Vol 16 No 5, May 2007
1009-1963/2007/16(05)/1472-05

Chinese Physics

Effects of South China Sea/western North Pacific summer monsoon on tropospheric biennial oscillation (TBO)*

Zheng Bin(郑 彬)†, Gu De-Jun(谷德军), Lin Ai-Lan(林爱兰), and Li Chun-Hui(李春晖)

Guangzhou Institute of Tropical and Marine Meteorology/the Key Open Laboratory for Tropical Monsoon, China Meteorological Administration, Guangzhou 510080*, China*

(Received 6 September 2006; revised manuscript received 19 October 2006)

Several theories have been developed to explain tropical biennial oscillation (TBO), as an air–sea interactive system to impact Asian and global weather and climate, and some models have been established to produce a TBO. A simple 5-box model, with almost all the key processes associated with TBO, can produce a TBO by including air–sea interactions in the monsoon regions. Despite that, the South China Sea/western North Pacific summer monsoon (SCS/WNPSM), a very important monsoon subsystem, is neglected. In this paper, based on the dynamical framework of 5-box model, the term of SCS/WNPSM has been added and a 6-box model has been developed. Comparing the difference of TBO sensibilities with several key parameters, air–sea coupling coefficient α, SST-thermocline feedback coefficient γ and wind-evaporation feedback coefficient λ, between the modified model and original model, TBO is more sensible to the parameters in the new model. The results imply that the eastern Pacific and local wind-evaporation play more important roles in the TBO when including SCS/WNPSM.

Keywords: South China Sea, western North Pacific, summer monsoon, tropospheric biennial oscillation

PACC: 9260S

1. Introduction

Many studies have indicated that a quasi-biennial period is pronounced in atmospheric circulations and interannual climate variability.[1−3] In comparison with stratospheric quasi-biennial oscillation, tropospheric (quasi-) biennial oscillation (TBO) is much weaker, so that it sometimes is covered or distorted by other stronger signals[4] or secular trend.[5] Despite that, TBO is a strong quasi-periodic phenomenon on a large scale in the troposphere, only weaker than the El Niño–Southern Oscillation (ENSO). Together with ENSO, TBO affects atmospheric circulation and leads to weather and climate anomalies. As an example, Feng *et al*[6] has found a marked quasi-biennial period in the precipitation probability of Yangtze River Delta in the last 50 years. Moreover, Lau and Weng[7] have identified three coherent modes of summertime rainfall variability over China and global sea surface temperature (SST) for the period of 1955–98 by singular value decomposition. They found that the second mode comprises 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 anomaly during the transition from El Niño to La Niña in 1997–98. Subsequently, Lau and Wu[8] investigated the covariability of the Asian summer monsoon and ENSO by 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.

At present, there are several TBO theories proposed. First, TBO is a result of air–sea–land interaction.[9,10] But some studies indicated that the land-surface condition is not essential for producing a TBO.[11,12] Secondly, tropical and mid-latitude teleconnection results in TBO.[13,14] Among the theories, some associate TBO with ENSO.[15−17] As mentioned above, there are certainly interactive impacts between TBO and ENSO, but TBO is independent rather than a result of ENSO. Thirdly, local air–sea interactions produce TBO.[18,19] Yu *et al*[20] performed TBO

*Project supported by the National Natural Science Foundation of China (Grant No 40505019) and the Open Research Fund of Laboratory of China Meteorological Administration (Grant No CMATG2006L03).

†E-mail: zhb@ustc.edu

experiments by a coupled general circulation model (GCM), and the results indicated that it would be led an opposite TBO to observations if only local air–sea interactions are taken into account. Fourthly, local and remote air–sea interactions establish TBO.[21,22] Chang and Li[21] pointed out clearly that TBO is a result of interactions among monsoon and tropical Indian Ocean and Pacific Ocean and is different from the ENSO air–sea system. They used a simple dynamical model to reproduce TBO that is close to the observations. Both Chang and Li[21] and Li *et al*[22] considered the South Asian and Australian monsoon by neglecting the South China Sea (SCS)/western North Pacific summer monsoon (WNPSM). Based on their model dynamical framework, we added into the term of SCS/WNPSM. The results from modified model will be compared with those from the original model to investigate the effects of SCS/WNPSM on TBO.

