TROPICAL TROPOSPHERIC BIENNIAL OSCILLATION AND ENSO

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Abstract

The tropospheric biennial oscillation (TBO) and El Nino-Southern Oscillation (ENSO) are interannual variations that affect the Asian-Australian monsoon region and the tropical Pacific. Some investigators view the two oscillations are actually one system, with TBO a "weaker sister" of ENSO. Others view the two as separate modes, with TBO primarily driven by the monsoon and ENSO primarily driven by atmosphere-coupling of the tropical eastern Pacific. In this paper the Chang-Li theory, which belongs to the latter category, is used as the basis to study the interactions of the two oscillations. It is shown that the nonlinear interaction of the two modes may produce a variety of solutions that may explain the complexity of the observed monsoon-ENSO relationship.

1. Introduction

Based on previous strong El Nino events, prediction of the effects of the 1997-98 El Nino on the precipitation anomalies in North America and the equatorial maritime continent were quite successful. However, the prediction of large deficits in rainfall over Australia and India did not materialize. This is indicative of important processes of different time scales that interfere with the effects of ENSO in these regions. We propose that the interactions between TBO and ENSO modes may determine the main interannual variations in the Asian-Pacific sector.

It is now well known that the El Nino and La Nina events result from active interactions between the ocean and atmosphere in the tropical Pacific. The anomalous sea-surface temperature (SST) conditions in the Pacific has be predicted with some skill up to several seasons in advance, using either a relatively simple model such as Cane and Zebiak (1985) or complex coupled ocean-atmosphere general circulation models . There are growing observational evidences that anomalous SST conditions in the equatorial Pacific can influence global weather and climate patterns. However, the predictions of these anomalous patterns using either a statistical or a dynamic method are still difficult. In particular, although observational records indicate that rainfall anomalies over Australia and Asia are somewhat correlated to the SST anomalies (SSTA) in the eastern Pacific, from time to time such a correlation is rather vague.

In addition to the external influence from ENSO which generally has a time scale of 3-7 years, the Asian-Australia monsoon system also experiences its own interannual variability, with a dominant time scale of 2-3 years (Meehl 1987, 1993, 1994, Yasunari 1990). The latter is referred to as the tropospheric biennial oscillation (TBO). Meehl (1997) showed that it is the result of coupling between atmosphere, ocean, and land. Chang and Li (2000) proposed that it involves active interactions between the northern summer and winter monsoon and between the anomalous monsoon heating and the planetary-scale east-west circulation. Whereas ENSO involves ocean-atmosphere interactions mainly in the Pacific, TBO covers an even wider spatial scale, including the Asian-Australian monsoon sector and the entire tropical Indian and Pacific oceans.

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Observations have established that the signals of the TBO are found not only in the Indian-Australia rainfall records, but also in the tropospheric circulation, SST, and upper-ocean thermal fields (e. g, Yasunari 1991, Ropelewski et al 1992, Lau and Yang 1996). One of the most remarkable features of TBO is its characteristic seasonal progression and dynamically coherent spatial structure. The conditions of a strong (or weak) monsoon over India and Southeast Asia in the northern summer often continues in the succeeding autumn and winter over the maritime continent and Australia monsoon region. So that a strong (weak) Asian monsoon is often followed by a strong (weak) Australian monsoon. Using a simple five-box model that considers SST-monsoon, evaporation-wind, monsoon-Walker circulation, and wind-thermocline feedback, Chang and Li (2000) illustrated that these processes can give rise to the seasonal evolution of TBO. Their theory explains why TBO can maintain the same phase from northern summer to northern winter and why a reversed phase of TBO can last three locally inactive seasons to affect the next year's monsoon.

Without the interference from and interaction with motions of other scale, Chang and Li (2000) obtained a pure, regular biennial oscillation. But in reality, TBO exhibits a rich, irregular spectrum. Such irregularity may, in large part, result from nonlinear interaction with ENSO. Since TBO is an inherent mode of the coupled ocean-atmosphere system in which monsoon is a major part, the prediction of anomalous rainfall over the Asian-Australian monsoon region based on ENSO forecast must consider the combined effects of TBO and ENSO.

