ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 25, NO. 5, 2008, 815–823

Spatial Patterns of Tropospheric Biennial Oscillation and Its Numerical Simulation

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(Received 10 May 2007; revised 18 December 2007)

ABSTRACT

In order to investigate the spatial patterns of the Tropospheric Biennial Oscillation (TBO) on the global scale, the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) monthly averaged precipitation and the Climate Diagnostics Center (CDC) monthly outgoing long-wave radiation (OLR) and SST are used in conjunction with TBO bandpass-filtering. The results indicate active biennial variability in the tropical eastern-central Pacific regions. It is evident that observations reflect the biennial component of the ENSO rather than the TBO itself. Since some studies have pointed out that the TBO is a broad-scale phenomenon differing from the ENSO, to investigate the pure TBO the ENSO signal must be excluded. The Scale Interaction Experiment-FRCGC (SINTEX-F) coupled general circulation model (CGCM) developed at Japan Frontier Research Center for Global Change (FRCGC) can capture both the ENSO and the biennial signals. Air-sea interactions in the tropical eastern-central Pacific are decoupled to eliminate the effects of ENSO in a experiment by SINTEX-F and the results show that biennial variability still exists even without ENSO. It seems to mean that the TBO and ENSO are independent from each other. Furthermore, the model results indicate that the two key regions are southwest Sumatra and the tropical western Pacific for the TBO cycle.

Key words: tropospheric biennial oscillation, air-sea interaction, spatial pattern

Citation: Zheng, B., D. J. Gu, A. L. Lin, and C. H. Li, 2008: Spatial patterns of tropospheric biennial oscillation and its numerical simulation. *Adv. Atmos. Sci.*, **25**(5), 815–823, doi: 10.1007/s00376-008-0815-9.

1. Introduction

Many studies had showed that the biennial variability is a very significant feature for the atmospheric circulations and interannual climate changes. In the 1960s, Reed et al. (1961) revealed quasi-biennial oscillations (QBO) by studying the lower tropical stratospheric zonal winds. Theories on the QBO developed quickly and recent studies have been concerned with the tracers of QBO and its numerical simulations (Zheng et al., 2003, 2006; Chen et al., 2005). Whereas, because of the strong signal in ENSO, tropospheric biennial oscillation (TBO) had already been neglected for a long time though the biennial variability for some variables in the troposphere had been investigated in the late 1960s and early 1970s (Angell and Korshover, 1968; Trenberth, 1975). The TBO was not a focus until the 1980s. Many variables, such as surface pressure over the Northern Hemisphere (NH) (Trenberth and Shin, 1984), NH averaged surface temperature (Ding et al., 2001) and tropical zonal winds (Rasmusson et al., 1990), had been revealed by the biennial variability. Furthermore, Asia-Pacific monsoon rainfall also exhibits pronounced biennial features [e.g., India summer rainfall (Mooley and Parthasarathy, 1983; Mukherjee et al., 1985) and East Asia rainfall (Tian and Yasunari, 1992; Shen and Lau, 1995)]. Interannual monsoon rainfall in China to a great extent behaves as a shift of the rainfall bands (Chen and Song, 1997; Liao and Wang, 1998; Kuang et al., 2002), which is different from the Indian monsoon rainfall. TBO is a strong quasi-periodic phenomenon on a large scale in the troposphere, second in strenght only to ENSO. Together with ENSO, the TBO affects atmospheric circulations and leads to weather and climate anomalies. Lau and Wu (2001) have identified three coherent modes of

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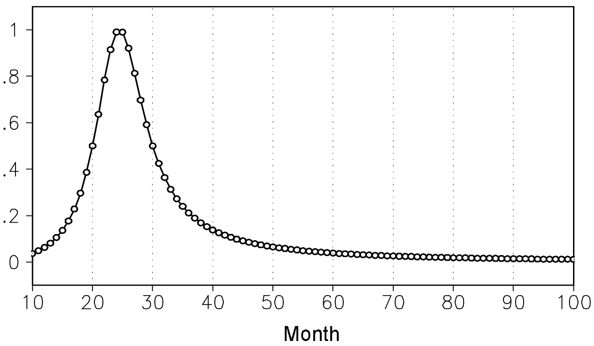

Fig. 1. Response function for the used bandpass filter.

summertime rainfall variability over China and global SST for the period of 1955–98 by Singular Value Decomposition (SVD). They found that the second mode is comprised of a quasi-biennial variability manifested in alternate wet and dry years over the Yangtze River valley and the severe flood over the Yangtze River valley in 1998 is associated with the biennial tendency of basin-scale SST anomalies during the transition from El Niño to La Niña in 1997–98. Subsequently, Lau and Wu (2001) investigated the covariability of the Asian summer monsoon and ENSO using global rainfall and SST data for the past two decades (1979–98) and found the first mode is characterized by a pronounced biennial variability. Hence, studies on the TBO are helpful for developing the skill of short-term climate prediction.

