle), western Pacific (130–150°E and 10–20°N, open square) and eastern Pacific (120–170°W and 5°S–5°N, close square) for the 2–3 year (top panel) and 3–7 year (bottom panel) bands. The correlation above 0.4 corresponds to the 95% significance level or above.

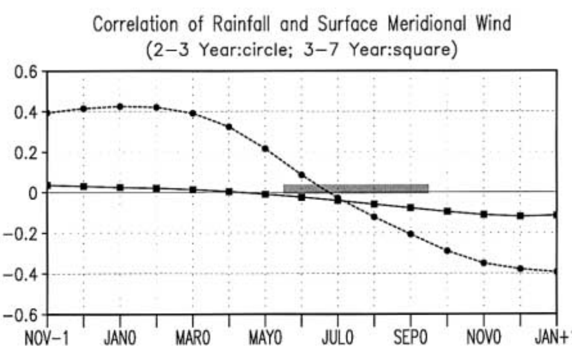

Fig. 4. Lagged correlation between the monsoon rainfall and the meridional wind at the surface in the central Indian Ocean (75–85°E, 0–12.5°N) for both 2–3 year (circle) and 3–7 year (square) bands.

reduced surface evaporation (because the mean wind is northerly in northern winter), and 2) anomalous meridional temperature advection that brings warmer water from the south. Thus, on the TBO time scale a strong Indian summer monsoon is preceded by a weak winter Asian monsoon[1] that is characterized by anomalous southerly over the northern IO.

Another notable feature associated with the monsoon TBO mode is the phase reversal of the EP SSTA in spring. The SST correlation coefficient changes its sign from a positive value in the preceding winter to a negative value in summer. Associated with this SSTA phase transition, the surface wind anomaly switches from westerly to easterly in the central equatorial Pacific.

In contrast to their high lagged correlation, the simultaneous correlation between the Indian monsoon and the IO SSTA is very low. This is because a strong Indian monsoon generates strong surface winds that further cool the ocean through enhanced surface evaporation and ocean mixing. As a result, the SSTA weakens rapidly in summer in the northern IO, especially along the eastern coast of Africa and in the Arabian Sea.

For the monsoon LFV mode, the most significant SST correlation appears in the Pacific. While the WP SSTA has a positive, lagged correlation (+0.4) with the monsoon in the preceding winter, the EP SSTA has a simultaneous negative correlation (−0.5). The correlation with the IO SSTA is weak for all seasons prior to the monsoon onset, whereas a negative correlation appears after the monsoon onset, indicating a strong monsoon impact on the IO SSTA.

A common feature in both the 2–3 year and 3–7 year bands is that there is a strong simultaneous, negative correlation between the Indian monsoon and the EP SSTA. This points out an interactive nature of the monsoon-ENSO system. On one hand, a positive (negative) SSTA may have a remote impact on the monsoon through large-scale vertical overturning (Meehl 1987). On the other hand, anomalous monsoon heating may alter the EP SST through the change of winds over the central and western Pacific (Chang and Li 2001).

3.2 Vapor transport and Walker circulation

To examine the effect of anomalous moisture transport, we composite the 1000 hPa moisture flux convergence field based on the NCEP/NCAR reanalysis data. Ten wettest and ten driest years are selected based on the filtered monsoon rainfall data for both bands. Form wet-minus-dry composites (Fig. 5), one can see that in the 2–3 year band there is significant low-level moisture convergence over the Indian subcontinent in the preceding winter and spring. In the 3–7 year band the anomalous moisture flux convergence is quite different—no significant moisture convergence appears in the Indian subcontinent.

The Walker circulation is regarded as an important agent that links the Pacific Ocean to the Asian monsoon. Since the vertical motion in the midtroposphere is related to upper and lower tropospheric divergent flows, we use the velocity potential difference (VPD) between 850 hPa and 200 hPa to represent the vertical overturning cell of the Walker circulation. A positive (negative) VPD center corresponds to a strong ascending (descending) motion. Figure 6 illustrates the composite VPD field for the 2–3 year and 3–7 year bands. A similar feature in both bands is that in summer (JJA), the Walker circulation is characterized by a strong ascending (descending) branch over the monsoon (EP) region. However, the evolution of the Walker cell in the two bands is quite different. For the LFV mode (right panel) the ascending and descending branches are almost stationary, whereas for the TBO mode (left panel) there is slow eastward propagation of the ascending and descending branches. This temporal evolution feature is somewhat similar to that found by Barnett (1991).

Another feature in the 2–3 year band is that even though there is a remote SSTA forcing in the EP in winter, an ascending branch appears in the equatorial IO. This ascending motion results from the direct impact of the warm SSTA in the IO, which compensates the effect of the El Niño in the EP. Thus, in addition to its moisture effect, the IO SSTA may have a dynamic impact on the vertical overturning of the Walker cell.

1 This result is consistent with Li (1990), Tomita and Yasunari (1996) and Zhang et al. (1997) who pointed out that there were simultaneous, negative correlations between the anomalous East Asian winter monsoon and El Niño.