The purpose of this study is to investigate the role of the nonlinear interaction between TBO and ENSO in the monsoon rainfall anomalies and the effect of ENSO on the irregularity of TBO. We will begin by summarizing Chang and Li's (2000) model for TBO, and then incorporate a simple ENSO mode based on Li (1997) in the model to study the interaction of the two modes in a unified dynamic framework.

2. The Chang-Li TBO Model

In the TBO model proposed by Chang and Li (2000) consists of five boxes representing, respectively, the South Asian and Australian monsoon regions and the equatorial Indian, and western and eastern Pacific oceans. The five regions interact among themselves through the SST-monsoon, evaporation-wind, monsoon-Walker circulation, and wind stress-ocean thermocline feedback.

The change of SSTA in the equatorial Indian Ocean (T_I), and the western (T_W) and eastern Pacific (T_E), is primarily determined by horizontal and vertical temperature advection and surface latent heat fluxes. The vertical temperature advection includes effects of both anomalous upwelling and thermocline-depth related subsurface temperature changes. Anomalous heating over the monsoon region excites both planetary-scale east-west circulation and local cyclonic flows near the surface, both of which further change the SST through the dynamic and thermodynamic processes. These SST equations may be found in Chang and Li (2000).

In the following equations we shall define δ_I and δ_A as switch-on coefficients for the Indian and Australian monsoons, respectively. They assume value of unity during the respective monsoon season and zero otherwise. On the interannual time scale, surface zonal winds over the Indian Ocean consist of two major parts, as illustrated in the following equation:

$$U_I = c_1{}'Q_I + c_2T_W = \delta_I c_1 T_I + c_2 T_W \,. \qquad (1)$$

Here the first term on the right hand side gives the direct response to anomalous monsoon heating and the second part reflects the change of intensity of a Walker cell over the Indian ocean due to the SST change in the maritime continent. Eq. (1) is schematically illustrated in Fig. 1a.

The surface wind in the western Pacific, on the other hand, is determined by the anomalous heating over Australia and by the Indian monsoon which, by altering the planetary-scale east-west circulation, changes the surface wind in both the western and central equatorial Pacific (Barnett et al., 1989; Meehl, 1997):

$$U_W = c_3' Q_A - c_4' Q_I = \delta_A c_3 T_W - \delta_I c_4 T_I \qquad (2)$$

This equation is shown schematically in Fig. 1b.

In addition to the influence by anomalous South Asian and Australia monsoon heating, the central equatorial Pacific wind is also affected by zonal SST gradients along the equator (Lindzen and Nigam 1987) so that the total anomalous zonal wind in the central Pacific is written as:

$$U_C = -c_5' Q_A - c_6' Q_I + \frac{A}{\varepsilon L_{EW}}(T_E - T_W)$$

$$= -\delta_A c_5 T_W - \delta_I c_6 T_I + \frac{A}{\varepsilon L_{EW}}(T_E - T_W)\,, \qquad (3)$$

where *A* is a SST-gradient momentum forcing coefficient (Wang and Li 1993), and ε is an atmospheric Rayleigh friction coefficient. The interactive coefficients c1, c2 …. c6 are determined from a scale analysis [see Chang and Li (2000) for details]. Eq. (3) is represented schematically in Fig. 1c.