Many scientists consider the TBO to be attributed to ENSO (Yasunari, 1988; Li et al., 2001a; Zou et al., 2002; Fasullo, 2004). In contrast, some studies indicated that the TBO is an independent air-sea coupling system (Chang and Li, 2000; Li et al., 2001b; Li et al., 2006). Using a 5-box model, Chang and Li (2000) firstly pointed out clearly the TBO is a result of interactions among monsoon cycles and the tropical Indian Ocean and Pacific Ocean and that it is different from the ENSO air-sea system. Thus, only using observations would make it very difficult to distinguish them and to study the "pure-TBO". The TBO theory is still in development (Zheng and Liang, 2005), but even its spatial patterns are unclear. In this paper, we first determined the distribution of the biennial variability by observations, and then explored the TBO activity centers using a coupled model.

2. Data and methods

The Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) monthly averaged precipitation and the Climate Diagnostics Center (CDC) monthly outgoing long-wave radiation (OLR) and SST are used in this paper. The precipitation and OLR have a horizontal resolution of $2.5° \times 2.5°$ and SST for $2° \times 2°$. All the data span the period from January 1979 to December 2003. Meanwhile, 20-year model data from the Scale Interaction Experiment-FRCGC (SINTEX-F) coupled general circulation model (CGCM) developed at Japan Frontier Research Center for Global Change are also used to investigate the TBO. By means of bandpass filtering and variance analysis, the distribution of the quasi-biennial component is obtained by the observational and model data and the TBO activity centers are found using the model data.

3. Global distribution of the tropospheric quasi-biennial oscillation

A bandpass filter, which has a complete response in the period of 24.5-month and half response periods by 20- and 30-month (as showed in Fig. 1), is used to obtain the quasi-biennial component of precipitation and SST. Then the ratios of variances of the quasi-biennial component to the total variances can be obtained as well (Fig. 2). Since many results of quasi-biennial variability in the past came from the precipitation in the monsoon regions, it gave us the misleading concept that the quasi-biennial component has larger variations in the monsoon continent (e.g. India or East Asia continent) than elsewhere. But it can be seen from Fig. 2 that this is not true.

Figure 2 shows the quasi-biennial variability is most active in the equatorial eastern-central Pacific Ocean. If calculating the ratios of variances of the quasi-biennial component relative to the total interannual variability instead of the total variance, a relative strong quasi-biennial variability would appear in India and China (see Li et al., 2006). Since seasonal variability absolutely dominates the monsoon regions, it is reasonable that there are small ratios in the same region of monsoon in comparison to the results of Li et al. (2006). Due to the strong ENSO signal located in the equatorial eastern-central Pacific, we generally consider the quasi-biennial oscillation to be a component of ENSO (further discussion will follow). There are also obvious quasi-biennial variabilities in the western Pacific (WP) and the southeast Indian Ocean (SEIO) off Sumatra. Although the quasi-biennial variability is manifested in some extratropical and mid- to high-latitude regions, they are not related to the main area of concern.

In brief, the global quasi-biennial variability is mainly active in the tropical oceans. This implies the quasi-biennial variability is produced by air-sea interactions. The quasi-biennial variability in the eastern-central Pacific may be attributed to ENSO, whereas