2. Model description

A good model is greatly helpful for us to understand some physical processes.[23,24] The model by Chang and Li[21] and Li *et al*[22] includes the tropical eastern Pacific, tropical western Pacific, tropical Indian Ocean, South Asian monsoon and Australian monsoon simplified into 5 boxes as Fig.1(a). The rates of time change of the Indian Ocean (with subscript I) and western Pacific (with subscript W) SST anomalies (SSTA) are given by

$$\frac{\partial T_\mathrm{I}}{\partial t} = \left[-\lambda\Delta\bar{q}_\mathrm{I}\frac{\bar{U}_\mathrm{I}}{V_0} - \bar{T}_\mathrm{I}^{(x)}\frac{\alpha}{\rho hr} + \bar{T}_\mathrm{I}^{(z)}\frac{(H-h)\beta\alpha}{\rho Hr^2}\right]U_\mathrm{I} - \left(\lambda V_0\kappa + \frac{\bar{w}_\mathrm{I}}{h}\right)T_\mathrm{I}, \tag{1a}$$

$$\frac{\partial T_\mathrm{W}}{\partial t} = \left[-\lambda\Delta\bar{q}_\mathrm{W}\frac{\bar{U}_\mathrm{W}}{V_0} + \bar{T}_\mathrm{W}^{(z)}\frac{(H-h)\beta\alpha}{\rho Hr^2}\right]U_\mathrm{W} - \left(\lambda V_0\kappa + \frac{\bar{w}_\mathrm{W}}{h}\right)T_\mathrm{W} + \left[-\bar{T}_\mathrm{C}^{(x)}\frac{\alpha}{\rho hr} - \bar{T}_\mathrm{W}^{(z)}\frac{2\alpha(H-h)}{L_\mathrm{EW}\rho r} - \frac{\bar{w}_\mathrm{W}\gamma\alpha L_\mathrm{EW}}{2\rho g' Hh}\right]U_\mathrm{C}, \tag{1b}$$

where the terms having the parameter λ in the right sides of Eq.(1) represent a linear form of surface wind–evaporation feedback, the second term of Eq.(1a) and the fifth of Eq.1(b) are the zonal temperature advection, and the third and last terms of Eq.1(a) and the second, fourth and last two terms of Eq.1(b) are the vertical temperature advection. In Eq.(1), U and w denote the surface zonal wind and ocean vertical velocity at the base of the mixed layer respectively. ρ is the density of water and β denotes the planetary vorticity gradient. Subscript C stands for the equatorial central Pacific. The surface zonal winds, U, are given by

$$U_\mathrm{I} = \delta_\mathrm{I}c_1T_\mathrm{I} + c_2T_\mathrm{W}, \quad U_\mathrm{W} = \delta_\mathrm{A}c_3T_\mathrm{W} - \delta_\mathrm{I}c_4T_\mathrm{I},$$
$$U_\mathrm{C} = -\delta_\mathrm{A}c_5T_\mathrm{W} - \delta_\mathrm{I}c_6T_\mathrm{I} + c_7(T_\mathrm{E} - T_\mathrm{W}), \tag{2}$$

where the interactive coefficients $c_1, c_2, \cdots, c_6$ are determined by a scale analysis[21] and c_7 an approximation based on observations.[22] The standard parameters values in Eq.(1) and (2) are specified in Table 1. Li *et al*[22] took an approximation by $T_\mathrm{E} = -2T_\mathrm{W}$ according to the observations. Finally, by substituting Eq.(2) into Eq.(1) with the specified parameter values (listed in Table 1), we derived a pair of homogeneous differential equations that is piecewise in time:

$$\frac{\partial T_\mathrm{I}}{\partial t} = aT_\mathrm{I} + bT_\mathrm{W}, \quad \frac{\partial T_\mathrm{W}}{\partial t} = cT_\mathrm{I} + dT_\mathrm{W}, \tag{3}$$

where $a = -3.15 - 9.9\delta_\mathrm{I}$, $b = -4.06$, $c = 13.34\delta_\mathrm{I}$ and $d = 0.27 + 12.53\delta_\mathrm{A}$. The solution for the equations (3) is not only oscillatory but also has a biennial period in summer, which implies the northern summer monsoon plays an important role in the TBO.