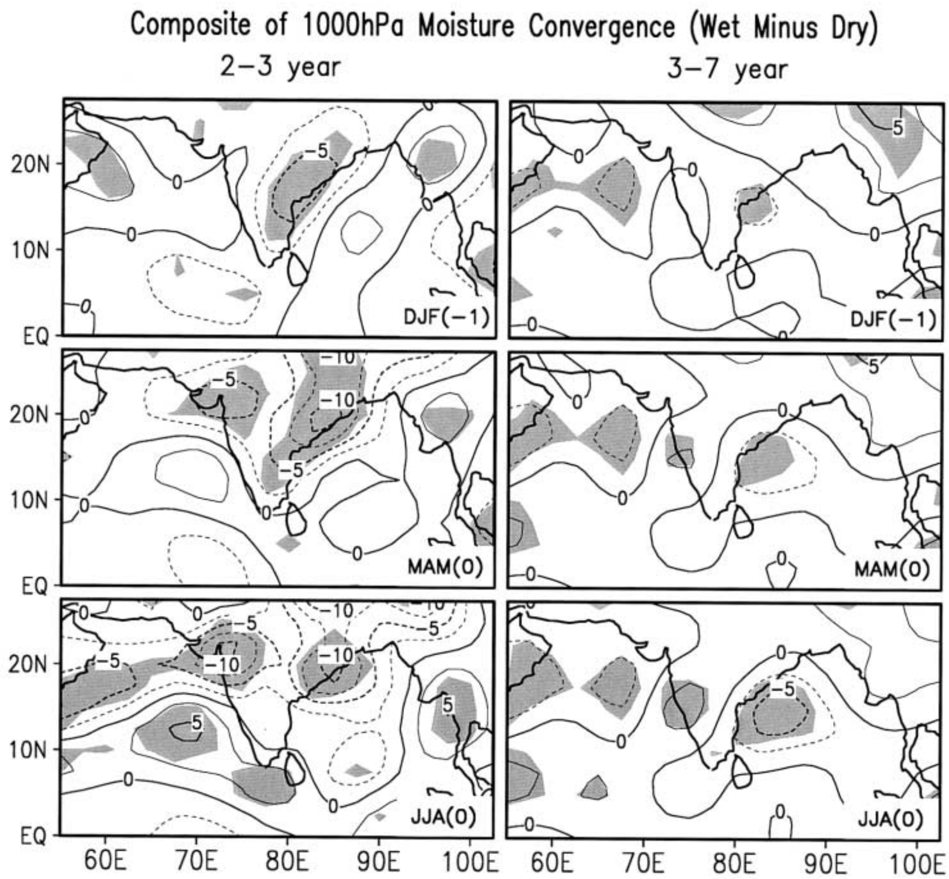

Fig. 5. Wet-minus-dry composites of 1000 hPa moisture flux convergence fields (units: 10^{-6} $gkg^{-1}s^{-1}$) in DJF (−1), MAM (0) and JJA (0) for the 2–3 year and 3–7 year bands. The composites are based on ten wettest and driest cases obtained from the time-filtered rainfall data. The shaded regions represent the statistical significance of 95% and above.

3.3 Land-ocean thermal contrast

The land-ocean thermal contrast between the Asian continent and Indian Ocean was regarded as a good indication of the monsoon strength (e.g., Li and Yanai 1996; Yang et al. 1996). To examine the role of the land-ocean thermal contrast on the monsoon variability, we calculated the lagged correlation between the monsoon rainfall and the mean (between 200 and 500 hPa) tropospheric temperature. Figure 7 shows the lagged correlation maps. At the 2–3 year window (left panel), the increase of the tropospheric mean temperature is concurrent with the warming of the SST in the IO in the preceding winter and spring, while no significant temperature changes are found over the Eurasian continent. At the 3–7 year window (right panel of Fig. 7), a significant warming of the tropospheric mean temperature appears over the subtropical Eurasian continent in the preceding winter, with the correlation coefficient greater than 0.7. Meanwhile, a negative correlation center is located over the western equatorial IO. This north-south thermal contrast is even enhanced in northern spring, and is significantly correlated with the summer monsoon rainfall. The physical processes that give rise to such a land-ocean thermal contrast are not clear, although several investigators (e.g., Meehl 1997; Yang and Lau 1998) have hypothesized that it might result from remote SST forcing in the tropics. The establishment of the meridional temperature gradient in the preceding season may help to set up the monsoon southwesterly earlier and stronger.