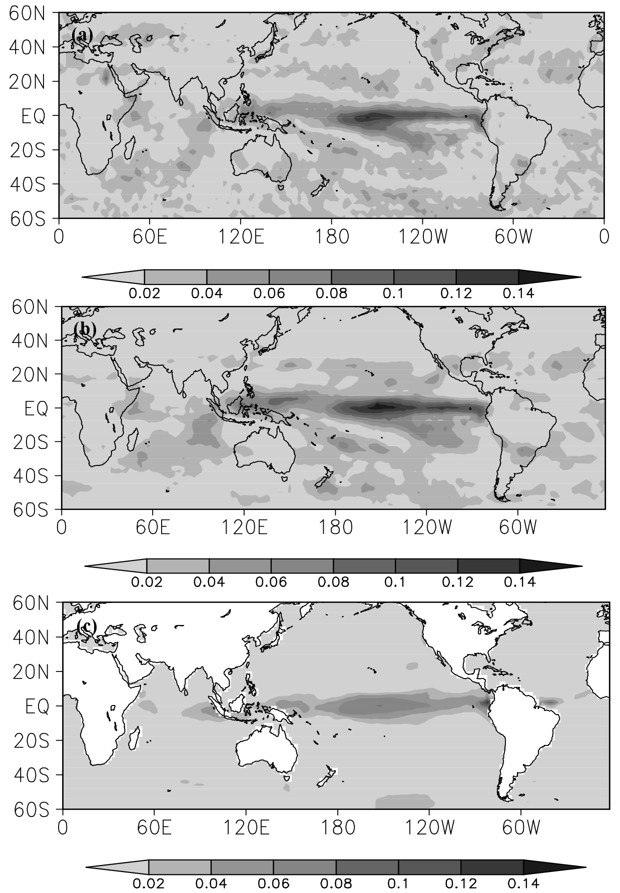

Fig. 2. Ratios of variances of quasi-biennial component of (a) CMAP precipitation, (b) OLR and (c) SST to the total variances during 1979–2003. (The total variance is computed with seasonal variations)

those in the WP and the SEIO seem to be a separate interactive air-sea system (Chang and Li, 2000). The following section will analyze the model data and focus on two issues: one is whether the TBO is independent from the ENSO, and another is how the TBO cycle is maintained.

4. Simulation results

The ocean component of the SINTEX-F CGCM is the reference version 8.2 of Océan Parallélisé—OPA (Madec et al., 1998) with the Option Rom Configuration for Arrays (ORCA) 2 configuration. The model resolution is $2° \times 2°$ with increased meridional resolution (as a cosine function of latitude) of 0.5° near the equator. It has 31 vertical z levels of which 19 lie in the top 400 m. The atmosphere component is the latest version of ECHAM4 (Roeckner et al., 1996) with a horizontal resolution of $1.1° \times 1.1°$. The coupling fields are exchanged every 2 h between the ocean and atmosphere by means of the Ocean Atmosphere Sea Ice Soil (OASIS) 2.4 coupler (Valcke et al., 2000). The coupled model does not apply any flux correction. The SINTEX-F CGCM can reproduce the features of ENSO well, including period, amplitude and spatial pattern. The detailed description of the model can be found in the works of Luo et al. (2003, 2005)

Two groups of model data are used in this paper. One came from the control experiment with global air-sea interactions, while another came from the experiment with air-sea interactions decoupled in the eastern-central Pacific (30°S–30°N, east of 180°E in the Pacific) (ESDEP run). The used model data are 20-year data (11–30 model-year).

4.1 *Results of control experiment*

In general, if a model can not reproduce the climatological seasonal variation well, then the simulated interannual variability will not be reliable. Figure 3 shows the 1979–2003 mean annual cycle of the CMAP precipitation, and Fig. 4 is the simulated 20-year mean seasonal variation. The differences between observations and simulations manifest in January and July. There is little observed precipitation in the southern portions of the equatorial eastern Pacific in January (Fig. 3a), however overestimated simulated precipitation appear there (Fig. 4a). In July, the model can not capture the precipitation in the mid-latitude Pacific and Atlantic Oceans and underestimates the rainfall over the East Asia monsoon region (Figs. 3c, and 4c). Although some differences exist between the observations and simulations, the simulated precipitation has similar large-scale spatial patterns and a reasonable seasonal match to the observations. The model simulates relatively realistic seasonal SST variation as well (Luo et al., 2005, figure omitted).

Figures 5a and 5b show the ratios of variances of the quasi-biennial component of simulated precipitation and SST to the total variances in the control experiment, respectively. Similar to the results of the observations (Figs. 2a and 2c), simulated quasi-biennial variability is mainly active in the tropical oceans.

4.2 *Results of ESDEP experiment*

In this run, the atmosphere reacts to climatological SST from the CTL run (not observed SSTs) east of 180° in the tropical Pacific, but SST of the ocean model there is allowed to evolve according to the atmospheric forcing.