Neither Chang and Li[21] nor Li *et al*[22] considered SCS/WNPSM in their box model. In recent years, SCS/WNPSM has been paid great attention to and considered as an important tropical monsoon subsystem.[25] Thus, it should be included in a TBO model. Based on the dynamical framework of Chang and Li[21] and Li *et al*,[22] the term of SCS/WNPSM is inserted into as shown in Fig.1(b). Then the central-western Pacific zonal wind anomaly is rewritten as $U_\mathrm{W} = \delta_\mathrm{A}c_3T_\mathrm{W} - \delta_\mathrm{I}c_4T_\mathrm{I} + \delta_\mathrm{W}c_8T_\mathrm{W}$. The additional term $\delta_\mathrm{W}c_8T_\mathrm{W}$ represents the excited surface zonal wind anomalies to the south of the SCS/WNPSM region corresponding to the anomalous vorticity induced by anomalous heating over the monsoon region. Additional parameters are listed in Table 1 by bold face. Now we are going to compare the effects of several key parameters on TBO with two box models to understand the role of SCS/WNPSM in TBO.

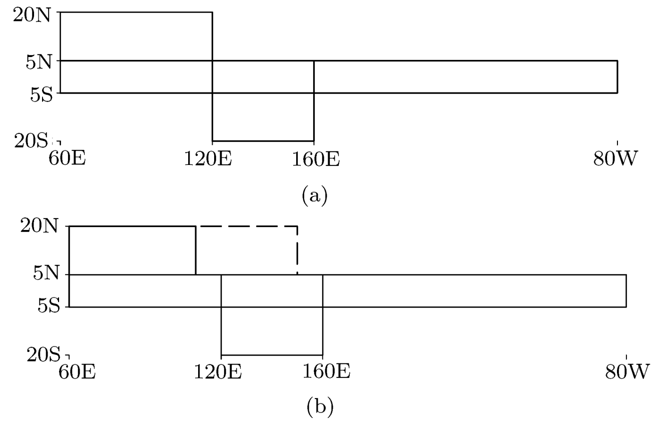

Fig.1. Schematic diagram of the 5-box model from Chang and Li [21] (a) and modified 6-box model (b).

Table 1. The list of key parameters of the model.

Parameter	Symbol	Value	Parameter	Symbol	Value
Ocean thermocline mean depth	H	150 m	Indian Ocean mean vertical temperature gradient	$\bar{T}_{\rm I}^{(z)}$	0.01 K/m
Ocean mixed layer mean depth	h	50 m	Western Pacific mean vertical temperature gradient	$\bar{T}_W^{(z)}$	0.01 K/m
Reduced gravity	g'	0.015 m/s^2	Air–sea coupling coefficient	α	9.6×10^{-3} kg· m$^{-2}\cdot$ s^{-1}
SST-thermocline feedback coefficient	γ	0.18 K/m	Wind-evaporation feedback coefficient	λ	2.1×10^{-5} K/m
SST-specific humidity constant	κ	7×10^{-4} K^{-1}	Interactive coefficient	c_1	4.0 mK$^{-1}\cdot$ s^{-1}
Oceanic Ekman layer friction coefficient	r	1×10^{-5} s^{-1}	Interactive coefficient	c_2	1.6 mK$^{-1}\cdot$ s^{-1}
Half-length of Pacific basin	$L_{\rm EW}$	8×10^6 m	Interactive coefficient	c_3	4.0 mK$^{-1}\cdot$ s^{-1}
Air–sea humidity difference	$\Delta\bar{q}$	5.6×10^{-3}	Interactive coefficient	c_4	4.2 mK$^{-1}\cdot$ s^{-1}
Indian Ocean annual mean zonal wind speed	$\bar{U}_{\rm I}$	3 m/s	Interactive coefficient	c_5	3.0 mK$^{-1}\cdot$ s^{-1}
Western Pacific annual mean zonal wind speed	$\bar{U}_{\rm W}$	0 m/s	Interactive coefficient	c_6	4.2 mK$^{-1}\cdot$ s^{-1}
Mean constant surface wind speed	V_0	4 m/s	Interactive coefficient	c_7	0.3 mK$^{-1}\cdot$ s^{-1}
Indian Ocean mean upwelling speed	$\bar{w}_{\rm I}$	2×10^{-6} m/s	Interactive coefficient	c_8	4.2 mK$^{-1}\cdot$ **s**$^{-1}$
Western Pacific mean upwelling speed	$\bar{w}_{\rm W}$	2×10^{-6} m/s	Seasonal switch coefficient for Australian monsoon	$\delta_{\rm A}$	$\begin{cases}1 & \text{during winter}\\ 0 & \text{rest of the year}\end{cases}$
Indian Ocean mean zonal temperature gradient	$\bar{T}_{\rm I}^{(x)}$	2×10^{-7}K/m	Seasonal switch coefficient for Indian monsoon and western North Pacific monsoon	$\delta_I=\delta_{\rm W}$	$\begin{cases}1 & \text{during summer}\\ 0 & \text{rest of the year}\end{cases}$
Central Pacific mean zonal temperature gradient	$\bar{T}_C^{(x)}$	-5×10^{-7} K/m			

