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Tropospheric Biennial Oscillation of ENSO-Monsoon System in the MRI Coupled GCM

By **Norimitsu Ogasawara**

Institute of Geoscience, University of Tsukuba, Tsukuba, Japan

Akio Kitoh

Meteorological Research Institute, Tsukuba, Japan

Tetsuzo Yasunari

Institute of Geoscience, University of Tsukuba, Tsukuba, Japan

and

Akira Noda

Meteorological Research Institute, Tsukuba, Japan

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Abstract

The mechanism of the tropospheric biennial oscillation (TBO) of the ENSO-monsoon system is investigated by an MRI coupled atmosphere-ocean general circulation model. In this mechanism, biennial variability of the South Asian monsoon affects the global scale climate variability through interactions with the air-sea coupled system over the Pacific and/or the extratropical circulation.

In the strong phase of the TBO, the area of relatively strong monsoon convective maximum over South Asia in the spring to summer season moves southeastward to Indonesia in the autumn to winter season. This movement superimposes on its climatological seasonal cycle. It suggests that the northern winter monsoon convection tends to be strong around Indonesia to northern Australia when the summer Asian monsoon is strong. The anomalous state of the air-sea coupled system in the Pacific sector which forms in the summer season, seems to dissipate from its eastern edge. This occurs by a local atmosphere-ocean coupling process through a large scale Walker circulation.

The convection anomalies persist during the entire monsoon season over Indonesia and northern Australia. As a response to this equatorial monsoon convection anomaly, a Matsuno-Gill type stationary Rossby wave is established over the South Asian region. The appearance of upper level anticyclonic circulation and lower level cyclonic circulation anomaly in a strong monsoon year is a favorable condition for bringing the cold air advection over the Eurasian continent. Cold air advection after the strong monsoon persists through the whole winter to spring season to form the cold tropospheric temperature around Central to South Asia. Then reduced land-sea, or north-south temperature contrasts sets up the following weak South Asian summer monsoon. The simulated TBO of the South Asian monsoon is tightly phase locked with a seasonal cycle. The phase of the TBO changes in northern spring, which suggests that the extratropical-tropical interaction be realized mainly during winter to spring through the onset of South Asian monsoon. Our results imply that the TBO is an inherent feature in the land-monsoon-ocean coupled system, and emphasize a more active role of monsoon-extratropical interaction in the Indian sector in winter to spring season for regulating the TBO cycle.

Corresponding author and present affiliation: Norimitsu Ogasawara, Japan Weather Association, Kitaku-Mizukusa 1-21-5, Nagoya, Aichi 462-0042, Japan.
E-mail: noga@jwa.or.jp

1. Introduction

There are a number of observational studies that indicate the tropospheric biennial oscillation (TBO) to be one of the dominant interannual variabilities in the tropical circulations and sea surface temperature (SST) (Trenberth, 1975; Yasunari, 1985; Kawamura, 1988; Ropelewski *et al.*, 1992; Shen and Lau, 1995). The biennial variability in the tropical circulation has a large spatial extent and generally exhibits an eastward propagating feature from the Indian sector to the Western Pacific. Its amplitude is largest along the equatorial band over the Indian to Pacific sector through the maritime continent, with a tendency for the amplitude to become loosely phase locked with the annual cycle (Meehl, 1987; Kawamura, 1988; Barnett, 1991; Shen and Lau, 1995).

As for the biennial mechanisms in the tropics, a conceptual model has been discussed based on local air-sea coupling between the upper ocean heat content and the seasonal cycle of tropical convection through the memory mechanism (Brier, 1978; Meehl 1987, 1993; Yasunari, 1990). The enhanced convective activity during a convective season, due to warmer SST, results in the cooler SST at the end of the season through the air-sea coupling. Then, the cooler SST that persists during the whole following season results in a weakened circulation in the convective season of next year, which in turn warms the SST in the end of the convective season. Although this idea is feasible to explain the apparent phase locking of biennial variability on the annual cycle, only the local air-sea feedback cannot satisfactorily explain the global structure of the TBO. In a larger point of view, the Asian monsoon is thought to play an important role in the mechanism of the TBO as it exhibits a strong annual cycle. Yasunari (1990) shows the existence of a clear relationship between the Indian monsoon rainfall and the Southern Oscillation Index (SOI) in almost two years as a characteristic time scale, and suggests the important role of the Asian monsoon in the mechanism of the TBO. He suggests that in the tropics, the interactive process between the anomalous Asian monsoon flow and the equatorial atmosphere-ocean coupled system in the Pacific, which realizes through the large scale east-west Walker circulation, maintains the biennial oscillation. Additionally, Yasunari and Seki (1992) proposes a role of the tropical-mid latitude interaction on the TBO mechanism; the effect of land surface condition over the Eurasian continent, especially snow cover, modifies the strength of the South Asian summer monsoon through land-sea thermal contrast as noted by previous studies (*e.g.*, Hahn and Shukla, 1976; Morinaga, 1992). The anomalous snow cover over the Eurasia continent is caused by a shift of mid-latitude circulation system that may be forced by the anomalous convection around the tropical Pacific. As a whole, they emphasize a concept of land-atmosphere-ocean coupled system to maintain the biennial variability of the global climate system. The general circulation model (GCM) study of Meehl (1994) also emphasizes the role of the Asian monsoon on the TBO cycle, and notes the tropical and mid-latitude interaction with the monsoon circulation. He suggests that the combination of convective heating anomalies over Africa and the Western Pacific in the northern winter to the spring season, is the most favorable condition to alter the mid-latitude circulation. The shifted extratropical circulation keeps the temperature anomaly over South Asia during the winter to the spring season to set up the following anomalous summer monsoon. However, the combination of convective anomalies in the tropics to alter the middle latitude circulation that he notes is somewhat ambiguous to clearly explain the mechanism of the TBO, because he vaguely discusses the processes in order to bring the complicated distribution of convective anomalies in each of the seasons. As an another component to maintain the TBO cycle, Tomita and Yasunari (1996) notices the "active" role of the feedback process between the northeast winter monsoon and South China Sea SST, to modify the biennial variability of the Asian monsoon and ENSO in the tropics. On the other hand, observational study of Ose *et al.* (1997) regards the variability of the South China Sea SST as an available "passive" indicator of the Asian monsoon and ENSO system.

While the above studies include some types of effect from the middle latitude on the TBO cycle, Chang and Li (1999) explains the mechanism of the TBO with a simple model without considering explicitly the tropical-middle latitude interaction. Their results suggest that the biennial variability of the Asian monsoon occurs due to the SST anomalies in the equatorial Indian Ocean. Although their model fails to properly simulate the change of sign of biennial SST anomaly in the equatorial Pacific in the northern spring, they could reproduce the seasonal migration of the anomalous monsoon convection from South Asia to Australia, which is accompanied by the interactive process between the monsoon region and the equatorial Pacific through Walker circulation.

Although the TBO is indeed the dominant mode in the global climate variability, there are processes that should be investigated in detail to clearly explain the mechanism of the TBO cycle. They include whether the anomalous middle latitude circulation affects (or interacts with) the tropical biennial variability such as the South Asian monsoon, or it merely reflects the response from the tropical TBO cycle. If some type of interaction between the tropics and middle latitude exists, it is important to note how it is realized. It is also important to examine how the seasonal eastward migration of the anomalous monsoon convection from South Asia to Australia during the northern summer to the winter season occurs, and its relation to equatorial Pacific SST variabilities.

In this study, we investigate the mechanism of the biennial variability of the South Asian monsoon, which strongly appears in the Meteorological Research Institute (MRI) atmosphere-ocean coupled GCM, to clarify the essential structure to keep up the TBO cycle. Using the lag-lead regression analysis technique for the biennial component of monsoon index, we discuss the interactive processes of South Asian monsoon circulations with other global phenomena through the global scale circulations. Although our coupled model simulates the realistic interannual variability such as the ENSO or interdecadal variability in the Pacific (Yukimoto *et al.*, 1996; Kitoh *et al.*, 1999b), there exist some differences from the observed climate. However, it is meaningful to investigate and describe the mechanism of simulated TBO in detail because it will give some implications on the study of observed biennial variability.

In Section 2, the model used in this study is described. Definition of a monsoon intensity index and the experimental design for regression analysis are shown in Section 3. The simulated TBO in the tropics concerned with the interaction between the South Asia and the Pacific sector, is described in Section 4. The TBO extratropical signals are then shown in Section 5. Discussion about the mechanisms of the TBO is presented in Section 6, and conclusion follows in Section 7.

2. Model

The global atmosphere-ocean coupled GCM used in this study is developed at the MRI. The atmospheric model (MRI GCM-II, Kitoh *et al.*, 1995) is a new version of the MRI-GCM (Tokioka *et al.*, 1984), which has 4° longitude and 5° latitude of horizontal resolution. There are 15 vertical levels with the model top at 1 hPa. The model includes standard physics, such as a complete hydrological cycle and interactive clouds, with surface fluxes of heat, water, and momentum. Precipitation occurs by the modified Arakawa-Schubert type cumulus convection, moist convective adjustment and large scale condensation. Ground temperature, ground wetness and frozen soil moisture are prognostic variables and calculated at four of each layer of the land surface. Low- frequency behavior of the AGCM forced by observed SST is described in detail by Kitoh *et al.* (1995).

The ocean model is a world ocean general circulation model developed at MRI (Nagai *et al.*, 1992). The model has extended to include realistic bottom topography. The OGCM has a resolution of 2.5° longitude and 2.0° latitude. Within the region of 12°N–12°S, meridional resolution varies nonuniformly from 0.5° to 2.0°, to resolve the equatorial waves and the narrow equatorial upwelling region. There are 21 irregularly placed levels in vertical, 11 of that are located in 300 m depths from the ocean surface. The oceanic mixed layer is calculated following the Mellor-Yamada level 2 turbulence closure schemes.

In this study, we investigate the mechanism of the TBO of the South Asian monsoon using 70 years of integrated data with the atmosphere-ocean coupled GCM. Before coupling, the AGCM was integrated for 3 years with the observed SST, and OGCM was spun up for 1500 years from motionless state with homogeneous potential temperature and salinity. Then preliminary coupling for 30 years integration was made to obtain flux adjustment values of heat and fresh water through the relaxation of sea surface temperature and salinity to climatology. After that, 70 years integration was made with flux adjustment obtained above to preserve realistic climate SST and surface salinity (Tokioka *et al.*, 1996).

Ability of this coupled model to simulate the climatological state and interannual variability of the Asian monsoon and atmosphere-ocean system over the tropical Pacific has been investigated by Kitoh *et al.* (1999b). Yukimoto *et al.* (1996) describes the interannual and interdecadal variabilities in the Pacific sector that is realistically reproduced in the model.

3. Methodology

3.1 Asian monsoon index

In this study, we used the area averaged northern summer (June–July–August) precipitation over South Asia (60°E–90°E, 4°N–26°N) as an index of the South Asian monsoon variability. The area is shown in Fig. 1a with the climatological 850 hPa wind vector and precipitation around South Asia in the northern summer. The model reasonably captures the gross features of the South Asian monsoon circulation such as the Somali Jet or the monsoon westerly flow, although its strength is somewhat weaker than the observed one (Kitoh *et al.*, 1999b). Figure 1b shows regressed northern summer (JJA) precipitation anomalies with the monsoon index for each grid point. The shaded area in the figure indicates where the correlation coefficient is significant at the 95 % level. The area of positive precipitation extending over most of the South Asian region indicates the appropriateness of the monsoon index defined above to represent the variability of the South Asian monsoon. The positive area also spreads around Southern Saudi Arabia in our model. On the other hand, areas of negative