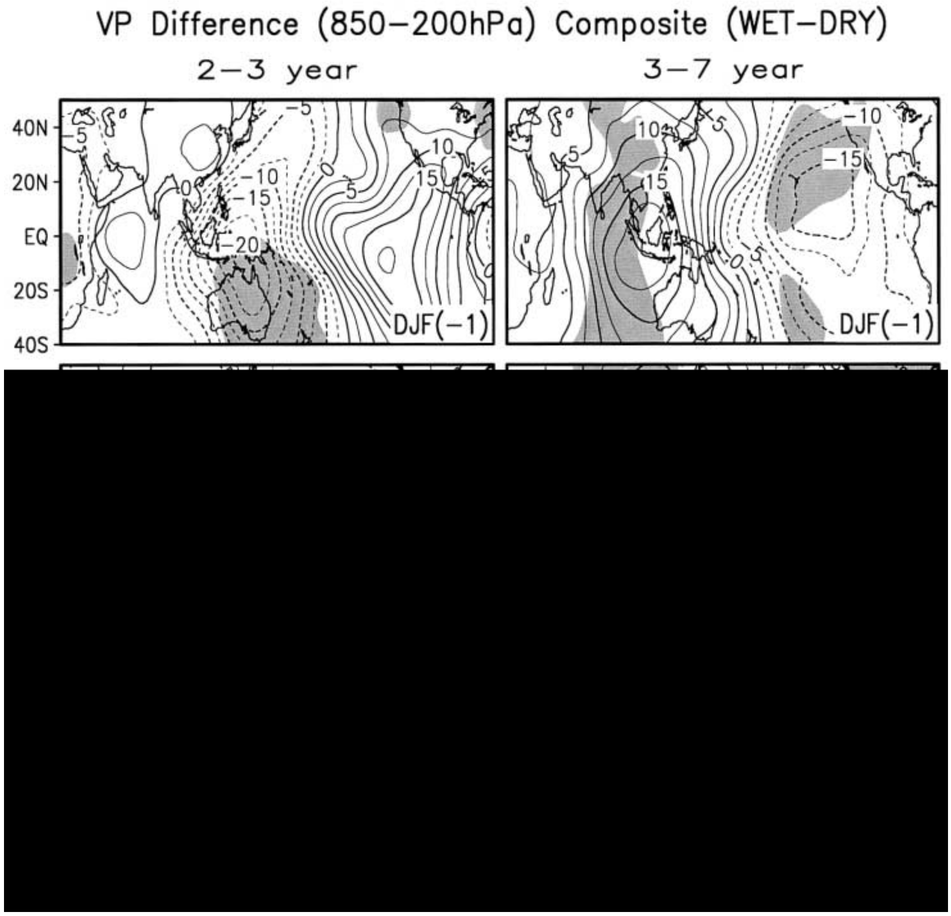

Fig. 6. Wet-minus-dry composites of the velocity potential difference (850 hPa minus 200 hPa) field (counter interval: 2.5×10^5 m^{2}s^{-1}) in DJF (−1), MAM (0), and JJA (0) for the 2–3 year and 3–7 year bands. The shaded regions represent the statistical significance of 95% and above.

4. Results from the composite analysis

The time filtering analysis above shows that the monsoon TBO and LFV modes have distinctive spatial and temporal structures. For the TBO mode a positive SSTA in the IO leads to a wet monsoon and the SSTA in the EP changes its phase in spring, whereas for the LFV mode a cold SSTA in the EP persists from winter to summer and is associated with a wet monsoon. To ensure that this result is not an artificial outcome of the band pass filters, here we further conduct a composite analysis using the original unfiltered data.

We partitioned the NINO3 SSTA into 3 groups: warm, cold, and normal, where the warm and cold categories are meant to represent El Niño and La Niña events and are defined by SSTA magnitude larger than 0.7 standard deviation for at least 2 months during September–February. The Indian summer rainfall anomaly is also partitioned into 3 categories: wet, dry, and normal, with the wet and dry categories defined by anomaly rainfall magnitude larger than 0.7 standard deviation. Based on this criterion, 12 El Niño and 14 La Niña cases are selected for 1949–98 (see Table 1). Different from previous studies that mainly focused on simultaneous monsoon-ENSO relationships, here we consider the lagged correlation between the peak phase of El Niño in winter and the rainfall anomaly in the following summer. Note that following 12 El Niño events there are 4 wet, 5 normal, and 3 dry monsoons, indicating that the probability of the monsoon can be wet, dry or normal following a peak El