3. Result analysis

The parameters α, γ and λ are selected to study TBO in the modified box model. Since TBO is a result of tropical air–sea interactions, α, representing air–sea coupling coefficient, is very important. γ is a parameter associating SSTA with ocean thermocline depth anomaly, and λ is related to a local air–sea interaction as a wind-evaporation feedback coefficient. Both γ and λ participate in the key processes for TBO.

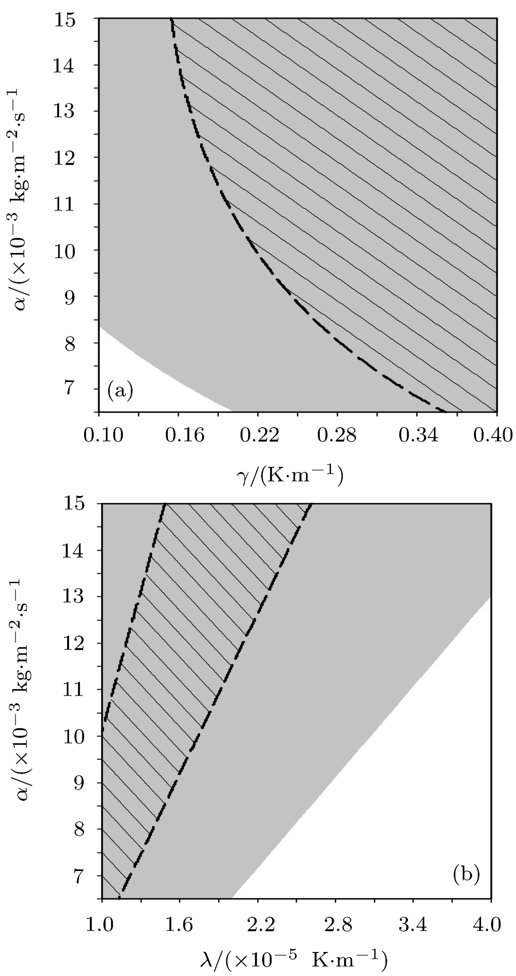

Fig.2. Sensitivities of TBO to several significant parameters for 5-box model from Li *et al*[22] and modified 6-box model (Shaded area is TBO region for the 5-box model and dashed line around shadow shows the TBO region for the 6-box model).

Figure 2(a) shows the difference of the TBO sensibility in the two models to the parameters α and γ. It can be seen that TBO is more sensible to α and γ due to SCS/WNPSM. For example, the values of α and γ in Table 1 can produce a TBO system in the 5-box model, but not in the 6-box model that has included SCS/WNPSM. So, with the same SST-thermocline feedback coefficient γ, stronger air–sea coupling is necessary to establish a TBO in the system containing SCS/WNPSM. On the other hand, with same air–sea coupling coefficient, 6-box model needs more sensible relationship between SST and thermocline depth to generate a TBO. Observations had told us that, among the tropical Indian Ocean, western Pacific and eastern Pacific, SST in the eastern Pacific is most sensible to the thermocline. And it affects TBO in the monsoon–ocean system by remote air–sea interactions. This implies that remote air–sea interactions should play an important role in TBO genesis in the system including SCS/WNPSM.