Since the SINTEX-F CGCM can capture the ENSO (Luo et al., 2005) and quasi-biennial signals (see Fig. 5), we are still uncertain of whether or not the TBO is a quasi-biennial component of ENSO. In order to analyze the TBO we should exclude the ENSO sig-

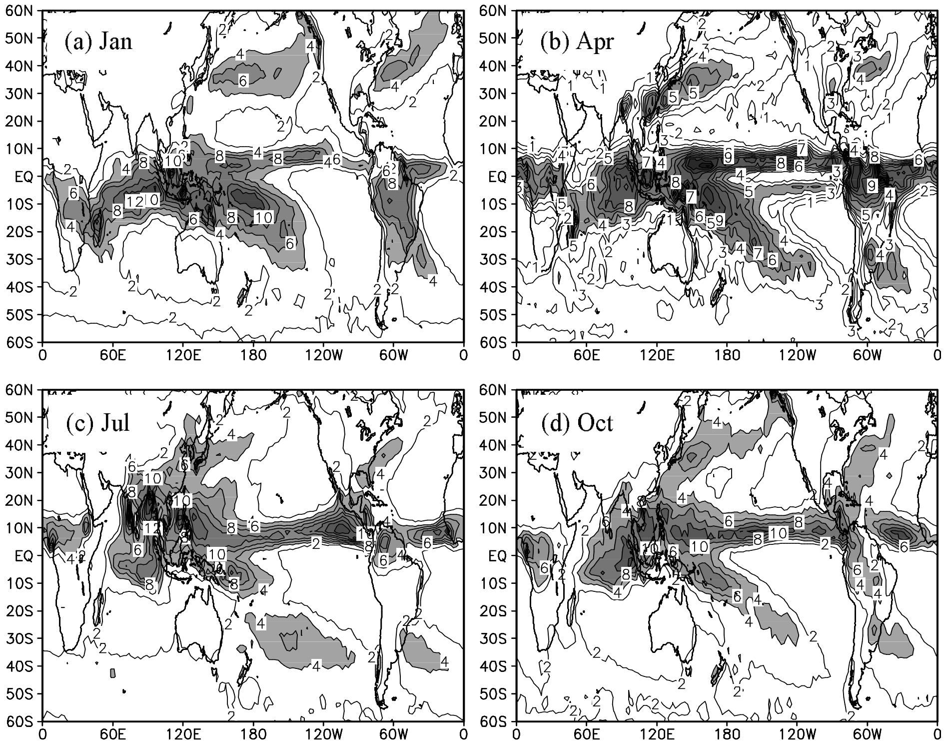

Fig. 3. 1979–2003 mean CMAP precipitation (mm d^{-1}) in (a) January, (b) April, (c) July, and (d) October.

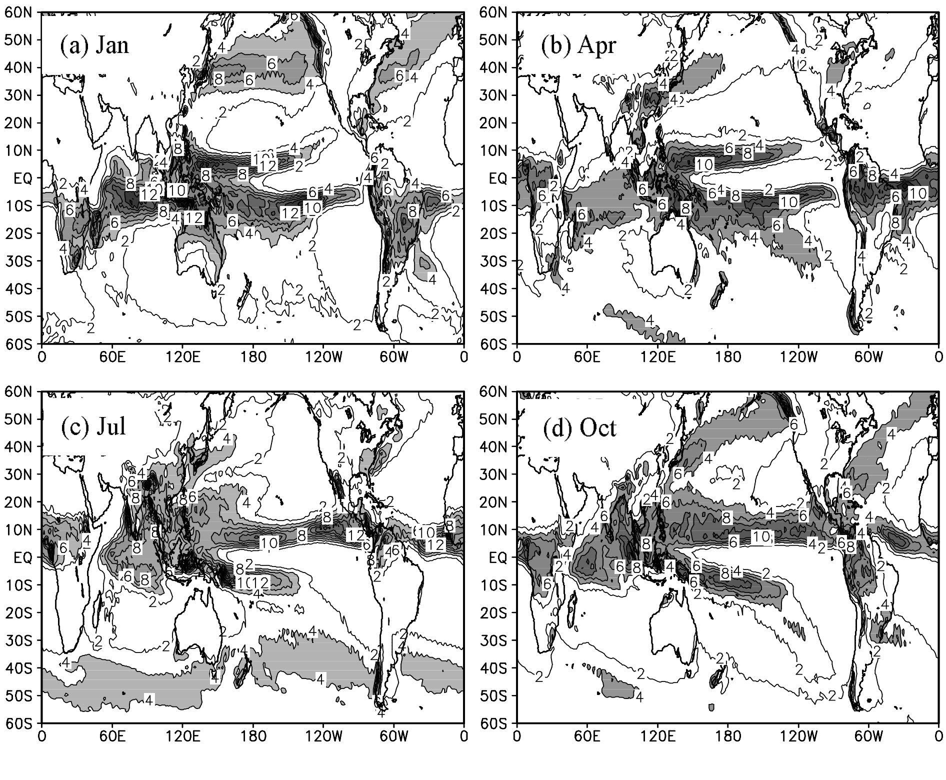

Fig. 4. 20-year mean simulated precipitation by control experiment (mm d^{-1}) in (a) January, (b) April, (c) July, and (d) October.