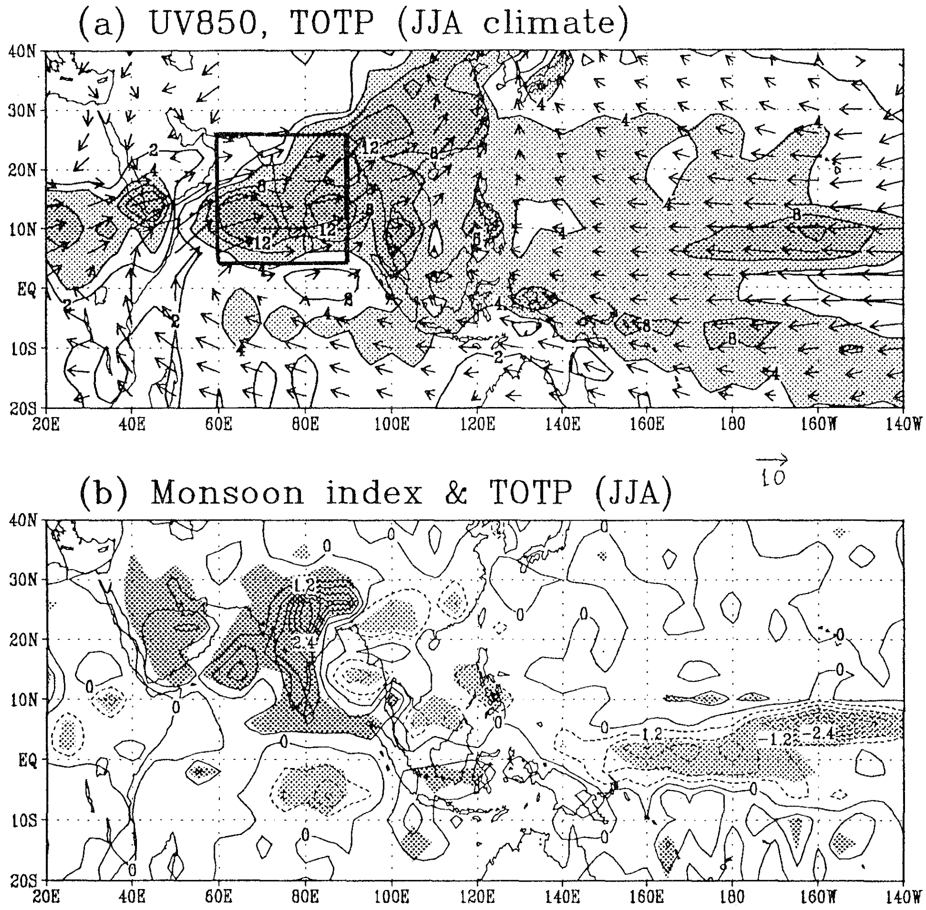

Fig. 1. (a) Climatological northern summer (JJA) precipitation and 850 hPa wind vectors simulated in the GCM. Contours of 2, 4, 8 and 12 are drawn. The rectangle with thick black lines in the figure indicates the area to calculate the monsoon index. The reference vector is 10 m/s. (b) The northern summer precipitation anomalies regressed with the monsoon index. The contour interval is 0.6 mm/day. The dark/light shaded area indicates where the correlation between the JJA precipitation and the index is significant at the 95 % level.

correlation spreads over East Asia and the equatorial central Pacific. These opposite distributions of the precipitation variability between South Asia and the area of South East Asia to the equatorial Pacific corresponds well with the empirical orthogonal function (EOF) of the observed annual mean station precipitation data as shown by Lau and Sheu (1988).

To see the overall feature of the interannual variability simulated in our model, the ratio of the surface temperature variance in biennial range (18 to 30 month) and lower frequency (36 to 84 month), to the total interannual variance is calculated in Fig. 2. The lower frequency variance is dominant over the equatorial Pacific and the Indian Ocean with the ratio exceeding 30 % (Fig. 2a). It is also noticeable over the eastern North Pacific around 140°W, 30°N, and the eastern South Pacific around 100°W, 25°S. The wedge like pattern of lower frequency variance in the Pacific corresponds well with the simulated model El Niño that has a dominant peak around 3 to 6 years (Yukimoto *et al.*, 1996). As for the biennial frequency, the area with the ratio more than 25 % extends over the South China Sea, and the maritime continent (Fig. 2b). An interesting feature is that over the land surface, especially over the Eurasian continent and Saudi Arabia, the biennial variability is dominant with the ratio more than 30 %.

The power spectrum of the monsoon index, calculated by MEM, is shown in Fig. 3. As is seen in the observed one (Parthasarathy *et al.*, 1994), there is a dominant peak around 2 years, as well as a peak around 4 to 8 years corresponding to the ENSO. Relative strength of the ENSO time scale peak to the biennial peak in the GCM is larger than that of the observed one.

The results described above imply that a close relationship exists in the biennial range between the South Asian monsoon variability and the surface temperature over the Eurasian continent. We examine later the physical process which realizes the apparent biennial relationship between them, and dis-

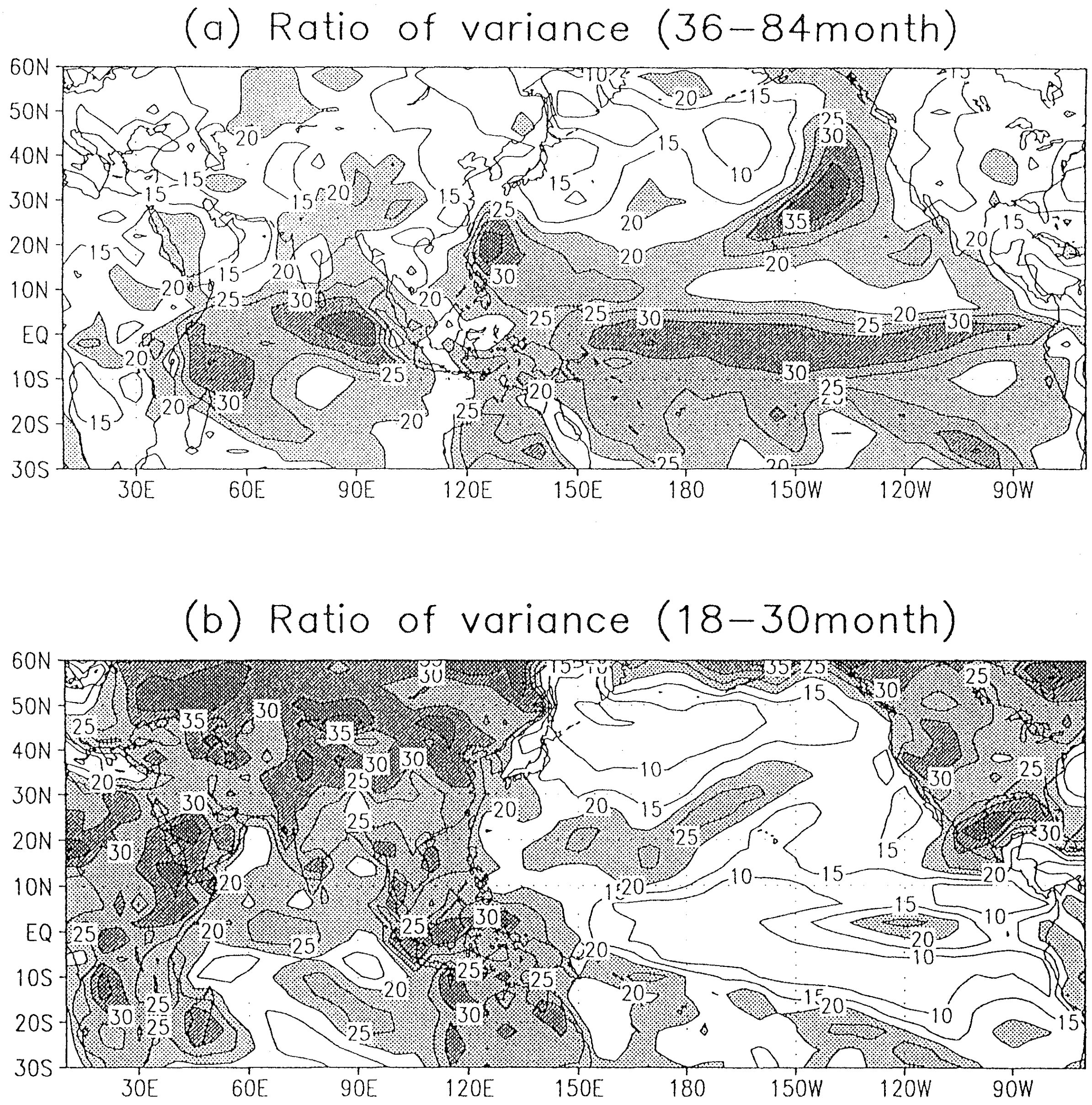

Fig. 2. Ratio of the surface temperature variance in (a) the low-frequency range (36 to 84 month bandpass filtered) or (b) the biennial range (18 to 30 month bandpass filtered), to the total interannual variance. Contour interval is 5 percent and the areas larger than 20 are shaded.

cuss the essential role of it to maintain the TBO variability of the South Asian monsoon and the equatorial Pacific ocean-atmosphere system.

3.2 The cross correlation technique

The cross correlation technique is used to search for relationships between the South Asian monsoon variability and the atmosphere-ocean coupled system of the equatorial Pacific, or the middle latitude circulation in the biennial range. The methodology used for the cross correlation is identical to that of Kiladis and Weickman (1992). At first, the entire monthly model output data are bandpass filtered into a 15–30 month band. The filtered monsoon index, which is defined as the northern summer season precipitation averaged over the South Asian region, is then regressed against the seasonal mean model parameters such as SST or u and v wind component at each grid point. Lagged regression relationships for the seasonal mean value during the northern spring season (March–April–May) prior to a strong summer monsoon (lag $= -1$ season) to the following spring season (lag $= +3$ season), that is prior to a weak monsoon, can be used to examine the evolution of convection and circulation signals associated with the TBO of the South Asian monsoon.

Following Livezey and Chen (1983), the "local" significance is calculated by assessing significance of

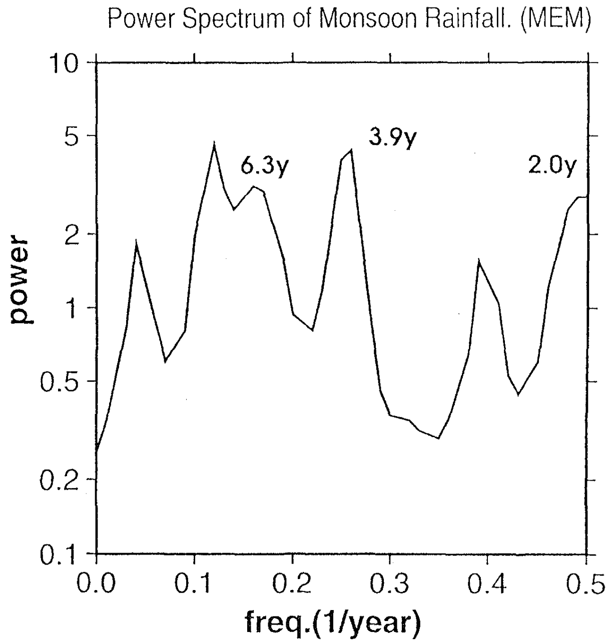

Fig. 3. Power spectral densities of the simulated South Asian monsoon index calculated by MEM.

the linear correlation coefficient between the monsoon index and various components, with an appropriate adjustment of the degree of freedom based on the autoregressive properties of filtered data. The shaded area in the following maps of regression coefficient indicates where the local linear correlation is significant at the 95 % confidence limit. Regression maps of winds are reconstructed where the local significance of either the u or v component is statistically significant at the 95 % level. The "field" significance of a given pattern is then tested with the Monte Carlo simulation. Using this technique, each of the maps that has significant field at the 90 % level is marked by an asterisk at the upper right side of each map.

4. Simulated TBO of the South Asian monsoon

In this section, circulation and convection anomaly, and SST anomaly field in the tropics corresponding to a strong South Asian monsoon year is described concerning the seasonal evolution of the TBO cycle.

4.1 850 hPa wind and precipitation fields

Figure 4 shows seasonal evolution of the regressed 850 hPa circulation field for a strong monsoon year. In the figure, the wind vector is plotted only where the cross correlation coefficient for either u or v component is significant at the 95 % level. The composite anomaly field of precipitation in the biennial range is also shown in the figure to examine the circulation anomaly field in relation to the tropical convective activity.

In the northern spring season (MAM) before a strong monsoon, there is a significant westerly anomalous flow over the Arabian Sea, which corresponds to an earlier onset of the South Asian summer monsoon (Fig. 4a). At the same time, the noticeable easterly anomalous flow is established from the South China Sea to the Bay of Bengal. Low level convergence of the stronger monsoon westerly and easterly anomalous flow forms an area of strong precipitation over the Indian subcontinent. As the easterly anomalous flow superimposes on the climatological westerly monsoon flow over the Bay of Bengal and the South China Sea, the area of weakened moisture flux and reduced precipitation is formed. It should be noted that the significant easterly anomalous flow is apparent over the equatorial central Pacific, with weak positive convection anomalies around the western Pacific and the SPCZ.

In the summer season (JJA), the anomalous monsoon flow over the Arabian Sea becomes stronger (Fig. 4b). The noticeable anomalous cross-equatorial Somali jet also indicates a stronger South Asian monsoon. Easterly wind anomalies over the Bay of Bengal, which has been seen from the previous season, extend zonally to cover the western to central equatorial Pacific. Positive precipitation anomalies over South Asia and Indonesia are apparent because of the low level moisture flux convergence there. Area of a negative precipitation spread over Southeast Asia to the equatorial central Pacific. This large scale east-west seesaw pattern of convection anomaly corresponds to an establishment of a stronger east-west Walker circulation.

In the following autumn (SON), the area of significant westerly wind extends to most of the South Asian region, with the easterly flow over the Bay of Bengal weakened (Fig. 4c). Corresponding to the weakened lower level convergence over South Asia, the area of strong monsoon precipitation reduces. The center of strong precipitation anomaly moves southeastward from South Asia in summer to Indonesia around 100°E, 0°N in autumn with a seasonal cycle. The easterly wind anomaly over the equatorial Pacific, which has transported moisture into South Asia in the previous season, reduces its zonal extent only to the equatorial western Pacific. Large scale east-west seesaw patterns of the convection anomaly still exists, though its zonal scale becomes smaller in autumn according to southeastward movement of the center of monsoon convection.