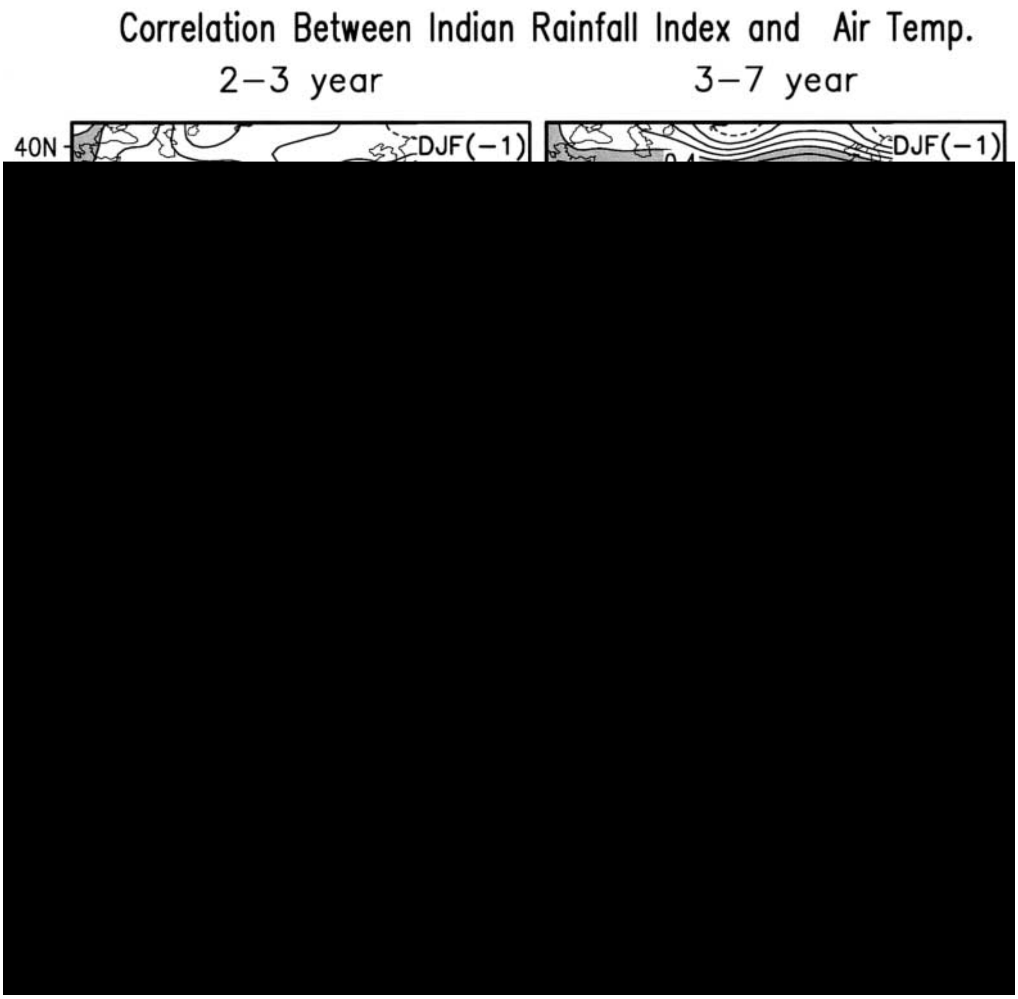

Fig. 7. Lagged correlation maps between the India monsoon rainfall and the mean tropospheric (200–500 hPa) temperature in DJF (−1), MAM (0) and JJA (0) for the 2–3 year and 3–7 year bands (contour interval: 0.1). The regions where the positive (negative) correlation exceeds 0.3 are heavily (lightly) shaded. The statistical significance exceeds the 95% level when the correlation is above 0.4.

Table 1. Years in which the Indian summer monsoon is wet, dry or normal following a peak El Niño or La Niña episode in the preceding winter

	El Niño (-1)	**La Niña (-1)**
Wet (0)	1964 1970 1983 1988	1956
Normal (0)	1958 1973 1977 1992 1995	1955 1957 1962 1971 1976 1989 1996
Dry (0)	1952 1966 1987	1963 1965 1968 1972 1974 1985

Niño episode. Following 14 La Niña episodes there are 1 wet, 7 normal, and 6 dry monsoons, suggesting greater probability towards wet or normal monsoons after a peak La Niña episode.

Figure 8 shows the time evolution of the composite SSTA from the preceding winter to summer following a peak El Niño. Although the EP SSTA is positive for all cases, the monsoon rainfall anomaly can be wet, dry or normal. A large difference exists in the IO between the wet and dry cases. The IO SSTA is mostly warm before a wet monsoon, and mostly cold before a dry monsoon. This situation in which