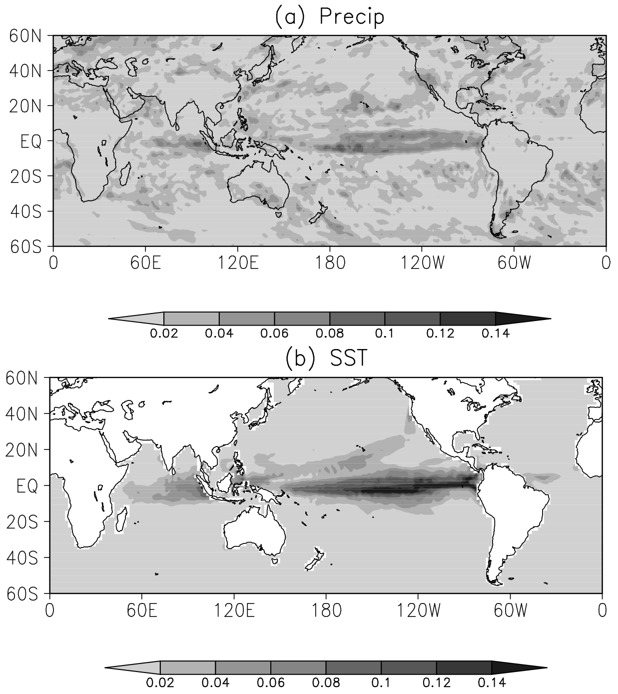

Fig. 5. Ratios of variances of the quasi-biennial component of simulated (a) precipitation and (b) SST to the total variances by control experiment. (The total variance is computed with seasonal variations)

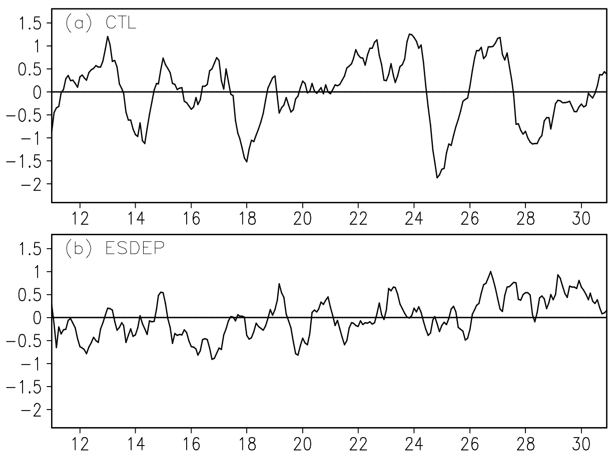

Fig. 6. The Niño3.4 index (unit: K) from (a) control run and (b) ESDEP run with variance of 0.455 K^2 and 0.178 K^2, respectively.

nal. The ESDEP experiment, with decoupled air-sea interactions in the tropical eastern-central Pacific since they are critical for creating ENSO (Zebiak and Cane, 1987; Wang et al., 1995), is designed to exclude the effects of ENSO. Figure 6 shows the Niño3.4 index from the CTL run and ESDEP run. There is much more significant interannual variability in the CTL run than in the ESDEP run, which confirms that ENSO is excluded in the ESDEP experiment.

Figure 7a shows the largest strength of the precipitation TBO in the SEIO off Sumatra, the same as Figs. 2a and 5a. There is another strength center of precipitation TBO in the Arabian Sea that is not showed in Fig. 2a and is relatively weak in Fig. 5a. We are not sure what role the Arabian Sea plays in the TBO or if this center is simply an error by model. In comparison with Figs. 2a and 5a, the most differences exist in the eastern-central Pacific where the active quasi-biennial components become much weaker.

It can be seen from Fig. 7b that the SST TBO in the WP and SEIO are as significant as shown in Figs. 2c and 5b. Contrary to the control experiment, the SST TBO in the eastern-central Pacific is suppressed but still has moderate strength. Since ENSO is excluded, the SST TBO in the eastern-central Pacific may be a variability forced remotely by the TBO in the warm ocean which is likely a key region for producing the TBO. Another possible cause of the TBO in the eastern-central Pacific is that the TBO is the internal mode of the ocean. Namely, oceans can produce the biennial variability by itself.