By the northern winter (DJF), the axis of the westerly wind anomaly shifts southward toward the equatorial Indian Ocean (Fig. 4d). Precipitation anomalies over the Indian subcontinent disappear, and the area of positive precipitation anomalies appear around the Indonesia and Australian region,

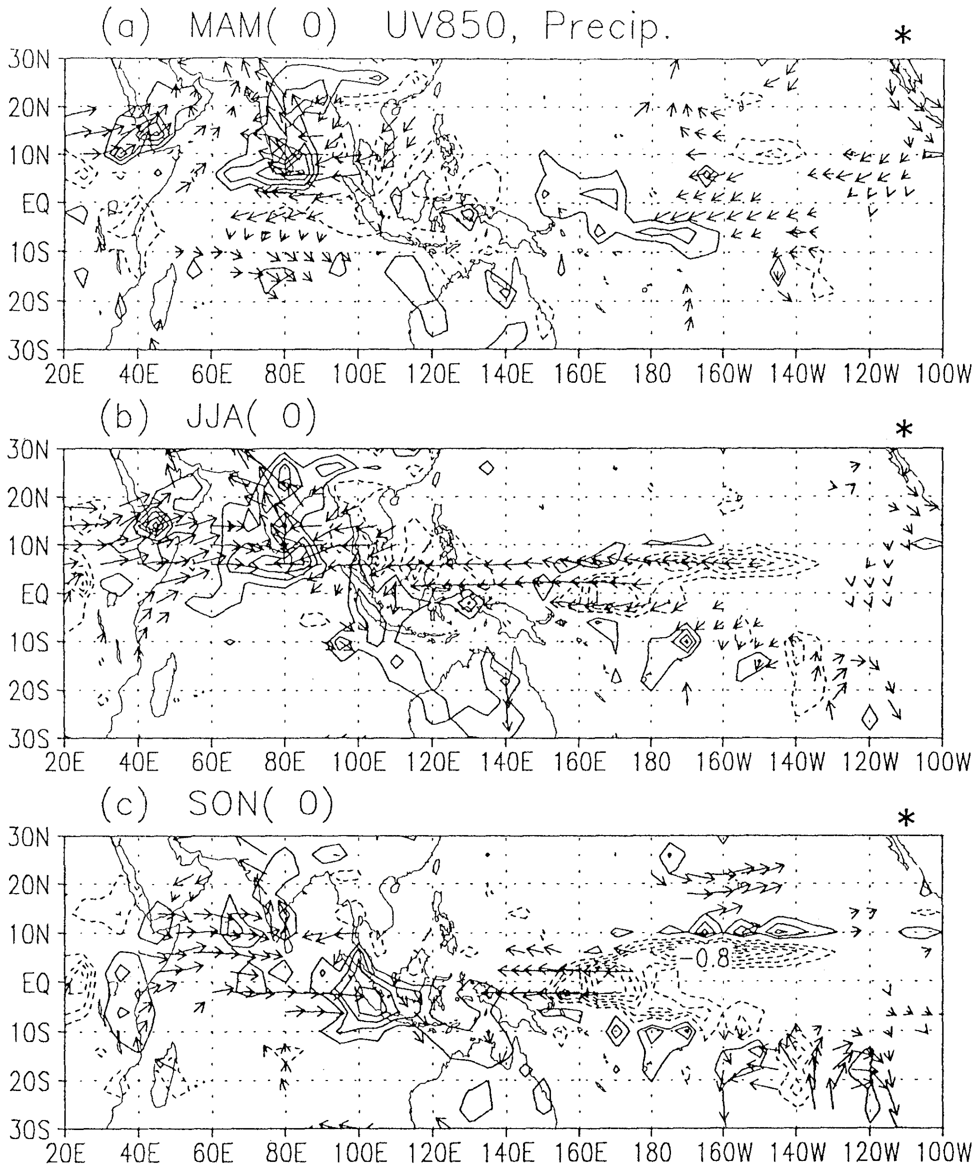

Fig. 4. 850 hPa wind (vector) and precipitation (contour) anomalies in the 15–30 month band. Only those wind vectors where the local correlation coefficient is significant at the 95 % level for either u or v component are shown. The contour interval is 0.2 mm/day. The reference vector is 0.4 m/s. Zero contours for precipitation anomalies are omitted. Seasonal panels for the strong monsoon year are shown for, (a) the spring (MAM0), (b) the summer (JJA0), (c) the autumn (SON0), (d) the winter (DJF0), and (e) the spring season (MAM + 1) of the next year. Asterisks on the upper right of each maps indicate where satisfies the significant field at the 90 % level.

indicating a stronger Australian monsoon. It is interesting to note that the significant northerly wind covers the Arabian Sea associated with a stronger winter monsoon flow. In the equatorial Pacific, the area of the easterly wind anomaly becomes smaller over the western Pacific, although a negative precipitation anomaly still exists.

In the next spring season following a strong monsoon, and prior to a weak monsoon (Fig. 4e), the wind field around South Asia is almost the same as the previous year's one, but with its sign reversed. The TBO cycle of a weak South Asian monsoon year continues starting at the northern spring.

4.2 Sea surface temperature

Figure 5 shows the seasonal evolution of the SST anomaly field in a strong monsoon cycle. The shaded areas in the figure indicate where they are statistically significant. In the figure, we can see two different types of noticeable features in the Indian sector and the Pacific sector, respectively.

In the spring to summer season over the Indian sector, there is a significant positive SST anomaly

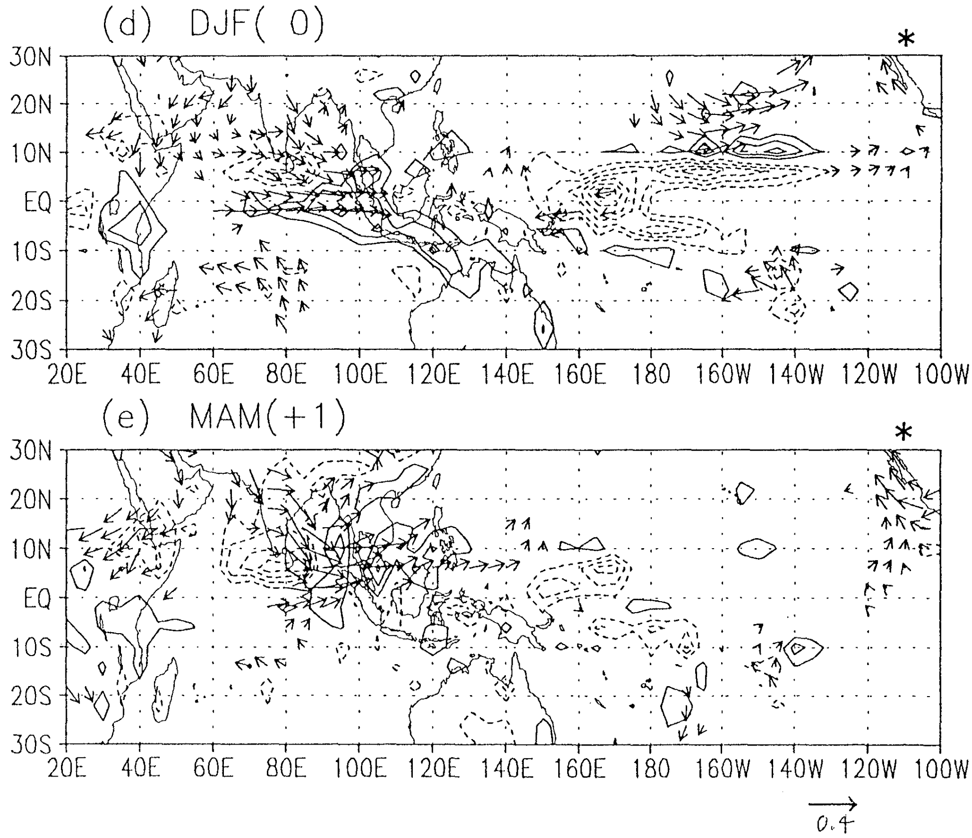

Fig. 4. (Continued)

over the Arabian Sea (Fig. 5a, b). The warmer SST over the Arabian Sea seems to be a favorable condition for a strong monsoon through increased evaporation and enhanced convection. The stronger summer monsoon circulation enhances the evaporative cooling at the ocean surface and vertical mixing of water over the Arabian Sea, which results to turn the warm SST anomaly to relatively cool in the following winter to spring season (Fig. 5d, e). As is shown in Fig. 4d, the significant northerly flow associated with a stronger winter monsoon over the Arabian Sea may contribute to keep the ocean surface cooled through the whole winter to spring season. In the following spring season, prior to a weak monsoon (Fig. 5e), cooler SST over the Arabian Sea contributes to a weak monsoon via reduced evaporation and suppressed convective activity.

In the northern spring of the equatorial Pacific sector, there is a small area of negative SST around 150°W (Fig. 5a). In the following summer season (Fig. 5b), significant negative anomalies spread over the central to eastern equatorial Pacific. This corresponds to the anomalous easterly wind over the equatorial central Pacific as seen in Fig. 4b. Our results agree with the observed features that when precipitation is above normal in the Indian monsoon region, SST and precipitation over the equatorial Pacific is below normal. This relationship is also found in the ENSO time scale so that the model El Niño year shows reduced Indian summer monsoon rainfall (Kitoh *et al.*, 1999b).

It is interesting to note that the negative SST anomaly over the equatorial central Pacific forms abruptly at northern summer, following an occurrence of strong Asian monsoon convection in the previous season (see Fig. 4a). This indicates that the TBO anomaly of the South Asian monsoon can be established without a direct effect of the variability of the central Pacific SST. According to Barnett *et al.* (1989) or Ose (1996), which studies snow-monsoon relationship using AGCM, the anomalous South Asian monsoon with springtime anomalous snow mass over Eurasia can result in forming an anomalous flow over the equatorial Pacific in summer through a Walker circulation. Therefore, it could be considered that the South Asian monsoon might have an effect on the variability of the central Pacific SST through the large scale Walker circulation. By the following autumn and winter (Fig. 5c, d), the area of negative SST becomes smaller by disappearing gradually from its eastern edge. Negative SST anomaly is only seen around 160°E–180°E at the equator in the winter. Then, in the next spring season before a weak monsoon, the negative SST anomaly almost disappears.

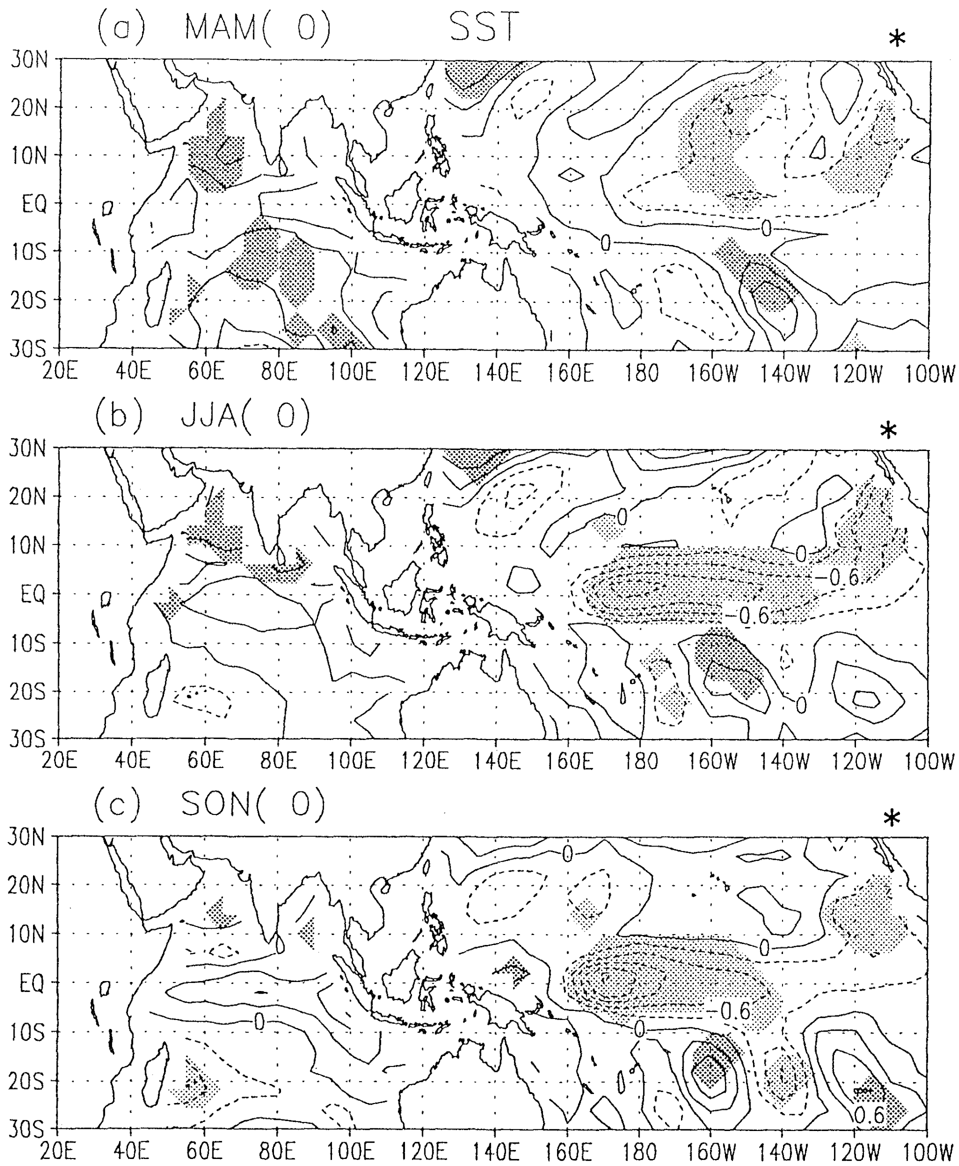

Fig. 5. The same as Fig. 4, except for the SST anomalies. The contour interval is 0.3 K. The dark/light shaded areas are significant at the 95 % level.