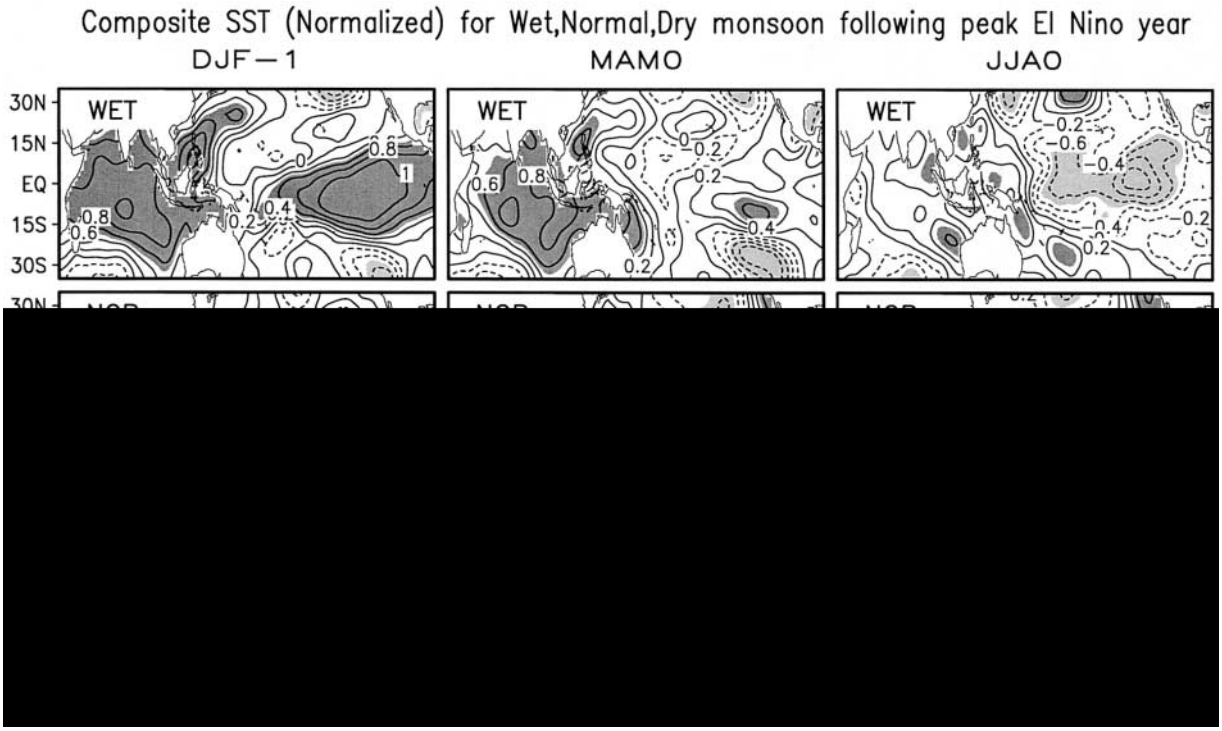

Fig. 8. Composites of anomalous SST fields in DJF (−1), MAM (0) and JJA (0) in association with wet, normal and dry Indian summer monsoons following a peak El Niño episode in northern winter. The SST fields are normalized with respect to their standard deviations at each grid point. The contour interval is 0.2.

the preceding IO SST is a better predictor than the EP SSTA for the Indian summer rainfall also appears in the northern spring composites. The relationship of the preceding IO SSTA with the subsequent Indian monsoon rainfall verifies the results from the time filtering analysis.

Another notable feature in Fig. 8 is the time evolution of the SSTA in the EP. For the wet cases, the composite warm SSTA in the EP decays rapidly in spring and changes its sign in summer, while for the dry cases the warm SSTA persists from the preceding winter to summer. The former resembles a typical life cycle of the TBO, while the latter represents low-frequency variability. Thus, in addition to the IO SSTA, the phase transition of the SSTA in the EP is another precursory signal for a wet Indian monsoon after a peak El Niño. Note that in this composite analysis we did not preseparate the biennial and lower-frequency time scales. Nevertheless, as the monsoon-ENSO relationship concerned, the two modes are naturally separated.

The composite VPD fields (figure omitted) show that in the wet composite an ascending branch persists over the IO from the preceding winter to summer. Another ascending branch appears in the EP in winter in response to the local SSTA there. Due to the phase transition of the SSTA in the EP, this ascending branch decays quickly in spring and becomes even a descending branch in summer. In the dry composite, an ascending branch persists in the EP, while the Indian subcontinent is under the control of the descending motion.

For the La Niño cases, the structure and evolution of the SSTA and VPD composites resemble a mirror image of the El Niño cases. While the EP SSTA changes its sign from a negative to a positive value in spring in the dry composite, the SSTA persists from the preceding winter to summer in the wet composite (Fig. 9).

The composite analysis with the original unfiltered data leads to the following conclusions that are consistent with the time-filtering

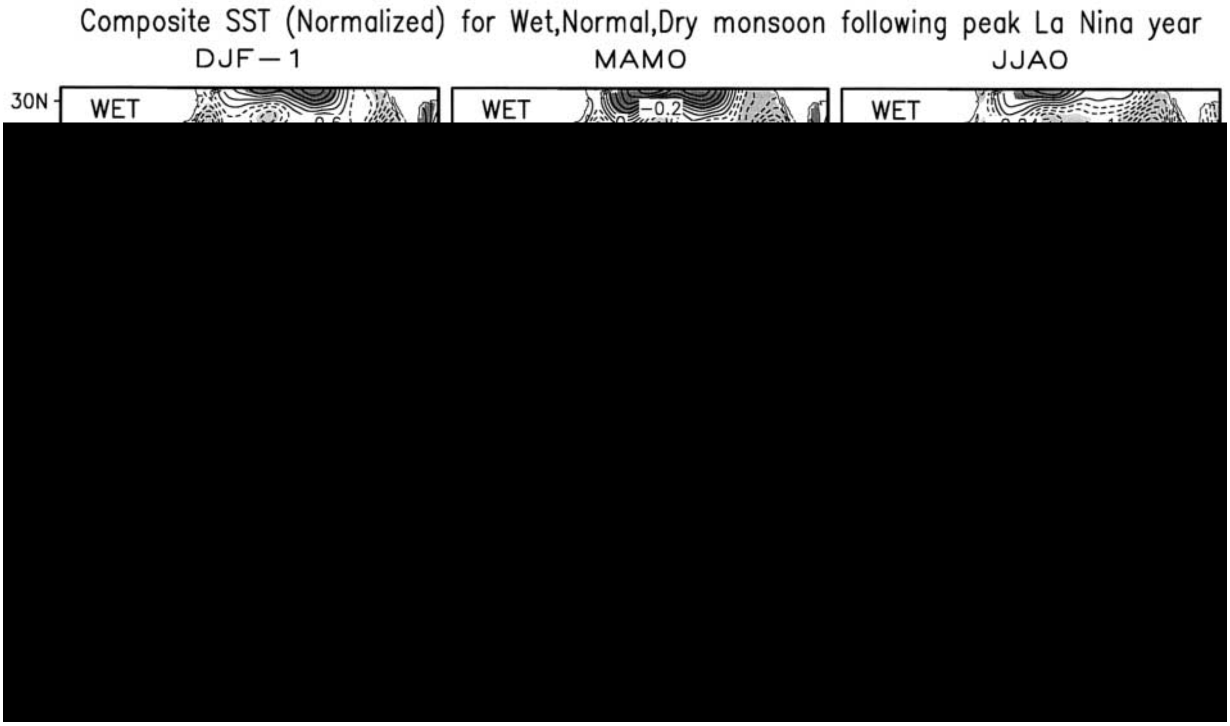

Fig. 9. Composites of anomalous SST fields in DJF (−1), MAM (0), and JJA (0) associated with wet, normal, and dry Indian summer monsoons following a peak La Niña episode in northern winter. The SST fields are normalized with respect to their standard deviations at each grid point. The contour interval is 0.2.

analysis. First, a warm SSTA in the EP in the preceding winter leads to a strong (weak) monsoon in the TBO (LFV) mode. Second, a warm SSTA in the IO in the preceding winter is a precursory signal for a wet monsoon. Third, following a peak El Niño (La Niña), a rapid decay of ascending (descending) motion in the central Pacific is a precursory signal for a wet (dry) Indian monsoon.