Figure 2(b) indicates that TBO in the modified model is also more sensible to λ. With a certain air–sea coupling strength in the modified model, λ is restricted in a narrow value area. That means neither extremely strong local air–sea interaction nor extremely weak one can establish a TBO in the system having SCS/WNPSM. The equations (1) and (2) show that the local air–sea interaction associated with λ is a negative feedback process that is a key for oscillating. For an extremely strong negative feedback which would result in an oscillation by the period less than 1.7 a that is not a TBO (TBO period is defined between 1.7–3.0 a in this paper), an extremely weak negative feedback may generate a period larger than 3 a or even a non-periodic variability.

4. Summary

Chang and Li[21] developed a theory for TBO, tropical ocean–Asian and Australian monsoon system (AAMS) interactions, by a simple 5-box model including tropical atmosphere–ocean local and remote interactions. Their model considered two subsystems in AAMS, namely the South Asian monsoon and Australian monsoon, but it skipped SCS/WNPSM. Based on the dynamical framework of Chang and Li,[21] the term of SCS/WNPSM was included by the modified 6-box model. Comparing the TBO sensibilities with several key parameters between the modified and original model, TBO is more sensible in the modified model to the parameters, air–sea coupling coefficient α, SST-thermocline feedback coefficient γ and wind-evaporation feedback coefficient. Additionally, it is noticeable that in the modified model the remote interaction from the eastern Pacific plays a more important role. Besides, the local air–sea interaction with certain intensity greatly contributes to establish a TBO as well, but the interactive strength could not be extremely strong or weak.

TBO theories are still in development,[26] and there is a need for further deep studies to understand it better.

References

[1] Zheng B, Chen Y J and Jian J 2003 *Chin. J. Atmospheric Sci.* **27** 280

[2] Zheng B, Chen Y J and Zhang H 2003 *Chin. J. Atmospheric Sci.* **27** 387

[3] Zheng B, Shi C H and Chen Y J 2006 *Chin. Sci. Bull.* **51** 1761

[4] Zheng H L, Gao X Q and Zhang W 2005 *Chin. Phys.* **14** 1265

[5] Shi N, Yi Y M, Gu J Q and Xia D D 2006 *Chin. Phys.* **15** 2180

[6] Feng G L, Dong W J and Li J P 2004 *Chin. Phys.* **13** 1582

[7] Lau K M and Weng H Y 2001 *J. Climate* **14** 1294

[8] Lau K M and Wu H T 2001 *J. Climate* **14** 2880

[9] Yasunari T 1988 *Meteor. Res. Report University of Tokyo* **1** 30

[10] Meehl G A 1994 *Science* **265** 263

[11] Meehl G A and Arblaster J M 2002 *J. Climate* **15** 923

[12] Meehl G A, Arblaster J M and Loschnigg J 2003 *J. Climate* **16** 2138

[13] Wu R G and Kirtman B P 2004 *J. Climate* **17** 1623

[14] Meehl G A 1997 *J. Climate* **10** 1921

[15] Li C Y, Sun S Q and Mu M Q 2001 *Adv. Atmos. Sci.* **18**(4) 554

[16] Zou L, Wu A M and Ni Y Q 2002 *J. Tropical Meteor.* **18**(1) 19 (in Chinese)

[17] Fasullo J 2004 *J. Climate* **17**(15) 2972

[18] Meehl G A 1987 *Mon. Wea. Rev.* **115** 27

[19] Clarke A J, Liu X and Gorder S V 1998 *J. Climate* **11** 987

[20] Yu J Y, Weng S P and Farrara J D 2003 *J. Climate* **16** 3072

[21] Chang C P and Li T 2000 *J. Atmos. Sci.* **57** 2209

[22] Li T, Tham C W and Chang C P 2001 *J. Climate* **14** 752

[23] Mo J Q, Wang H, Lin W T and Lin Y H 2006 *Acta Phys. Sin.* **55** 6 (in Chinese)

[24] Mo J Q, Wang H and Lin W T 2006 *Acta Phys. Sin.* **55** 3229 (in Chinese)

[25] Wang B, Wu R G and Lau K M 2001 *J. Climate* **14** 4073

[26] Zheng B and Liang J Y 2005 *J. Tropical Meteor.* **21** (1) 79 (in Chinese)