4.3 Velocity potential field at 200 hPa

To examine a large scale aspect of circulation anomaly associated with the simulated TBO, we show the regressed anomalies of velocity potential at 200 hPa in the strong monsoon cycle in Fig. 6. The negative and positive anomaly means a large scale divergence and convergence field, respectively. Shaded area indicate where significant linear correlation relationship is satisfied at 95 % level.

In the northern spring (Fig. 6a), a significant negative anomaly exists over the Arabian Sea, corresponding to the strong monsoon convection (see Fig. 4a). In the following northern summer (Fig. 6b), the area of strong divergence expands to cover the entire South Asia with its center located at 10°N, 80°E. At the same time, the significant convergence anomaly is formed over the equatorial Pacific with its center located at 170°W, corresponding to the negative precipitation anomaly (Fig. 4b). Establishment of this global scale Walker circulation is expected from the low level wind and precipitation anomaly fields as discussed in the above section.

By the northern autumn to winter season (Fig. 6c, d), center of strong divergence moves southeastward to 100°E, 0°N in autumn to 110°E, 0°N in winter, with its spatial extent decreasing. At the same time, area of upper level convergence in the Pacific region becomes smaller dissipating from its eastern edge. Moreover, the center of convergence maximum moves westward, around 180°E, 0°N in autumn and 170°E, 0°N in winter. It is interesting to note that the anomalous state of a Walker

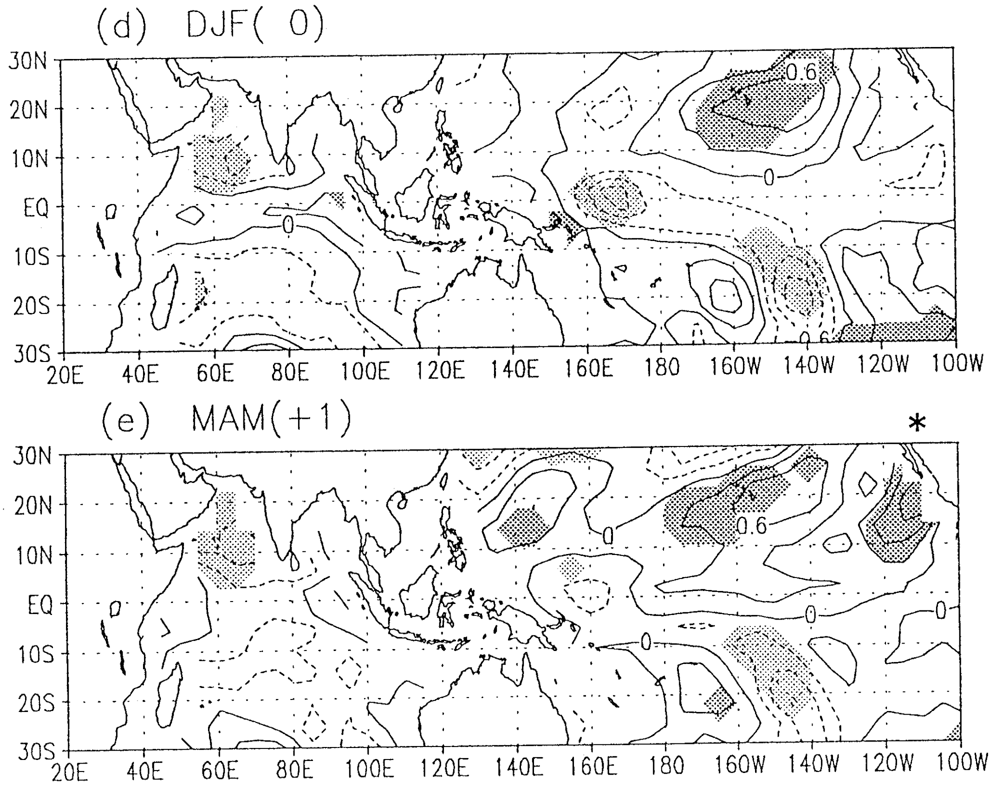

Fig. 5. (Continued)

circulation, between the Indian sector and the Pacific sector, is kept through the northern summer to the winter season, though its spatial extent becomes somewhat smaller.

In the following spring season (Fig. 6e), there seems no noticeable upper divergent anomaly over the equatorial Pacific. An upper convergent anomaly associated with a weak South Asian monsoon is established around the Arabian Sea. Then the course of weak TBO cycle of upper circulation continues at almost the same as that for a strong monsoon cycle, with its sign reversed.

4.4 Wind-SST interaction in the equatorial Pacific region

It is suggested that the moisture transport from the equatorial Pacific to the South Asian region associated with an anomalous Walker circulation may contribute to the monsoon convective activity through the northern summer to winter season (*e.g.*, Fig. 4). We discuss here the mechanism of the TBO seasonal cycle in the tropical Pacific atmosphere-ocean system. Figure 7 shows a longitude-time cross section of the regressed SST anomaly at the equator, during a weak monsoon year (annotated by −1 at y-axis) to a strong monsoon year (annotated by 0 at y-axis). The regressed 850 hPa wind vectors in a biennial range are also plotted to show the equatorial east-west circulation.

We can confirm in the figure that in a weak monsoon year, the SST is warmer over the central to eastern Pacific, and the westerly anomalous wind extends over the Indian sector to the central Pacific, and vice versa for a strong monsoon year. The SST anomaly over the Pacific reverses its sign gradually from its eastern edge during winter to spring. It is interesting to note that in the autumn season, when the SST anomaly peaks, we can see an area of zonally convergent (divergent) wind anomaly around 160°W in the weak (strong) monsoon year. This indicates the establishment of the secondary circulation due to the stronger upward (downward) motion at the eastern periphery of a Walker circulation (see Fig. 6c).

In autumn of a weak monsoon year, the anomalous easterly flow over the eastern Pacific enhances the climatological easterly trade wind. This anomalous easterly wind may enhance the surface latent heat flux and oceanic upward motion through the Ekman pumping to cool the SST around there. Then the anomalous warm SST over the Pacific region dissipates from its eastern edge, and the center of the warm SST apparently moves westward during the autumn to winter season. As expected from Fig. 4, convection activity around the equatorial central Pacific is enhanced corresponding to the warmer SST anomaly. Reduced shortwave radiation at the ocean surface due to the increased cloudiness, may also contribute to cool the SST. Westerly

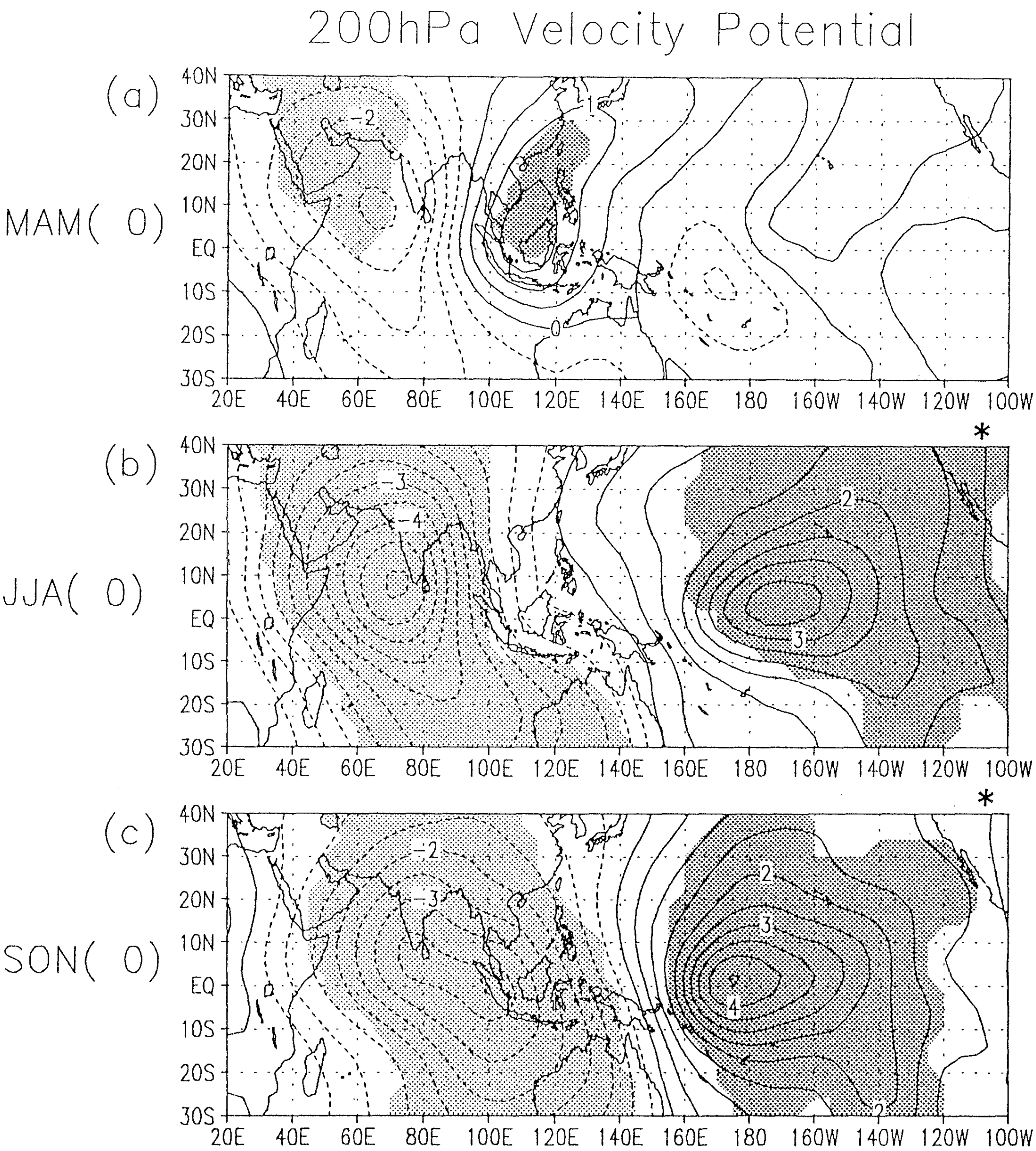

Fig. 6. The same as Fig. 4, except for the velocity potential anomalies at the 200 hPa level. The dark/light shaded areas are significant at the 95 % level.

anomalous flow appears over the eastern Pacific in autumn of a strong monsoon year. It corresponds to downward motion associated with the cooler SST and reduced convective activity over the central Pacific (Fig. 4), then it results in the reduced climatological easterly trade wind. The reduced surface latent heat flux and suppressed oceanic equatorial upward motion due to the weak trade wind, and more incoming shortwave radiation due to reduced cloudiness contributes to warm the SST. Then the negative SST anomaly in the Pacific dissipates from its eastern edge.

These results suggest that the Pacific side of the anomalous state of Walker circulation decays gradually through the wind-SST negative feedback during autumn to winter season. As noted by the result of Kitoh *et al.* (1999a) using the slab ocean GCM, interannual variability of the SST in the Pacific sector could be realized through the atmosphere-SST feedback without the ocean dynamics. It may be said that the anomalous state of the equatorial Pacific SST, which is responsible for the monsoon circulation variability, could decay through its internal atmosphere-SST negative feedback. Then the reversed Pacific SST anomaly is formed in the next year, associated with the opposite anomalous state of the monsoon circulation.