5. Discussions

5.1 Role of the Indian Ocean SSTA

Our time-filtering analysis indicates that the Indian monsoon rainfall is significantly correlated with the IO SSTA in the preceding winter and spring on the TBO time scale. A natural question is through what process the IO SSTA in the preceding seasons influences the monsoon.

The IO SSTA may influence the Indian monsoon via both dynamic and thermodynamic effects. The dynamic impact is through induced upward motion in the monsoon sector that may compensate the effect of El Niño forcing from the EP. The thermodynamic impact is through the moisture effect. We argue that a warm SSTA can increase local moisture over the ocean through enhanced surface evaporation. The overall increase of specific humidity over the Indian Ocean is a good precondition for a strong monsoon, because after the monsoon onset the southwesterly flows would transport these excess moistures into the monsoon region. A strong monsoon enhances surface winds that cool the ocean through surface evaporation and ocean mixing, resulting in a colder than normal IO SSTA that further reduces the moisture accumulation and leads to a weak monsoon next year.

In addition to the IO SST, the low-level moisture convergence in the preceding spring is significantly correlated with the monsoon on the TBO time scale. We argue that the anomalous moisture convergence may influence the monsoon through the accumulation of local moisture. As we know, during the dry seasons the local atmosphere over India is in a convectively stable regime. Because of that, the convergent water vapor prior to the monsoon onset can be used only for moistening the local air. The increase of local humidity may initially help strengthen the monsoon intensity, which may further induce anomalous southwesterly flows during the monsoon season and help to bring more moistures from the tropical ocean. Note that this anomalous moisture convergence mechanism differs from the effect of the IO SSTA. The former is associated with the moisture flux by anomalous winds, whereas the latter is related to anomalous moisture advection by the mean monsoon flow.

5.2 Remote SST impacts

While the monsoon TBO is primarily controlled by local processes, the monsoon LFV might be due to the remote forcing of the SSTA in the Pacific. We argue that three possible processes may contribute to the rainfall anomaly on the lower-frequency time scale.

The first is the direct impact of the EP SSTA through the vertical overturning of the large-scale east-west circulation (e.g., Meehl 1987). This mechanism can be readily seen from the wet-minus-dry composites of the velocity potential difference field (Fig. 6).

The second is the effect of the anomalous SST in the WP. Associated with a cold SSTA in the equatorial EP, a positive SSTA appears in the WP. This warm SSTA may further affect local convective activity, and induce anomalous lower tropospheric circulation off Philippines (Tomita and Yasunari 1993). The enhanced convective activity can be readily seen from the bottom panel of Fig. 10, which shows anomalous cyclonic circulation and anomalous low-pressure center at 850 hPa over the Philippine Sea. As we know, the WP monsoon trough is a region of rapid development of synoptic-scale disturbances. Many of the disturbances that develop in the region move northwestward toward Southeast Asia, and some reach the Indian subcontinent (Lau and Lau 1990; Chang et al. 1996). It is speculated that the enhanced convective activity in the WP may increase the frequency or intensity of the northwestward-propagating synoptic-scale disturbances. It is seen from Fig. 10 that an enhanced monsoon trough, oriented from the equatorial WP to the Indian subcontinent (shaded region), appears in the 3–7 year band. In contrary, in the top panels of Fig. 10 (the 2–3 year band), an anomalous high-pressure center and anomalous anticyclonic circulation[2] appear in the WP. Thus, the tendency for more convective activity in the WP near the Philippines during the wet monsoon season appears in the LFV mode, but not in the TBO mode.

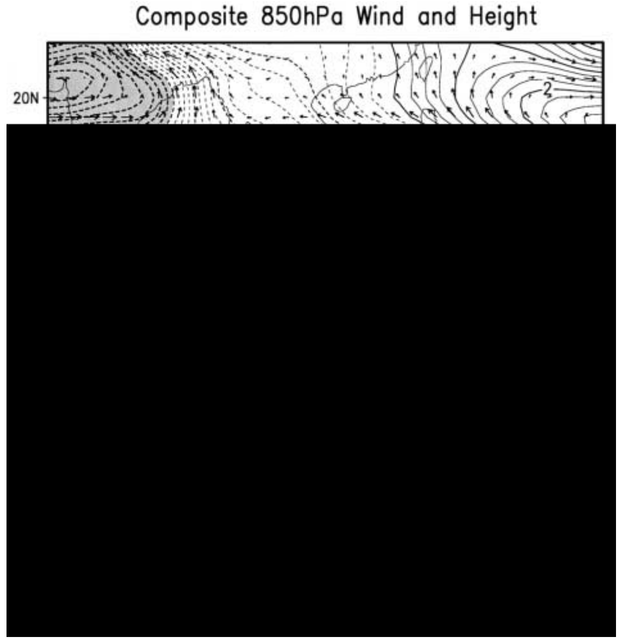

Fig. 10. Wet-minus-dry composites of 850 hPa wind (vector) and geopotential height (contour interval: 0.25 gpm) fields in JJA (0) for the 2–3 year and 3–7 year bands. The composites are obtained from ten wettest and ten driest cases based on the filtered rainfall data.