It is noticeable that the SST TBO in the eastern-central Pacific is moderate but the precipitation TBO almost disappears. Since the atmosphere only reacts to climatological SST in the eastern-central Pacific in the ESDEP run, it is reasonable that the TBO nearly disappears in the corresponding precipitation anomalies in the cold ocean. This suggests that the biennial variability is likely not an internal mode of the atmo-

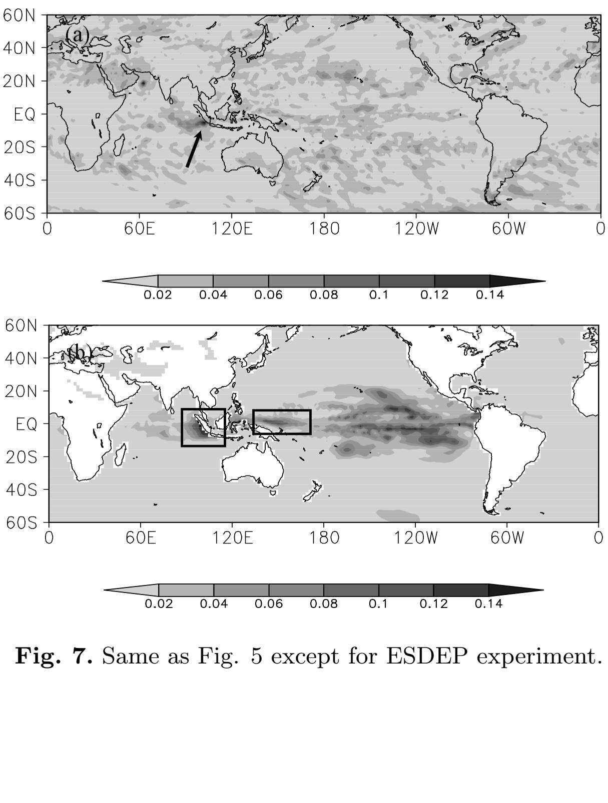

Fig. 7. Same as Fig. 5 except for ESDEP experiment.

sphere. Also, Fig. 6 shows that the Niño3.4 SSTA variance without air-sea interaction is much less than the full coupled one, and the calculated variance of the biennial component in the eastern-central Pacific in the ESDEP run is also reduced by 74.3% in comparison with that in the CTL run. On the other hand, the variance of the biennial variability in the warm pool ocean (5°S–5°N, 130°–160°E) in the ESDEP run decreases just 37.5% relative to that in the CTL run. Since the cold ocean can still be forced by the biennial variation of the atmosphere and can also vary by itself combined with the fact that the warm ocean is coupled with the atmosphere, the reductions of the biennial component in the ESDEP indicate that the TBO is a coupled air-sea system mode rather than an internal mode of ocean. We also calculated the ratio of the quasi-biennial component relative to the total variance with the seasonal cycle removed in the warm pool ocean. The ratio from the CTL run is 72.7% and that from the ESDEP run is 83.3%. It suggests that the TBO has the most dominant variability after the seasonal cycle and ENSO. Additionally, the ratio of the quasi-biennial component from the ESDEP relative to the CTL run is 62.5% in the warm pool ocean. This indicates the independent mode of the TBO separate from the ENSO mode.

5. TBO cycle

The analyses above confirm that the TBO is an independent system, but it is unclear how the TBO cycle is established and maintained. Figure 8 show the lagged correlation of the quasi-biennial component of the ESDEP July SST averaged over 5°S–5°N, 130°–160°E (see the right box in Fig. 7b) with surface winds and SSTAs.

In July of year 0 (a reference year), anomalous warm SSTs in the WP excite anomalous cyclones over the western North Pacific (WNP), Southeast Asia (SEA) and to the Northeast of Australia. The anomalous cyclone over the South SEA (SSEA) and to the Northeast of Australia enhances the cross-equatorial flows over the SEIO and the Maritime Continent (MC). This also cools the SST there due to wind-evaporation feedback. Cold MC SSTs then excite an anomalous anticyclonic circulation over the South Indian Ocean (SIO). Because of the bias of the SSTA over the MC in the Southern Hemisphere, the excited Rossby wave is rather weak over the NH. The anomalous anticyclone over the SIO also intensifies the cross-equatorial flows over the MC, so the anomalous anticyclonic circulation and SSTA in the MC and SEIO can be self-maintained (the detailed mechanisms can be seen in Li et al., 2003) even several months after the warm SSTAs in the WP have died out [see Oct(0) and Jan(1) in Fig. 8]. Meanwhile, the anomalous cyclone over the WNP results in the cooling of the WP SSTAs which simultaneously weakens the WNP anomalous cyclone.