The surface wind in turn drives the ocean surface current and induces ocean vertical overturning. A simplified Cane-Zebiak (Zebiak and Cane 1987) model is used to calculate the ocean surface currents and Ekman pumping velocity along the equatorial Indian and Pacific Oceans:

$$u_I = \frac{\alpha U_I}{\rho h r}\,, \qquad (4)$$

$$w_I = -\frac{(H-h)\beta\alpha U_I}{\rho H r^2}\,, \qquad (5)$$

$$u_C = \frac{\alpha U_C}{\rho h r}\,, \qquad (6)$$

$$w_W = -\frac{(H-h)\beta\alpha U_W}{\rho H r^2} + \frac{2h(H-h)u_C}{L_{EW}}\,, \qquad (7)$$

$$w_E = -\frac{(H-h)\beta\alpha U_C \sigma}{\rho H r^2} - \frac{2h(H-h)u_C}{L_{EW}} \qquad (8)$$

Here β denotes the planetary vorticity gradient; *H* the mean depth of the ocean thermocline; and *r* a friction coefficient in the oceanic Ekman layer. The vertical velocities in (7) and (8) consist of both meridional and zonal convergence components. The meridional component of the Ekman convergence in the eastern Pacific is estimated with a ratio of σ from its counterpart in the central Pacific. The interactive coefficients c_1,

$c_2 \ldots c_6$ are determined from a scale analysis of the relevant dynamic and thermodynamic balances.

After integrating the model in time, a biennial oscillation emerges in a reasonable parameter regime, with model SST and wind variations resembling many aspects of the observed TBO (Fig. 2). A strong (weak) South Asian monsoon is always followed by a strong (weak) Australian monsoon.

The physical mechanism that causes the TBO in the model can be summarized as follows. Since the starting point of an ideal TBO cycle is arbitrary, we will start from a warm SSTA condition in the Indian Ocean prior to the South Asian summer monsoon season. To aid our discussion, a schematic diagram illustrating how the interactions between the monsoon, ocean, and atmosphere promote a biennial oscillation in the model is given in Fig. 3. In this figure the main cause-effect for the strong monsoon phase are given in the left hand side, with the sequence of development driven by the northern summer events indicated by red solid thin and double arrows, while those driven by the northern winter events by black dashed double arrows. The chain of events in the weak monsoon phase is only symbolically sketched on the right hand side since the details are exactly mirror images of the left hand side. The two phases are separated by a ribbon. Fig. 4 graphically summarizes all the important interactive processes shown in Fig. 1a-c, and is presented as a companion diagram to Fig. 3. A warm SSTA condition in the equatorial Indian ocean increases the local lower tropospheric moisture through surface evaporation. As the South Asian summer monsoon develops, southerly winds to the south of the Indian subcontinent bring the excess water vapor into South Asia, which intensifies the convective rainfall and leads to a strong monsoon. The convective heating associated with the stronger monsoon on one hand induces a westerly wind anomaly over the Indian Ocean due to the enhanced low-level vorticity (indicated by a thin arrow), and on the other hand intensifies a planetary-scale east-west circulation leading to anomalous easterlies over the western and central Pacific (indicated by the solid double arrow). The westerly anomaly over the Indian Ocean decreases the local SST, primarily through the evaporation-wind feedback. This cooling will help the development of a weak phase of the South Asian summer monsoon if the cold SST can persist for nearly one year.

The anomalous easterlies associated with the intensified east-west circulation affect the western Pacific SST in two ways. Over the western Pacific the seasonal mean is westerly, therefore the mean wind is reduced resulting in a reduced evaporation, leading to a warming effect. Meanwhile, the central Pacific easterly anomalies deepen the ocean thermocline in the western Pacific, so that the SST also increases due to anomalous vertical temperature advection. (The easterly anomaly also tends to cool the local SST through zonal advection and Ekman-induced upwelling, but both effects are small – the latter due to the weak mean vertical stratification in the western Pacific.) As a result, a warm SST persists in the western Pacific through the northern fall, which eventually leads to a stronger Australia monsoon in the northern winter and the beginning of the winter-driven event sequence indicated by the dashed double arrows.