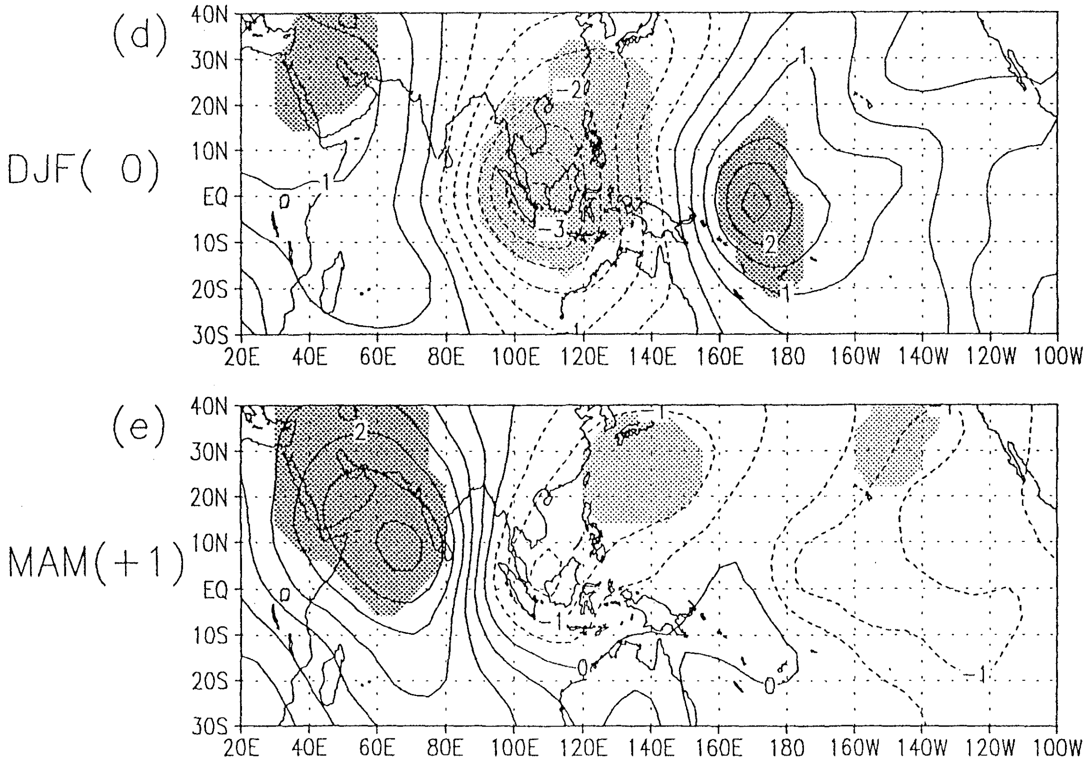

Fig. 6. (Continued)

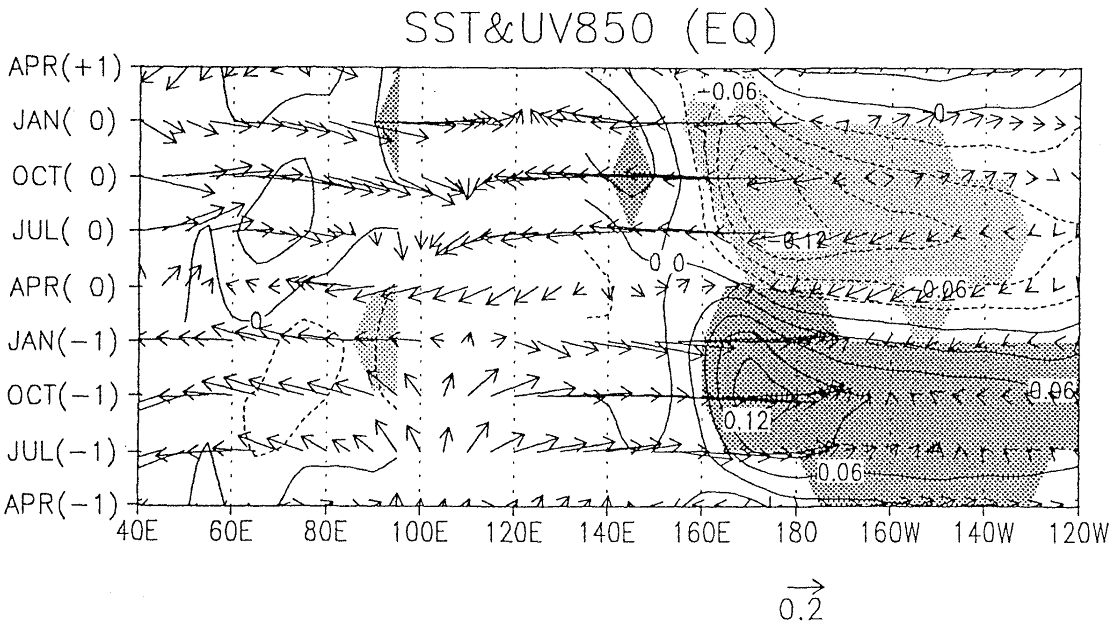

Fig. 7. Longitude-time cross section for the equatorial SST and 850 hPa wind anomalies in the 15–30 month band. Seasonal sequences during the weak monsoon year (noted by −1 on y-axis) to the strong monsoon year (noted by 0) are shown. The dark/light shaded areas indicate where they are significant at the 95 % level for the SST. The contour interval is 0.03 K. The reference vector for 850 hPa wind is 0.2 m/s.

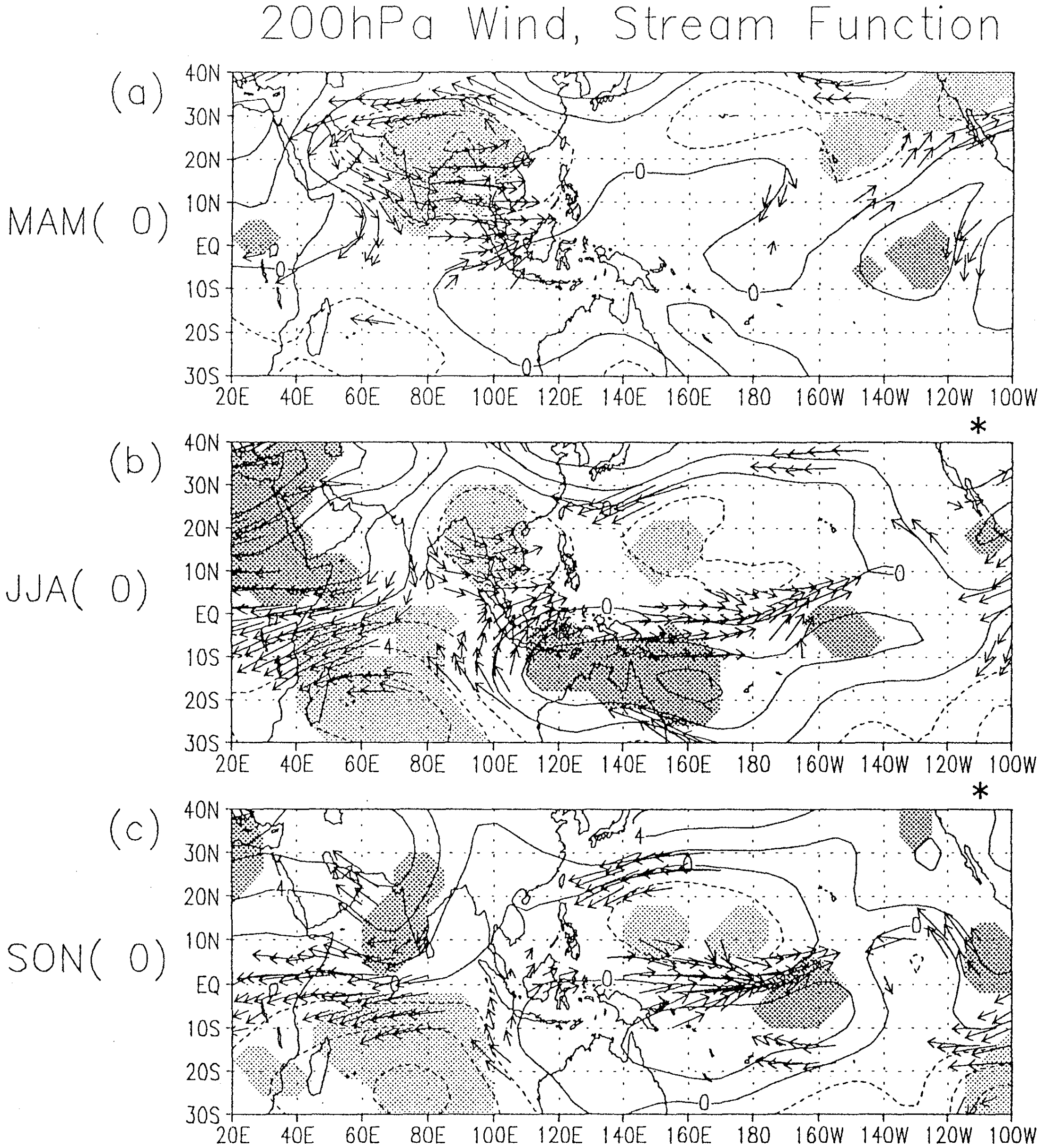

Fig. 8. 200 hPa stream function and wind anomalies in the 15–30 month band. The dark/light shaded areas are significant at the 95 % level for the stream function. Only locally statistically significant wind vectors are shown. The reference vector is 0.6 m/s. Seasonal panels for the strong monsoon year is shown for, (a) the spring (MAM0), (b) the summer (JJA0), (c) the autumn (SON0), (d) the winter (DJF0), and (e) the spring season of the next year (MAM + 1). Asterisks on the upper right of each maps indicate where satisfies the significant field at the 90 % level.

5. Extratropical signals in relation to the TBO

5.1 Midlatitude atmospheric response to the tropical convection anomaly

The extratropical circulation signals associated with the TBO tropical convection anomaly is investigated to see whether the South Asian monsoon variability could affect the extratropical climate, and the tropical-extratropical interaction contributes to the TBO cycle. Figure 8 shows the regressed stream function and wind anomaly field at 200 hPa for a strong monsoon year. The positive and negative stream function anomalies imply the anticyclonic and cyclonic circulation, respectively. The shaded area indicates where the linear correlation coefficient between the stream function and the monsoon index is significant at the 95 % level. Wind vectors are plotted only where either the u or v component is locally significant at the 95 % level.

Around the equatorial Indian Ocean to the Pacific, there is a significant upper level zonally divergent flow in the summer, corresponding to the

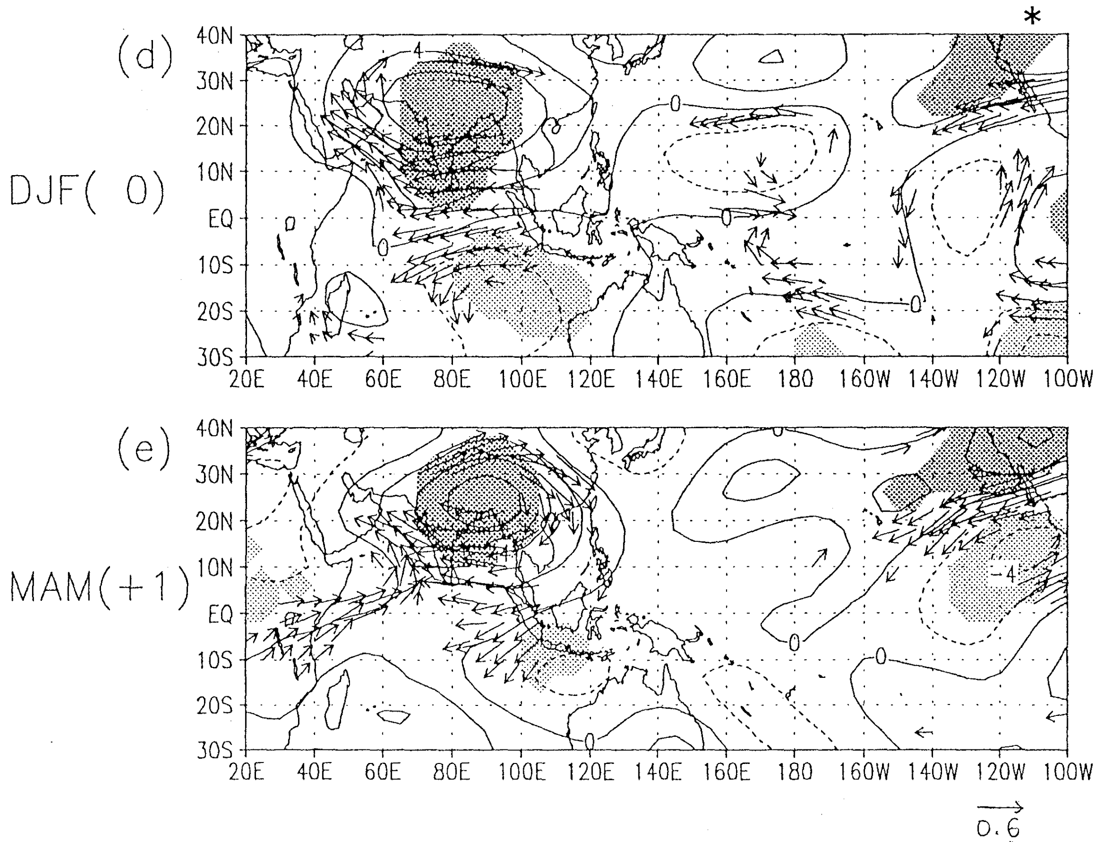

Fig. 8. (Continued)

strong monsoon convection (Fig. 8b). By the autumn to winter season (Fig. 8c, d), the center of the upper divergence moves eastward to 120°E–140°E from 80°E in the previous season.