The third process is attributed to the impact of the remote tropical SSTA forcing on the midlatitude atmospheric circulation. The wet-minus-dry mean tropospheric temperature composite shows that 3–6 months prior to a wet monsoon a north-south thermal contrast has already been established across South Asia and the IO, with the warm core centered over the Tibetan Plateau (Fig. 7, right panel). The

2 Wang et al. (2000) and Chang et al. (2000) pointed out that this anomalous anticyclonic circulation plays an important role in linking the variability of the East Asia monsoon and ENSO.

location of this warm core is consistent with the hypotheses that Tibetan heating and/or Eurasian snow cover prior to the monsoon onset play an important role in the strength of the monsoon (e.g., Mooley and Shukla 1987; Yanai et al. 1992). This differs markedly from the monsoon TBO mode, in which the Indian subcontinent and the IO are both covered by an elongated warm anomaly belt in the preceding winter and spring. Thus, an enhanced (reduced) land-ocean thermal contrast precedes a strong (weak) monsoon on the lower-frequency time scale, but not on the TBO time scale.

5.3 *Net effects of the TBO and LFV modes*

It is likely that the interannual variability of the monsoon is strongly regulated by both the TBO and LFV modes. To quantitatively assess their effects, we constructed linear regression models for both TBO and LFV modes, based on the filtered rainfall data. We selected four predictands for each mode based on the time-filtering analysis results. Table 2 lists all these predictands. It turns out that the regression model well captured the rainfall variability on the two time scales. The correlation coefficients between the regressed and actual rainfall time series exceed 0.73 for both modes.

In the following, we particularly focus on the TBO and LFV mode contributions after 12 major El Niño events during 1950–1999. Table 3 lists these El Niño episodes (in an order of the Niño 3 SSTA magnitude), the sign of actual rainfall anomalies, and contributions from the TBO and LFV modes. Of about a half of these cases the two modes have the same sign (with regard to the rainfall contribution), whereas of the other half they are against each other.

Of particular interest are three strongest El

Table 2. Four predictands for the summer rainfall regression models of the monsoon-TBO and the monsoon-LFV modes

2-3 Year			3-7 Year		
Variable	Season	Domain	Variable	Season	Domain
SST	DJF	70-90°E, EQ.-10°N	SST	MAM	180-160°W, 10°S-10°N
U (850hPa)	DJF	115-135°E, 10–15°S	SST	DJF	140-160°E, 15-25°S
V (Surface)	MAM	65-75°E, 10-20°N	U (850hPa)	DJF	175°E-175°W, 7.5-17.5°N
SLP	DJF	130-150°E, 10-20°S	Thickness (200-500hPa)	DJF	70-90°E, 25-35°N

Table 3. List of 12 major El Niño events (in order of El Niño magnitude) and signs of actual rainfall anomalies and contributions from the TBO and LFV modes

	Nino3 SSTA December (-1)	Observed rainfall JJA(0)	TBO rainfall contribution	LFV rainfall contribution
1997	3.53	+	+	–
1982	3.15	+	+	–
1972	2.22	+	+	–
1965	1.33	–	+	+
1957	1.33	–	+	–
1991	1.29	–	–	–
1987	1.06	+	+	+
1963	0.96	+	+	+
1994	0.93	–	–	–
1969	0.91	+	+	+
1976	0.87	–	–	–
1968	0.81	–	+	–

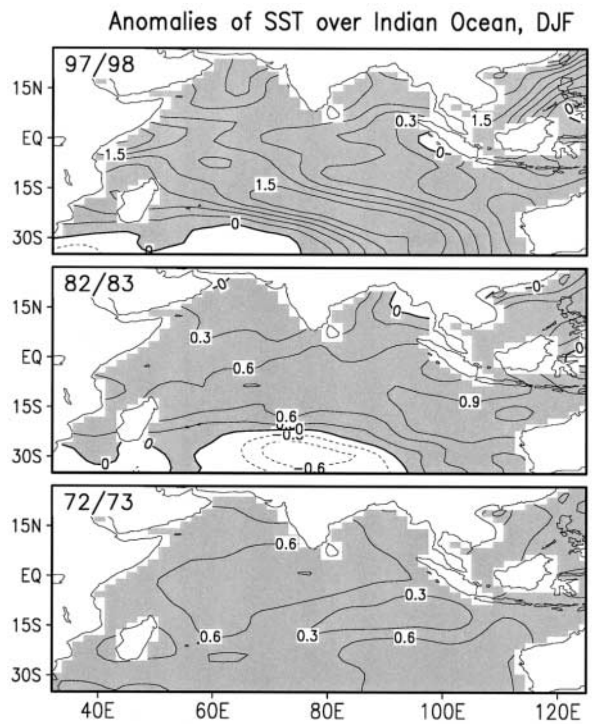

Fig. 11. Anomalous SST patterns over the Indian Ocean in DJF of 1997/98, 1982/83 and 1972/73 (contour interval: 0.3°C).