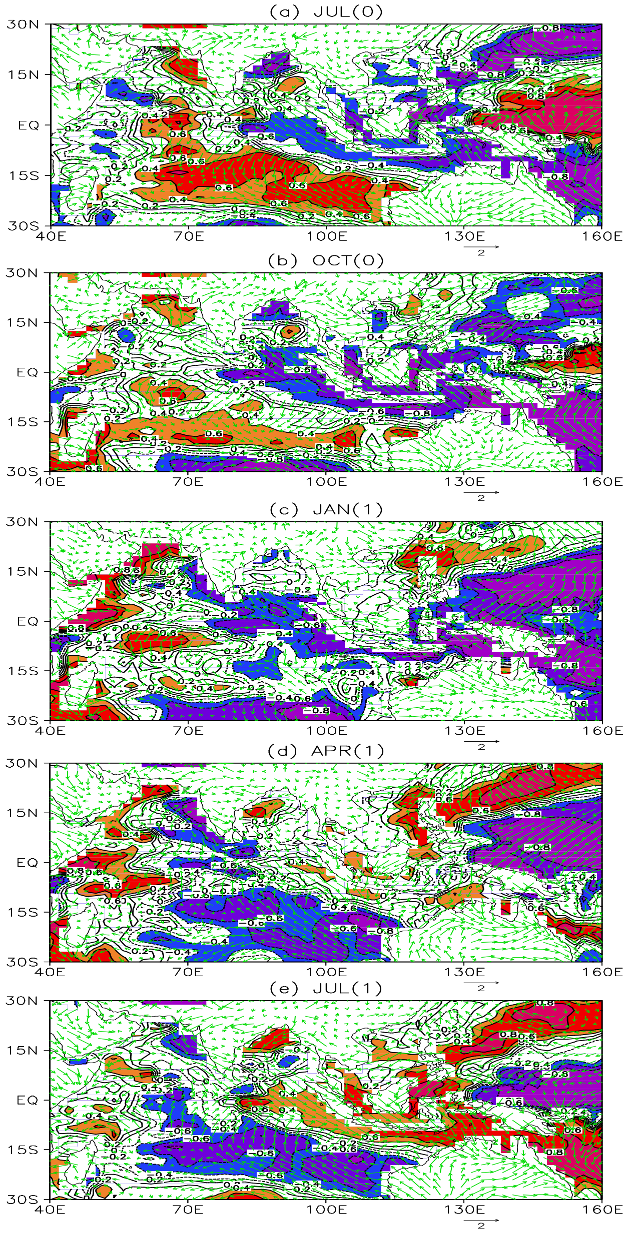

Fig. 8. Lagged correlation of quasi-biennial component of July SST by the ESDEP experiment averaged in 5°S–5°N, 130°–160°E (see the right box in Fig. 7b) with surface winds and SSTAs (Shaded areas denote the correlation coefficients with SSTA exceeding 95% significant level. The meridional component of the vector represents the correlation coefficients with surface meridional wind and the zonal component of the vector with the zonal wind).

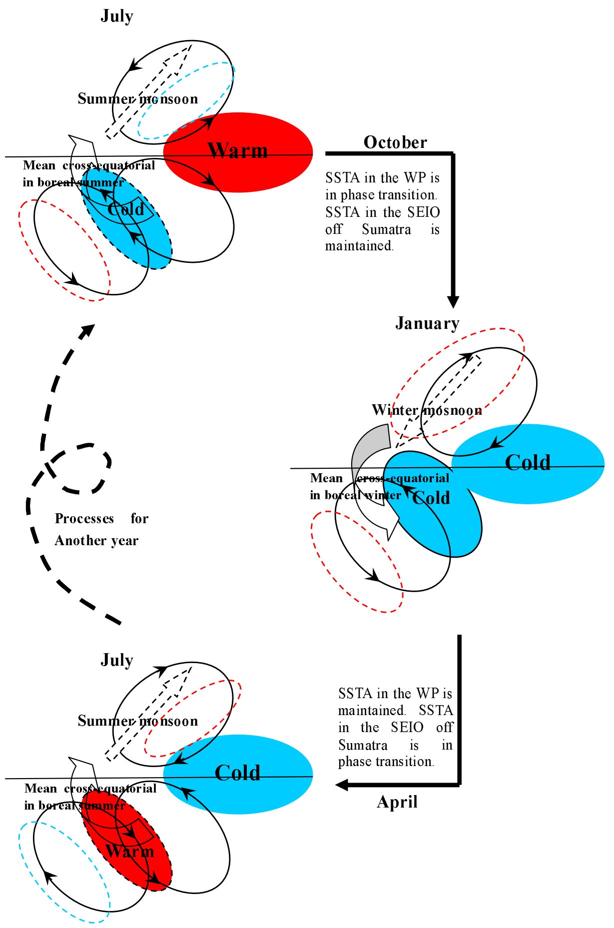

Fig. 9. Sketches for the TBO air-sea interactions in the warm oceans from Fig. 8.