An interesting feature in the extratropical band is the formation of a significant anticyclonic circulation anomaly over the South Asian region in the northern winter after the strong monsoon. It locates northwestward of the anomalous monsoon convection around Indonesia (Fig. 8d). In the following spring (Fig. 8e), there still exists the anticyclonic circulation over South Asia with its center around 90°E, 25°N, that is northwestward of the anomalous convection around the maritime continent. Before a strong monsoon season, there is significant cyclonic circulation at 200 hPa over South Asia (Fig. 8a). The formation of this extratropical circulation anomaly during the winter to spring season, matches well with the seasonal migration of the relatively strong convective maximum from South Asia to the Australian region (see Fig. 4).

Figure 9 shows the anomalous stream function at 850 hPa around South Asia to the Eurasia continent in the northern spring season. The thick black line in the figure means a 1500 m contour line of the model orography and interpreting the stream function in the circle should be attentively made. To represent an anomalous lower level circulation, regressed surface wind stress vectors are also plotted in Fig. 9. It should be noted that the magnitude of the surface wind stress is generally larger over the land surface than that over the oceans. The distributions of surface wind stress anomalies correspond well to the 850 hPa stream function anomalies even around the Tibetan Plateau. In spring before the strong monsoon (Fig. 9a), a significant anticyclonic circulation anomaly spreads over South Asia, and southerly wind penetrates to Central Asia. This low level anomalous circulation corresponds well to the upper level anomalous circulation, with the opposite sign. In the spring season of next year (Fig. 9b), almost the same wind fields spread over South Asia with its sign reversed to the previous years'. There are also significant stream function anomalies west of the Caspian Sea to Europe. It exhibits a cyclonic (anticyclonic) circulation anomaly before a strong (weak) monsoon, which captures an observed feature associated with biennial variability of the South Asian monsoon (Yasunari and Seki, 1992).

The circulation anomaly in winter to spring over South Asia exhibits an upper level cyclonic/anticyclonic circulation, with the opposite one at the lower level. This formation of anomalous circulation is explained as a stationary Rossby

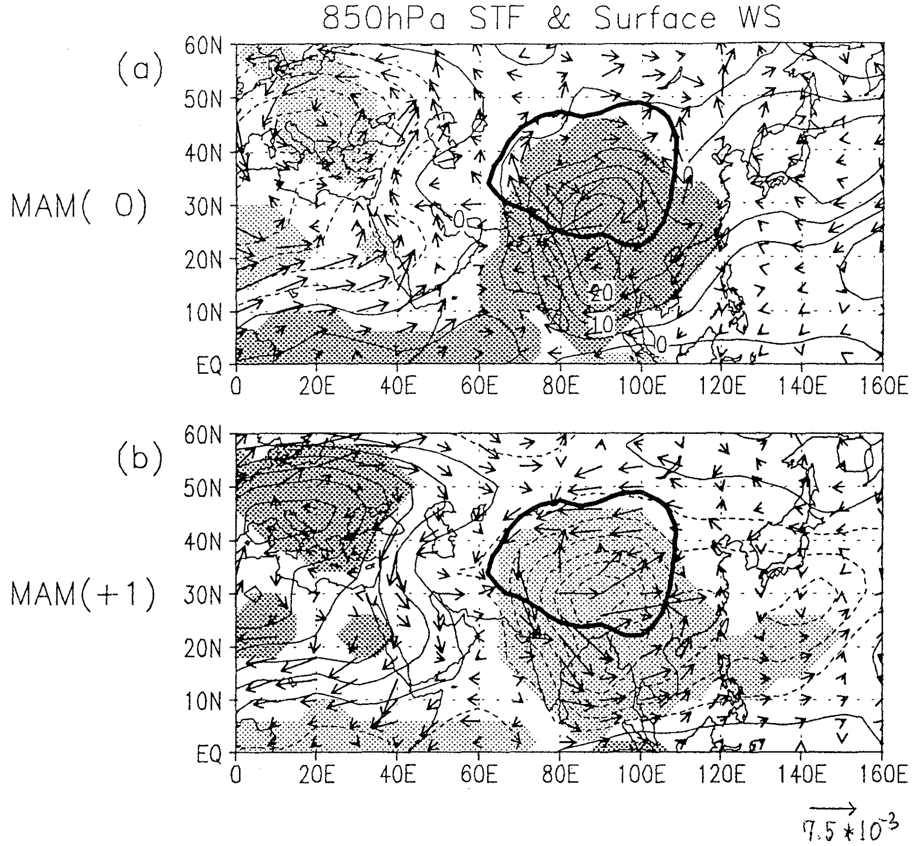

Fig. 9. 850 hPa stream function and the surface wind stress vector anomalies in the 15–30 month band. The dark/right shaded areas are significant at the 95 % level for the stream function. The reference vector is 0.075 N/m2. The thick black line is 1500 m model topography. (a) MAM(0), before the strong monsoon. (b) MAM(+1), before the weak monsoon.

wave response to the equatorial monsoon convection anomaly around the Australian region, as introduced by Matsuno (1966) and Gill (1980). Particularly, it resembles the solution of Gill (1980), when the heating source is placed north of the equator (see Fig. 3 and 4 of Gill, 1980). It seems to be reliable because the anomalous convection persists through the whole winter to spring season over the Australian region. The persistent low level cyclonic (anticyclonic) flow after the strong (weak) monsoon over South Asia around 10°N to 40°N seems to be a favorable condition for bringing the temperature advection over the land through the winter to spring season, then modifying a land-sea tropospheric temperature contrast.

5.2 Land-sea tropospheric temperature contrast in the Indian sector

It has been studied that the fundamental forcing of the South Asian monsoon is a land-sea or north-south tropospheric temperature contrast, and interannual variability of the monsoon might be realized through anomalous heating to change the temperature contrast. In this context, the possibility of a close link between the monsoon variability and winter to springtime surface conditions over the Eurasian continent has been discussed (*e.g.*, Hahn and Shukla, 1976; Morinaga, 1992). These studies suggest that the summer monsoon rainfall tends to be below normal when the snow cover over the Eurasian continent in the preceding winter and spring is above normal. There are also a number of studies of the relationship between the Eurasian land surface condition and the following monsoon strength by GCMs. Yamazaki (1989) shows that the increased land surface albedo weakened the subsequent summer monsoon. The increased snow mass experiments (Barnett *et al.*, 1989; Yasunari *et al.*, 1991; Ose, 1996) noted the two physical processes of anomalous snow to produce a cooling source in the atmosphere that leads to a weak summer monsoon. One is the effect of increased surface albedo by anomalous snow cover. Another is the hydrological effect caused by melted snow, which inhibits surface sensible heating through the absorption of the latent heat of fusion. The result common in these GCM studies is that the land surface condition is an important factor affecting the Asian monsoon strength through the tropospheric temperature gradient.

Figure 10 shows the seasonal cycle of regressed

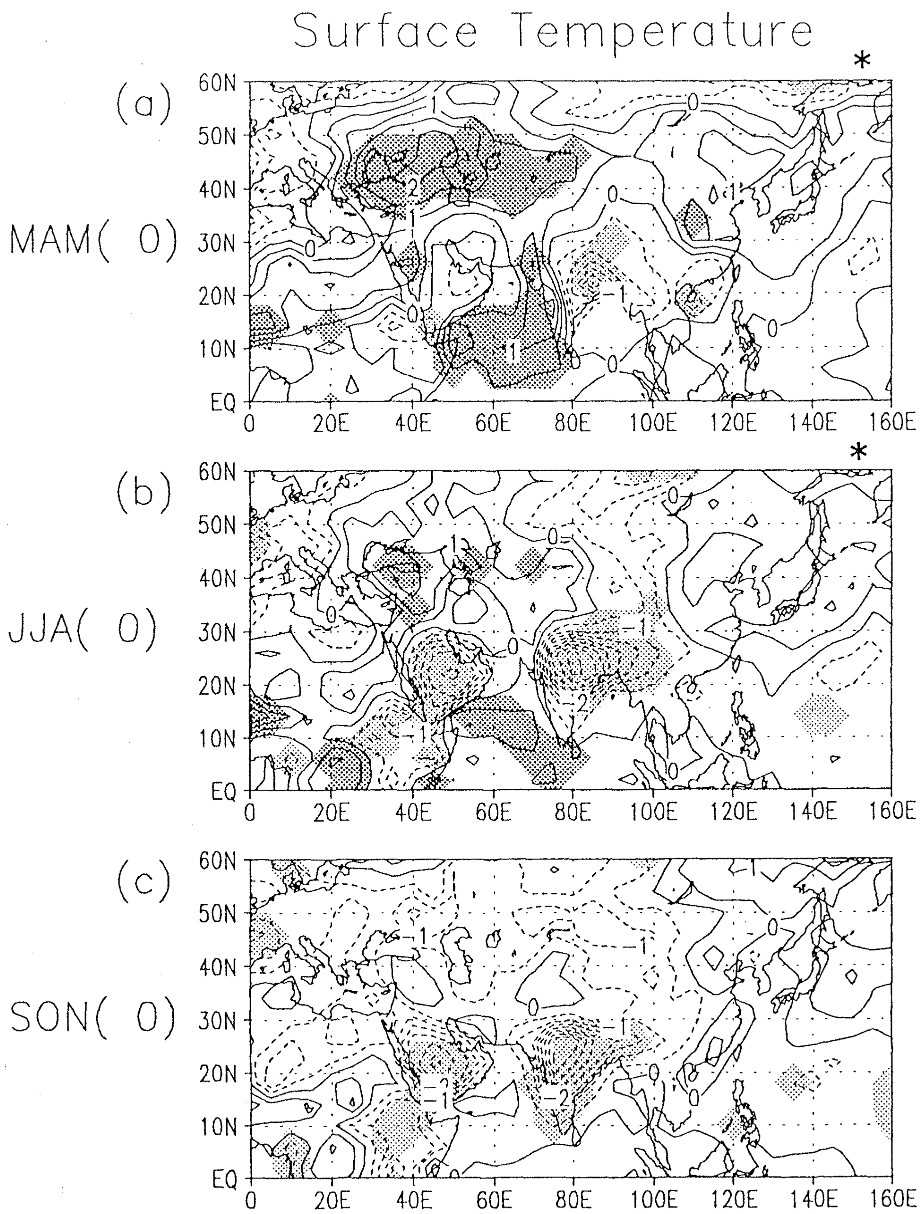

Fig. 10. The same as Fig. 8, except for the surface temperature anomalies. The dark/light shaded areas are significant at the 95 % level. Contour interval is 0.1 K.

surface temperature anomalies in a strong monsoon year. The shaded area represets the linear correlation coefficient is significant at the 95 % level. It is clear in the figure that the surface temperature over the Eurasian continent around 30°E–80°E, 35°N–50°N is significantly warm before a strong monsoon (Fig. 10a). It is also significantly warm over the Arabian Sea as shown in Fig. 5a. The area of significantly warm temperature around the South to Central Asia corresponds well to the area of southerly wind anomaly (see Fig. 9a). The cool land surface over the eastern Indian subcontinent is due to the wetter land surface and reduced surface solar radiation by the earlier onset of monsoon rainfall there. The significantly warm temperature dissipates by autumn and turns its sign to significantly cool in the following winter season (Fig. 10b, c, d). The cooler surface temperature over land and the Arabian Sea persists during the winter and spring season to form a reduced land-sea temperature contrast, then sets up a following weak South Asian summer monsoon (Fig. 10d, e). The significantly warm temperature over the eastern Indian subcontinent in spring, indicates a drier land surface and enhanced solar radiation by a weak monsoon rainfall.