Whenever there is warming in the western Pacific, it strengthens the western branch of the Walker cell and thus a surface westerly anomaly over the Indian Ocean. Therefore, the westerly anomalies over the Indian Ocean is reinforced during the northern fall (the solid double arrow). This helps the cold SST to persist throughout the succeeding seasons. Meanwhile, the easterly anomaly in the central Pacific, through anomalous upwelling and shoaling of the thermocline, also cools the SST in the eastern Pacific. The east-west SST gradient further intensifies the anomalous easterly over the central Pacific, reinforcing the tilting of the thermocline and the warming of the western Pacific SST.

The warm SSTA in the western Pacific are weakened in late northern fall by anomalous cold advection from the eastern Pacific. However, they are reinvigorated in the northern winter when the Australian monsoon becomes stronger. This is accomplished through two processes: the western Pacific westerly anomaly that weakens the seasonal mean easterlies resulting in a reduction of evaporation, and the central Pacific easterly anomaly (associated with the strengthened east-west circulation) through the change in the tilting of the thermocline. Thus, the winter driven events (indicated by the dashed double arrows) become a third mechanism to keep the Indian Ocean's surface wind anomaly westerly and SST cold, leading to a weaker Asian monsoon in the following summer. (The warm western Pacific SST further intensifies the eastern Walker cell and helps to keep the eastern Pacific cold.) This interactive system explains why the Indian Ocean SSTA, can be kept from rapid weakening for several seasons.

The reversed cycle associated with a weak Asian monsoon is sketched to the upper-right of the ribbon in Fig 3. The thin arrows that form a closed loop surrounding this ribbon represent the local wind-evaporation negative discussed by Meehl (1987, 1997) for the TBO of the South Asian monsoon. In order to explain the coherent temporal and spatial structure, the interactions between the monsoons, east-west circulation, and ocean thermocline variations, as represented by the double arrows (both solid and dashed) in the strong monsoon phase in Fig. 3, are required. These processes are responsible for an Australian monsoon whose TBO phase follows the South Asian monsoon, and for a sustainable SST anomaly in the Indian Ocean to cause a phase reversal of the TBO cycle one year later.

The proposed TBO theory is based on processes within the tropics, thus it excludes the possible feedback through tropical-midlatitude interactions. However, such interactions will not change the phase of the simulated oscillation, because their effects would give rise to the same sign of response in the biennial cycle as that of the equatorial SST produced by the present model.

3. Incorporating the stationary SST ENSO Mode

When ENSO is included in the model, the wind anomalies over the Indian Ocean and the western and central Pacific consist of two parts. One is related to the internal TBO mode that is associated with the east-west circulation due to anomalous monsoon heating and SST gradients as described in the preceding section. The other part describes the effect of ENSO on TBO. Thus, we have,

$$U_I = \delta_I c_1 (T_I - \delta_{ET} T_E) + c_2 T_W , \qquad (9)$$

$$U_W = \delta_A c_3 (T_W - \delta_{ET} T_E) - \delta_I c_4 (T_I - \delta_{ET} T_E), \qquad (10)$$

$$U_C = -\delta_A c_5 (T_W - \delta_{ET} T_E) - \delta_I c_6 (T_I - \delta_{ET} T_E) + c_7 [\delta_{ET} T_E - (1+\psi) T_W] , \qquad (11)$$

where δ_{ET} is a switch-on coefficient representing the strength of ENSO influence on TBO. The interactive coefficient c7 is defined in (3).

In the TBO model, we intentionally filtered out the ENSO mode by considering only the zonally asymmetric thermocline depth variation that is in Sverdrup balance with the wind or SST anomaly. However, on the ENSO time scale the zonal mean thermocline depth variation is clearly not in equilibrium with the wind or SST anomaly. Li