In OCT(0) (October of the reference year), the region of the warm SSTAs over the WP has reduced and the anomalous cyclone over the WNP has almost disappeared. Anomalous cross-equatorial flows blow into the WNP and continue cooling the SST there; they come back from the western equatorial Pacific to cool the SST there as well. More importantly, the cyclonic wind stress curl in the WP also cools the SST.

In JAN(1) (next January relative to the reference year), the WP SSTAs become cold and excite two anomalous anticyclonic circulations west of the WP. The anomalous anticyclone over the WNP (namely Philippine anticyclone, see Wang et al., 2000) can be self-maintained into July with corresponding anomalous SSTAs (see APR(1) and JUL(1) in Fig. 8). With opposite mean cross-equatorial flows (to the north in summer)relative to boreal summer, the anomalous anticyclone over the SIO warms the MC SST and the latter weakens the former in reverse.

In APR(1), the anomalous anticyclone over the SIO has died out, while the anomalous anticyclone over the SSEA and to the Northeast of Australia induced by cold SSTAs in the WP continues warming the SST in the MC. This excites an anomalous cyclone over the SIO though it is still weak.

In JUL(1), the anomalous cyclone over the SIO has been established completely and it can be self-maintained for several months. Meanwhile, the Philippine anticyclone is weakened due to reversed mean background winds.

The processes have finished one phase and another has formed with similar processes. Then the TBO cycle is established. The 40-year model run has also been carried out, and the results indicate a similar cycle (figure omitted). Under idealized conditions, the TBO cycles with a period of exactly 2 years, whereas due to internal factors (e.g., strength of air-sea coupling) and external dynamics (MJO, ENSO and so on) the TBO evolves with an irregular period (Li et al., 2001b).

Figure 9 shows how the TBO is established and maintained. It is notable that the maximum amplitude of the SSTA in the SEIO off Sumatra is in boreal fall and phase transition in northern spring, while the maximum amplitude of the SSTA in the WP is in boreal spring and phase transition in northern autumn. This out-of phase setting is responsible for the maintenance of the TBO.

6. Conclusions and discussions

It is obvious that observational quasi-biennial variability includes the TBO and the quasi-biennial component of ENSO. The quasi-biennial component associated with ENSO is mainly distributed in the eastern-central Pacific and the TBO is located in the warm oceans (the WP, MC and SEIO). The simulations indicate that the TBO is likely an independent air-sea system. Li et al. (2006) have had similar findings regarding the TBO in the Indo-Pacific warm ocean regions, but they considered the WNP monsoon variability and convective activity over the SEA and MC to be two of the key processes. In contrast, we consider the convective activity over the WNP and MC to be nonessential for the TBO cycle. As mentioned above, the TBO is most likely an air-sea self-sustained system in the warm ocean region and convection variability is a result rather than a cause of the TBO. The phase lag between SST in the WP and SEIO causes them to form a cycle and establish the TBO.

Figure 7b implies the TBO may affect ENSO. Li et al. (2006) pointed out that with three "atmospheric bridges" the TBO in the warm ocean region would cause the biennial variability of ENSO. These are the north-south teleconnection that connects the WNP monsoon and the SEIO, the east-west teleconnection that connects the Indian Ocean and the Pacific cold tongue and the El Niño-WNP monsoon teleconnection. It is noticeable that a rainfall belt with quasi-biennial variability exists over the North Pacific. The belt along the tropical eastern Pacific into East Asia is always located there no matter whether ENSO exists or not (see Figs. 2a, 5a and 6a). Thus, the quasi-biennial component of the East Asia precipitation should be attributed to the TBO rather than ENSO, but it is unclear how the TBO affects the precipitation over East Asia.

Acknowledgements. The authors would like to express their thanks to the Climate Prediction Center (CPC) and the Climate Diagnostics Center (CDC) for providing good data products. We also appreciate the comments from Drs. T. Li and J. Luo. This study is supported by the National Natural Science Foundation of China (Grant No. 40505019)

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