Fig. 11a, referring to the surface temperature anomalies, represents the seasonal evolution of the 850 hPa temperature anomalies averaged over the land area (40°–80°E, 30°–50°N), and the sea area (40°–80°E, 0°–20°N) during the strong monsoon

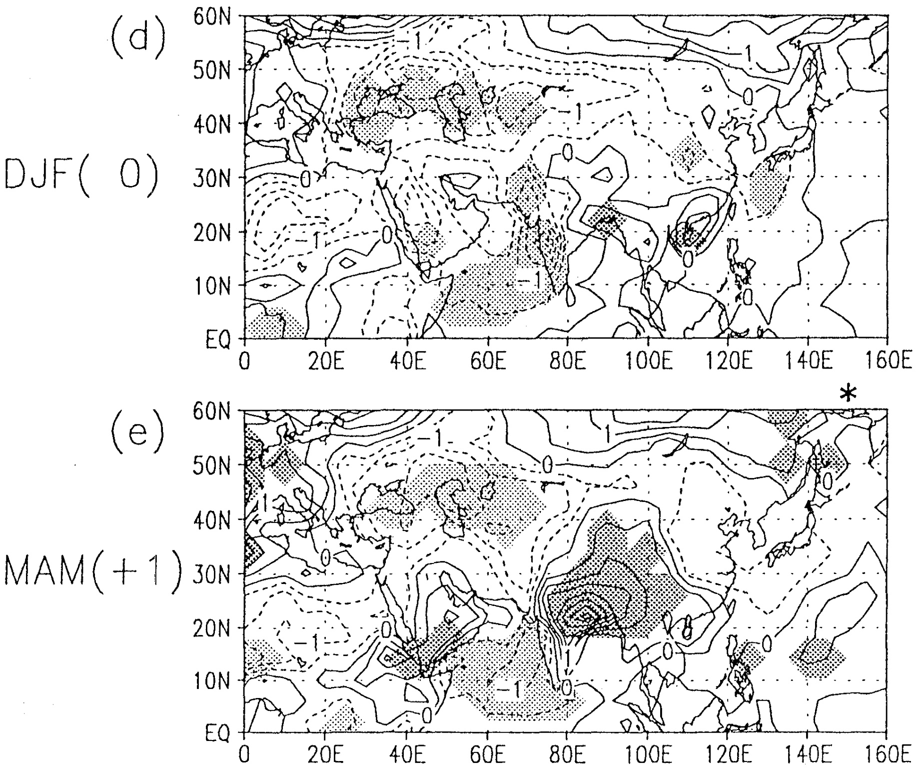

Fig. 10. (Continued)

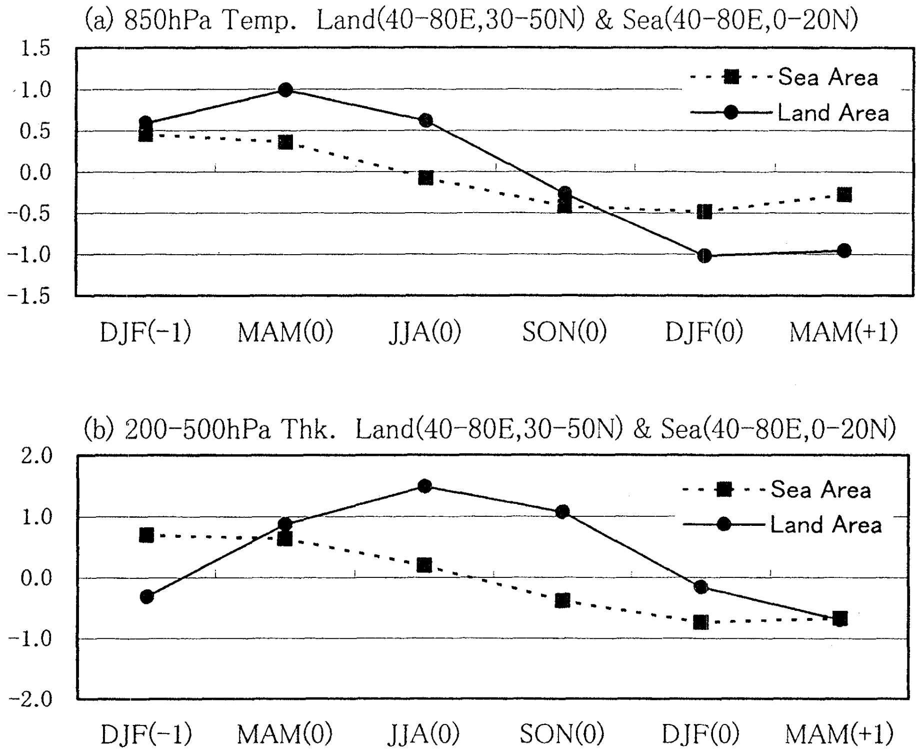

Fig. 11. (a) Regressed seasonal evolution of the 850 hPa temperature anomalies averaged over the land area (40–80E, 30–50N) and the sea area (40–80E, 10–30N), during the winter season (DJF) before the strong monsoon year (noted by "−1") to the spring season (MAM) following the strong monsoon year (noted by "+1"). (b) Same as (a), but for the 200 hPa to 500 hPa thickness anomalies.

year. It is seen that over the land area, the temperature anomaly is already positive in the winter season prior to the strong monsoon. At the same time, the temperature anomaly is also warm over the sea area. In spring, it becomes warmer rapidly over the land area than the sea area, and establishes a stronger land-sea temperature gradient that is favorable to set up a stronger monsoon. After the strong South Asian monsoon, the temperature anomaly over the sea area turns its sign to negative, associated with the negative SST anomaly there (Fig. 5c, d), in the northern autumn. In the following winter to spring, the temperature anomaly over the land area exhibits a rapid cooling than the sea area, to set up a weak land-sea thermal contrast and then, a weak South Asian monsoon. Figure 11b is the seasonal evolution of the 200–500 hPa (upper troposphere) thickness anomaly averaged over the same area. The upper tropospheric temperature contrast between the land and sea to exhibit an anomalously strong monsoon is apparent in northern summer season, which lags that of the lower troposphere by about one season. These results indicate that the circulation anomaly in the winter to spring over the South Asia brings warm temperature advection over the land, and forms a tropospheric land-sea thermal contrast to realize a stronger South Asian monsoon.

It should be noted that in our model, there is no significant snow mass anomaly over the Eurasian continent at the 95 % significant level in relation to the TBO of the South Asian monsoon. Although not statistically significant, there is a negative snow mass anomaly in central Asia in the spring prior to the strong summer monsoon (not shown). This may be due to earlier snow melt in the model climatology over the Eurasian continent by about a month as described in Kitoh *et al.* (1999b). We note here, from our results, that the excessive snow mass is not the only condition, but one of the preferred conditions to affect the monsoon strength. It should also be noted that for bringing the temperature advection around Central Asia, there also seems to be a contribution from the middle latitude circulation anomaly around the Caspian Sea to Europe (Fig. 9). To see the anomalous middle latitude circulation, the regressed 500 hPa height in northern winter (DJF), after a strong monsoon season, is shown in Fig. 12. The area of positive height anomalies around the Europe corresponds well to the anomalous surface wind distribution (Fig. 9b), indicating a barotropic structure. Although not statistically significant, the overall distribution, that is positive height around the Europe and the central Atlantic and negative one around the Central Asia and northern Atlantic, does not contradict the observed distribution of middle latitude circulation anomaly associated with the monsoon rainfall (Yasunari and Seki, 1992). Yasunari and Seki (1992) suggest a possibility of the chaotic behavior of middle latitude circulation to modify the tropical TBO system through the snowmass and temperature anomalies. The possibility of some effects from the anomalous middle latitude circulation may remain, although our results emphasize the role of Matsuno-Gill type circulation anomalies around South Asia in the winter to spring season as a key factor in the TBO mechanism.

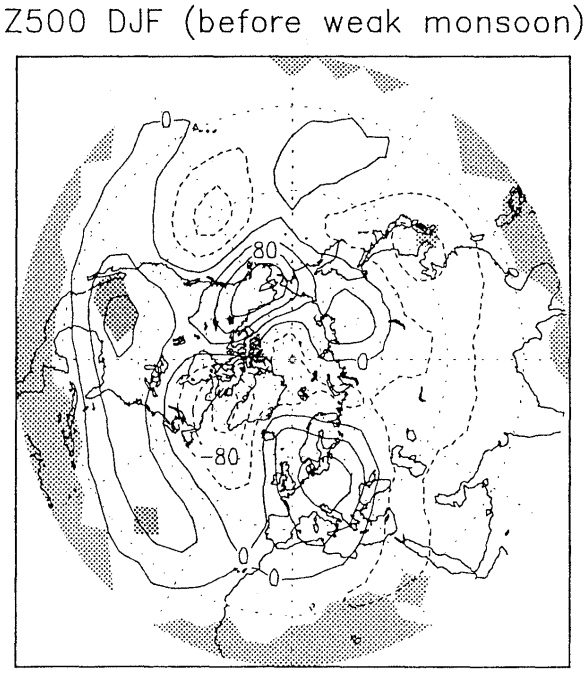

Fig. 12. Regressed 500 hPa height anomalies in DJF after a strong monsoon season (*i.e.*, before a weak monsoon season). Contour interval is 40 m. Shaded area indicates where is statistically significant at the 95 % level.

6. Discussion

Considering the results mentioned above, a schematic mechanism of the biennial variability of the South Asian monsoon could be constructed. In the mechanism, the South Asian monsoon variability itself, and the seasonal migration of the anomalous monsoon convection during the northern summer to winter with its climatological seasonal cycle, play an essential role to keep up the TBO cycle through the land-atmosphere-ocean coupled processes. The seasonal evolution of the TBO is illustrated in Fig. 13 during the strong monsoon year to the weak monsoon year.

In the northern summer of the strong monsoon year (JJA-S), anomalous monsoon westerly flow appears over the Arabian Sea. At the same time, noticeable easterly anomalous flow extends from the central equatorial Pacific to the Bay of Bengal, contributing to the stronger South Asian monsoon precipitation. The SST and convective activity over the

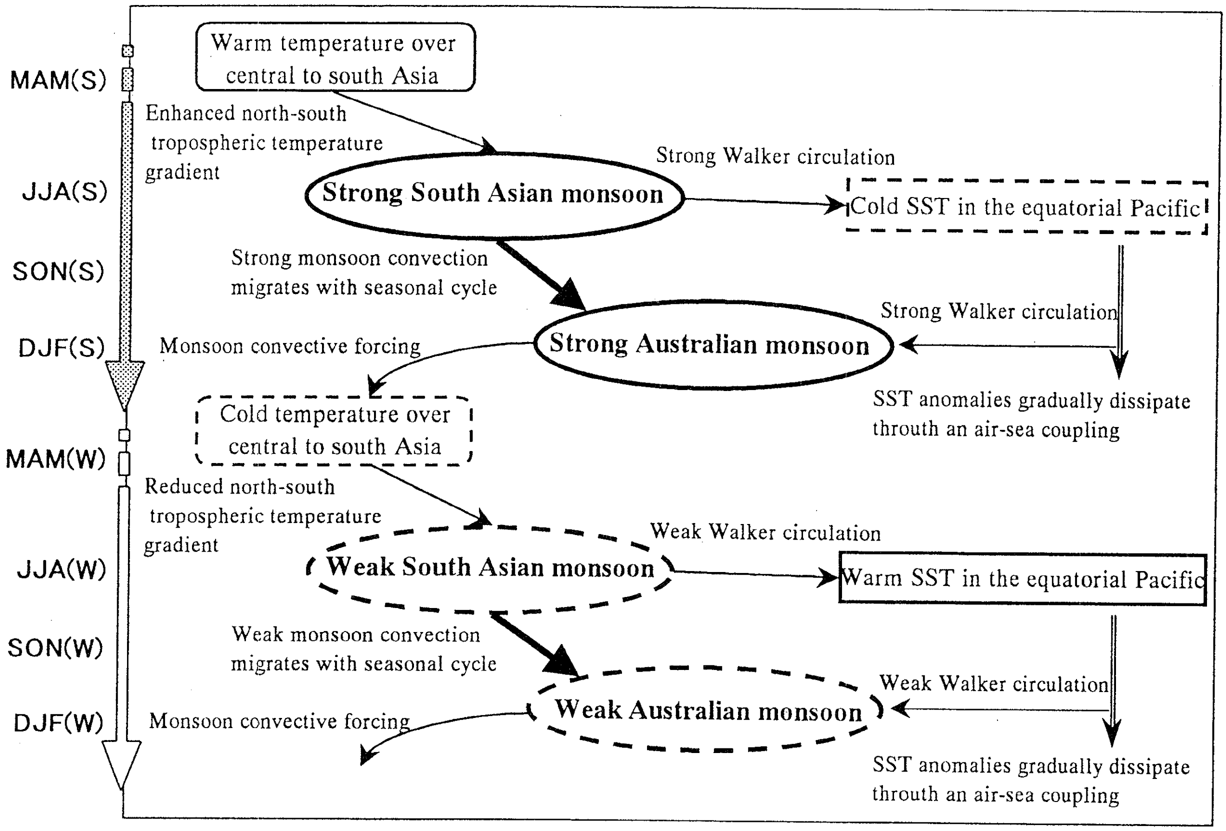

Fig. 13. Schematic diagram illustrating the seasonal evolution of the TBO cycle during the strong monsoon year (noted by S on y-axis) to the weak monsoon year (noted by W on y-axis). Arrows in the figure indicate the interactive processes that keep up the biennial signals.

central Pacific is below normal like a La Niña condition. The large scale stronger east–west Walker circulation is formed between South Asia and the equatorial Pacific, which indicates the close relationship between them.