(1997) pointed out that while the asymmetric variation influences the coupled instability through a positive feedback between the ocean and atmosphere, the zonally symmetric variation is associated with a negative feedback and is essential for the phase transition of ENSO. Following Li's stationary SST mode ENSO model, with <h> denoting the zonal mean thermocline depth anomaly, the eastern Pacific SSTA equation may be written,

$$\frac{dT_E}{dt} = \Psi U_C^{(ENSO)} - (\lambda V_0 \kappa + \frac{\overline{w}_E}{h}) T_E + \frac{\overline{w}_E}{h} \gamma < h > , \quad (12)$$

$$\frac{d < h >}{dt} = -\Lambda T_E - \varepsilon_h < h > , \quad (13)$$

where $\Psi = \frac{\overline{T}_E^{(z)} (H-h) \beta \alpha \sigma}{\rho H r^2} + \frac{2h(H-h) \alpha \overline{T}_E^{(z)}}{\rho h r L_{EW}} + \frac{\overline{w}_E \gamma L_{EW} \alpha}{2 h \rho g' H}$, $\Lambda = 1.3 \times 10^{-7} ms^{-1} K^{-1}$ and $\varepsilon_h = (2 year)^{-1}$ are constants.

The TBO mode influences the ENSO mode through the central Pacific zonal wind that is a function of both eastern and western equatorial Pacific SSTs:

$$U_C^{(ENSO)} = c_7 (T_E - \delta_{TE} T_W) , \quad (14)$$

where δ_{TE} is a switch-on coefficient describing the influence of TBO on ENSO.

Equations (9-14), together with the SSTA equation for T_I and T_W, form a simple dynamic framework for ENSO-TBO interactions. The two modes interact in such a way that a strong monsoon due to TBO would produce anomalous easterly in the central Pacific that further leads to an anomalous cooling tendency for the eastern Pacific SST, as represented by the δ_{TE} term. The anomalous warming in the eastern Pacific associated with ENSO, on the other hand, has both direct and indirect effects on the monsoon rainfall anomalies. Due to the associated Walker cells, the SST anomaly in the Indian Ocean in general tends to be in-phase with the eastern Pacific but out-of-phase with the western Pacific. This indirect effect causes the eastern Pacific El Nino condition to lead to a warm SST anomaly over the Indian Ocean and thus a stronger Indian monsoon. Meanwhile, a large-scale descent over the Asian-Australian monsoon region due to the anomalous vertical overturning associated with El Nino directly produces a deficient monsoon rainfall anomaly. The combination of these two effects would determine the net effect of ENSO on TBO-monsoon rainfall anomalies.

4. Nonlinear TBO-ENSO Interaction

When $\delta_{ET} = \delta_{TE} = 0$, this system has two independent solutions, with one being the TBO mode and the other the ENSO mode. The former involves the dynamic feedback between the Asian-Australia monsoon heating and the planetary-scale east-west circulation, and the latter involves interactions between the zonal mean (and zonally asymmetric) thermocline depth variation and SST anomaly. Fig. 5 shows the model solutions for the two modes. The TBO mode (top panel) is characterized by regular biennial oscillations in India and Australia rainfall anomalies. A strong Australian monsoon follows a strong Indian monsoon, as observed. The ENSO mode (bottom panel) is characterized by SST variations in the eastern equatorial Pacific, which has an oscillation period of 3-4 years and an amplitude of 4°C.

The top panel of Fig. 6 illustrates the net effect of the model ENSO on TBO ($\delta_{ET} = 1, \delta_{TE} = 0$). Note that the regular biennial oscillations of anomalous rainfall over both the Asian and Australia monsoon regions have been perturbed. A weak rainfall anomaly is not always followed by a strong one. The amplitude of the rainfall anomaly varies with time. A strong warm event in the eastern Pacific associated with El Nino does not always coincide with a drought condition in Australia (e.g., compare rainfall anomaly in Australia at year 9 in Fig. 6 with that in Fig. 5). When $\delta_{ET} = 0, \delta_{TE} = 1$, ENSO is also interfered by TBO as shown in the bottom panel of Fig. 6. The eastern Pacific SST oscillation becomes more irregular, owing to the change of the central Pacific wind associated with TBO.