Niño cases (1997, 1982 and 1972), in which the LFV mode tends to reduce the monsoon rainfall, while the TBO mode tends to enhance it. The actual rainfall anomalies for all three cases are positive, indicating that the TBO mode dominates the LFV mode in these cases. By examining the SSTA pattern over the IO, we found that in all three cases there were basin-wide warmings over the tropical IO in the preceding winter (Fig. 11), a precursory signal for the monsoon TBO mode. In particular, the SSTA in winter of 1997/98 was characterized by a strong warming in the equatorial western IO. The basin-scale warming was associated with the eastward expansion of positive SST anomalies from the western IO, following a strong Indian Ocean dipole event that developed rapidly in the preceding summer and fall (Webster et al. 1999; Saji et al. 1999).

6. Concluding remarks

In this paper the spatial and temporal structures of atmospheric circulation and SST associated with the Indian monsoon rainfall variability on quasi-biennial (2–3 year) and lower-frequency (3–7 year) time scales are investigated by use of both time-filtering and composite analysis approaches. The observational study indicates that physical processes that affect the monsoon rainfall variability on the two time scales differ significantly. Figure 12 is a schematic diagram that encapsulates such a difference. The monsoon biennial variability is primarily affected by local anomalous SST and moisture flux convergence over the tropical Indian Ocean. The lower-frequency variability of the monsoon, on the other hand, results primarily from remote SST forcing in the Pacific through the changes of the large-scale east-west circulation, convective activity along the monsoon trough, and/or land-ocean thermal contrast associated with the tropical-midlatitude teleconnection.

The Indian Ocean SSTA has long been thought to play a minor role in the monsoon variability. The time filtering analysis indicates that on the TBO time scale the Indian Ocean SSTA is highly correlated, at a leading time of 3–6 months, with the Indian monsoon rainfall. The effect of this SST influence is quite different from the remote forcing of the eastern Pacific SSTA, which is dominant on the lower-frequency time scale. This points out the important role the Indian Ocean SST plays in the TBO. It supports the TBO hypothesis that a positive SSTA in the Indian Ocean leads to the increase of surface moisture (due to the enhanced surface evaporation), and thus a strong monsoon owing to the increased moisture fluxes into the monsoon region (Chang and Li 2000).

Our regression model results suggest that

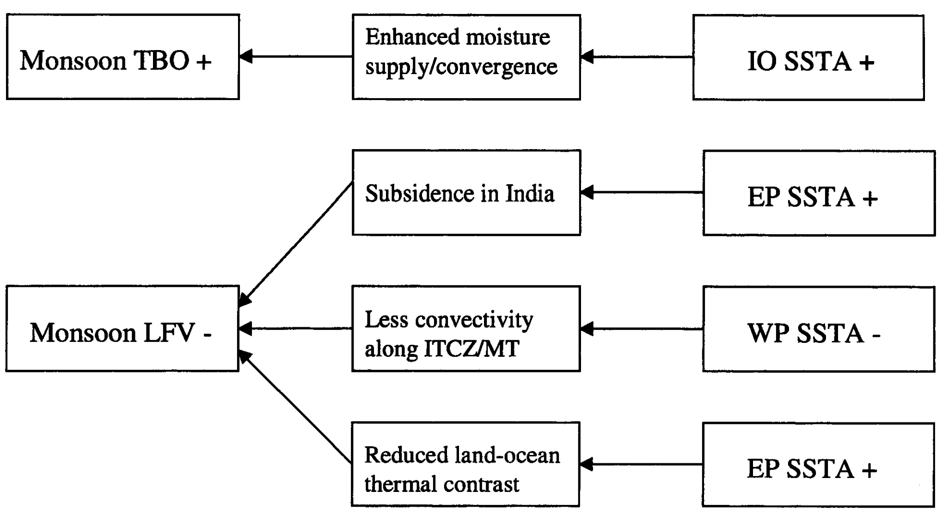

Fig. 12. Schematic diagram showing major processes that influence the quasi-biennial and lower-frequency variability of the Indian monsoon rainfall.

the interannual variability of the monsoon in general depend on the net effect of the TBO and LFV modes. Therefore it is essential to understand the physical mechanisms associated with the two modes. So far there are different views on the origin of the TBO. Some believe that the TBO is an extreme case of ENSO so that the fundamental causes of the TBO and ENSO are the same (e.g., Meehl 1997). Others believe that the TBO is an inherent monsoon mode, resulting from the monsoon-ocean interactions (e.g., Nicholls 1978; Chang and Li 2000, 2001). This observational analysis supports the notion that the origin of the TBO lies in tropical atmosphere-ocean-land interactions in the monsoon sector. While this observational analysis points out the important role of the IO SSTA on the monsoon, it does not rule out possible influences from the EP SSTA, as it is well known that the Niño 3 SSTA also has a biennial component. Further modeling studies are needed to understand the processes through which the Indian Ocean SSTA impacts the monsoon, and the teleconnection mechanism through which the tropical SSTA remotely affects the midlatitude atmospheric circulation.

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