In the following autumn (SON-S) to winter (DJF-S) season, the area of relatively strong convective maximum moves southeastward from South Asia to around Indonesia. This movement superimposes on the climatological seasonal cycle, as described by the observational study of Meehl (1987). Associated with this seasonal migration, the axis of the anomalous westerly flow over South Asia also moves southeastward, to spread over the entire South Asia. Observational results of the eastward propagating feature of the tropical circulation anomaly in the biennial range (*e.g.*, Yasunari, 1985; Kawamura, 1988) may be explained as another aspect of this seasonal migration of the biennial convection anomaly, from South Asia in summer, to Australia in winter.

The anomalous state of the east–west Walker circulation also persists during this season. In the central to eastern equatorial Pacific, a westerly anomalous wind spreads due to the secondary circulation, which corresponds to downward motion at the eastern periphery of the Walker circulation. The westerly anomalous flow weakens the climatological easterly trade wind. Then the reduced surface latent heat flux and the weakened upwelling motion by Ekman pumping or reduced upper ocean mixing, in addition to the increased shortwave radiation due to reduced cloudiness, leads to warm the SST. This atmosphere-SST negative feedback in the Pacific atmosphere-ocean coupled system may dissipate the TBO variability by the end of winter season. On the other hand, continuous easterly anomalous flow over the equatorial western Pacific to the area of relatively strong monsoon convection, besides these from the Indian Ocean, seems to contribute preserving the monsoon convection activity and seasonal movement of it during the summer to the winter season.

Forced by the monsoon convection anomaly in the northern winter to spring season, anomalous upper (lower) level anticyclonic (cyclonic) circulation over South Asia is formed as a stationary Rossby wave responding to the near equatorial monsoon heating. The persistent lower cyclonic circulation anomaly during the season is a favorable condition to cool the temperature over land through the cold air advection and more snow mass. Cold temperature around the South to Central Asia reduces the tropospheric north- south temperature contrast and then, sets up a weak South Asian monsoon the following spring. Relatively cool SST over the Arabian Sea may also contribute to maintain a weak monsoon by the reduced evaporation and convection. Then the sequence of the weak monsoon year starts as same as strong year's one, but its anomalous state is reversed.

From the result mentioned above, we see two noticeable features in the TBO mechanism. One is the interactive process between the monsoon and the air-sea coupled system in the Pacific through Walker circulation. The anomalous SST is formed over the equatorial Pacific, associated with the anomalous South Asian monsoon circulation in summer season. Although its spatial scale reduces gradually, the moisture fluxes from the Pacific to the Indian sector by Walker circulation continues during the following season. Then it contributes to maintain the anomalous monsoon convection, and also contributes the seasonal migration of the anomalous convection during the summer to winter season, so that the Australian monsoon tends to be strong (weak) when the summer South Asian monsoon is strong (weak).

Another process to be noted is the monsoon-middle latitude circulation interaction during the winter to spring season in the Indian sector. The anomalous monsoon convection around Indonesia to northern Australia persists during the whole winter season. Forced by this anomalous monsoon convection, the anomalous circulation is established over South Asia through the Rossby wave propagation. The circulation brings the temperature advection around Central to South Asia, and changes the tropospheric temperature anomaly over land with the opposite state to that of the previous years. Then the next year, South Asian monsoon with its sign reversed follows through its seasonal onset. It should also be noted that the Matsuno-Gill type circulation anomalies apparently affect the monsoon convection around Southeast Asia through modifying the climatological westerly wind around there, then results to form the contrast of precipitation between South Asia and Southeast Asia (Fig. 1, Fig. 4 and Fig. 9). Because the biennial variability of the South Asian monsoon changes its sign in the northern spring with its seasonal onset, a sequence of the tropical-extratropical interaction during the winter to spring should be regarded as an important factor to switch the biennial variability.

Considering the seasonal migration of the areas of anomalous monsoon convection, the monsoon system in the Indian sector should be considered to exhibit biennial variability as a natural course through its annual progression and the interactive processes with the Pacific sector and middle latitude circulations. Previous studies that emphasize the role of tropical-extratropical interaction in the TBO mechanism vaguely explains the cause of the shift of middle latitude circulation, although they note the role of tropical convection anomalies to alter the extratropical circulation in some way. Meehl (1997) suggests the complex combination of convection anomalies over the African region, equatorial central Pacific and Australian region during the winter to spring season to form the favorable condition to shift the extratropical circulation that affects the winter to spring Asian land surface temperature. In our study, there are little convection anomalies over the African region in the winter to spring season, and it suggests that the monsoon convective activity around Indonesia to Australia in northern winter to spring season is the main force to affect the extratropical circulation over Central to South Asia.

The results mentioned above show that an occurrence of the biennial change of the South Asian monsoon variability in the northern spring season is not to be affected directly by the Pacific SST. Rather, monsoon-midlatitude interaction in the Indian sector in winter to spring season seems to be a main factor to establish the biennial variability on the monsoon system through the seasonal onset of South Asian monsoon.

Our results emphasize the monsoon-midlatitude interaction in the Indian sector to regulate the tropical climate through the biennial variability of the tropical air-sea coupled system. As for the longer time scale variability, Kitoh *et al.* (1999b) discusses the monsoon-ENSO relationship using the MRI coupled GCM. They note that the Indian summer monsoon rainfall variability seems to be affected by the tropical Pacific SST in the ENSO time scale (4 to 8 years). Discussion on the relative importance of the biennially oscillating land-atmosphere-ocean system in the Indian sector and the ENSO for determining the South Asian monsoon variability is beyond the scope of this paper. This problem should be investigated with observational data, or sensitivity experiments using the coupled GCM.

7. Summary and conclusion

In this paper we have investigated the mechanism of the tropospheric biennial oscillation of the South Asian monsoon, as simulated by the MRI coupled GCM. Our results strongly suggest that the biennial variability of the South Asian monsoon occur through the interactive processes with the Pacific sector or middle latitude circulation. Each of the processes largely depends on a course of seasonal march of the monsoon convection anomaly, which moves from South Asia to Australia during northern summer to winter season. As noticed by Yasunari (1990) and Meehl (1994) the strong (weak) phase of the TBO could be considered to start from the northern spring, and persists until the following winter season. As an essential factor to maintain the TBO cycle, we raise here the two types of interactive processes; one is between the monsoon and the tropical Pacific in northern summer to winter, and the other is between the monsoon and the middle latitude during the northern winter to spring season.

The monsoon-tropical Pacific interaction is realized through the anomalous state of the east–west Walker circulation during the northern summer to winter season. In the summer of a strong TBO year, when the South Asian convection anomaly reaches its maximum, SST is below normal and convective activity is weak over the equatorial Pacific region. This corresponds to a La Niña type condition, and thus indicates a stronger Walker circulation. Although the upper divergence/convergence of Walker circulation reduces its spatial extent with a progress of seasonal cycle, the anomalous state persists through the whole summer to winter season over the Western Pacific. Persistent moisture fluxes from the equatorial Pacific to the area of anomalous monsoon convection due to stronger Walker circulation, in addition to those from the South Asian region, may contribute to keep the activity of strong monsoon convection until the winter season. In the Pacific sector, the downward motion at the eastern periphery of an anomalous Walker circulation brings a westerly anomalous flow to reduce the climatological easterly trade wind in autumn to winter. The reduced surface latent heat flux and weakened equatorial upwelling current and vertical mixing at the oceanic mixed layer, caused by the weaker trade wind, in addition to the increased shortwave radiation due to the reduced cloudiness, may result in warming the SST there. By this local atmosphere-ocean coupling, the SST anomaly formed in the summer season decays gradually from its eastern edge and the center of the anomalously warm SST apparently moves westward during the autumn to the winter season. Convection and SST anomalies over the equatorial Pacific almost dissipate in the following spring. Because none of the anomalous state is seen over the Pacific region in springtime, when the abrupt phase change of the Asian monsoon occurs, the other components to "directly" affect the South Asian monsoon strength should be noted.

The monsoon-middle latitude interaction occurs mainly during the northern winter and the following spring season. In the northern autumn of a strong monsoon year, the center of anomalously strong monsoon convection starts to move southeastward around Indonesia to Australia along with a seasonal cycle, keeping its strength through the whole winter season. As a response to this tropical convection anomaly, upper level anticyclonic circulation and lower level cyclonic circulation is formed over South Asia. This is explained as the formation of a stationary Rossby wave responding to the remote forcing of the monsoon convection. The lower cyclonic circulation over South Asia in the winter to spring season is a favorable condition to keep the land surface cool around the Middle East and Central Asia by northerly cold air advection, and lead to form the cooled tropospheric temperature there. The reduced land-sea, or north-south tropospheric temperature contrast in spring leads to a following weak South Asian monsoon. The snow mass is also above normal around Central Asia, although it is not significant at the 95 % level. So our result suggests that the snow mass anomaly is one of the condition to affect the monsoon variability, but not the necessary condition. The circulation also modifies the climatological westerly flow over Southeast Asia, and results to form the contrast of monsoon precipitation between South Asia and the Southeast Asia. Then the monsoon-middle latitude interaction in the biennial range is regarded as the internal feedback process. A stronger monsoon convection results in a weak one in the next year, through the weakened land-sea thermal contrast that is established by an anomalous circulation driven by a wintertime stronger monsoon convection itself. These results imply the importance of the tropical-extratropical interaction to maintain the TBO cycle.

Our result strongly suggests the important role of South Asian monsoons to keep the TBO cycle through the interaction with the Pacific region, or the middle latitude. This point of view is roughly the same as noted by Yasunari and Seki (1992) or Meehl (1994, 1997). However, it should be noted that the mechanism of the TBO discussed here necessarily requires the combination of tropical convection anomalies to shift the middle latitude circulation, that realizes the tropical-middle latitude interaction. We emphasize the importance of the seasonal migration of monsoon convection anomaly itself during the northern summer to winter season, and its role to directly affect the wintertime circulation around South Asia through the Rossby wave response. Because the scenario of the TBO described here is based on the simulated monsoon variability by a coupled GCM, the cycle should be compared with the observational data. But we believe that our results would have some implication on the mechanism of the TBO, emphasizing the active role of the monsoon circulation in the whole sequence of convective season from the northern summer over South Asia, and the northern winter around Australia.

Although the mechanism of the TBO discussed above closes its cycle completely in two years, the TBO seems to be not completely biennial (Barnett, 1991), and it does not occur in every year. Certainly, the South Asian monsoon shows the longer variability associated with the Pacific SST such as the ENSO. Discussing the relationship between the TBO and the ENSO is beyond the scope of this study, but it could be said that the other scale of variabilities like the ENSO may occasionally affect the TBO cycle. The amplitude of the TBO cycle may be modulated by the chaotic behavior of the winter time middle latitude circulation as proposed by Yasunari and Seki (1992). Searching for other processes which affect the TBO cycle remains in the future.

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気象研究所大気海洋結合モデルで再現された ENSO—モンスーン系の対流圏2年周期振動

小笠原範光[1]
(筑波大学地球科学研究科)
鬼頭昭雄
(気象研究所)
安成哲三
(筑波大学地球科学系)
野田　彰
(気象研究所)

ENSO—モンスーン系において卓越する2年周期振動のメカニズムを、気象研究所全球大気・海洋結合モデルを用いて調べた。2年周期振動において、アジアモンスーンは熱帯太平洋の大気海洋結合系と、中緯度循環場との相互作用という2つのプロセスを通して中心的な役割を果たしている可能性が確認された。

モンスーンが強い年の春から夏の南アジアで相対的に活発な対流活動域は、秋から冬にかけてインドネシア付近へ南東進していく。この動きは気候値に見られる熱帯域での対流活動活発域の季節変化と良く一致しており、南アジアからオーストラリアのモンスーンが一連に強い状態となる事を示唆している。夏期に南アジアと熱帯太平洋間の東西循環偏差を通して形成された、熱帯太平洋における海面水温や対流活動偏差は、冬までの間に解消される傾向にある。

冬から春にかけての南アジア付近では、冬の間持続されるオーストラリアモンスーンに伴う対流活動偏差を熱源として、松野-Gillタイプの定常ロスビー波が形成される。モンスーンが強い年には南アジア上空で高気圧性循環、下層で低気圧性循環ができるため、寒気移流がより南まで入りやすい条件となっている。寒気移流は翌春まで持続し、これに伴い地上気温や上空の気温偏差は、春には中央アジアや南アジアで有意に冷たくなっている。このため大陸—海洋間の南北温度コントラストが弱められ、アジアモンスーンの強い年の翌年には逆にモンスーンは弱められると解釈できる。2年周期振動が年サイクルに見かけ上位相固定している理由は、春先の南アジアモンスーンの開始を通した、中緯度−熱帯間の相互作用が本質的な役割を果たしているためであると考えられる。モデル結果は、2年周期振動が陸面−モンスーン−海洋結合系特有の変動であることを示唆しており、冬から翌春にかけてもたらされる、モンスーンと中緯度との相互作用が2年周期振動の維持において果たす役割を強調している。

1 現所属：(財) 日本気象協会東海本部