The results presented above illustrate one-way interactions, either TBO affecting ENSO or ENSO affecting TBO. The nonlinear two-way interactions between the two modes exhibit even richer characteristics. Figure 7 is a scatter diagram in the phase space of eastern Pacific SST anomaly versus the Australian rainfall anomaly in northern winter, obtained from a 1000-year integration of the model with $\delta_{ET} = \delta_{TE} = 0.1$. In most cases there is no sequential relationship between adjacent solutions in this diagram. There are at least three types of limit cycles in the phase space. The first is a narrow cross-center type, diagonally oriented from top-left to bottom-right, that reflects the conventional ENSO theory: A positive (negative) SST anomaly in the eastern Pacific leads to a deficient (excess) rainfall in northern Australia. This narrow cycle is accompanied by a round cycle around the center (0,0). This limit cycle is related to the regular TBO mode, since it does not depend on SST anomaly in the eastern Pacific. A maximum (minimum) Australian anomaly occurs near $T_E = 0$. The third type is like a bow with its two ends bounded by the previous two limit cycles. It may result from the nonlinear interaction between the TBO and ENSO modes. Overall, this phase solution indicates that the anomalous rainfall in Australia depends on the combination and competition of both the ENSO and TBO modes, and that a strong El Nino may cause either a deficient rainfall (greater than a standard deviation) or small positive or negative rainfall, but it is unlikely to reach an opposite extreme. The 1997-98 El Nino event may be described by the circle area in the phase space.

The results presented here demonstrate that it may be possible to predict certain important aspects of anomalous rainfall over the Asian-Australian monsoon regions, but only when both ENSO and TBO modes are included in a dynamic coupled ocean-atmosphere model. Many previous models, either statistical or dynamic ones, failed because they only considered the ENSO effect.

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$$U_I = c_1' Q_I + c_2 T_W = \delta_I c_1 T_I + c_2 T_W$$

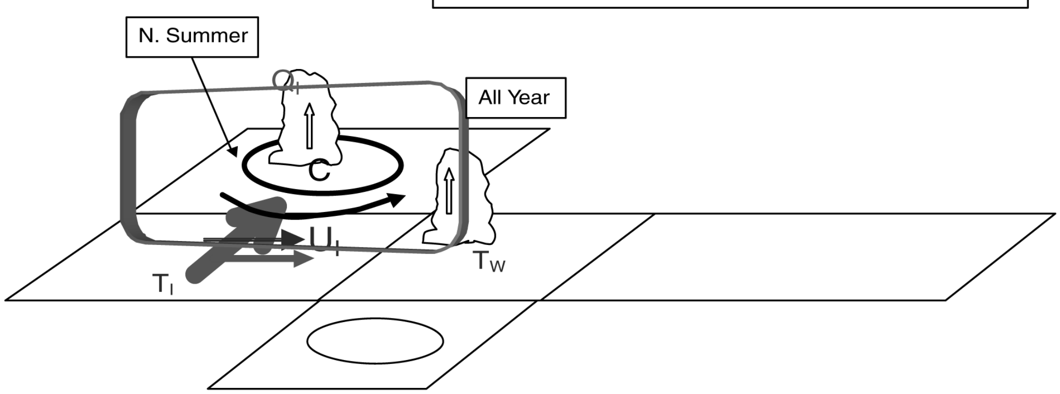

Fig. 1a. Schematic diagram showing the relationship between U_I and T_I and T_W in the TBO model. The Indian Ocean SST determines the anomalous moisture convergence (green thick arrow) into South Asia, which in turn determines the convective heating (Q_I) and cyclonic vorticity (***C***) of the northern summer monsoon, therefore relating T_I and U_I during northern summer only. The western equatorial Pacific SST determines the strength of the western Walker cell. This cell operates in all four seasons and causes T_I to influence U_I continuously throughout the year. (from Chang and Li 2000).

$$U_W = c_3'Q_A - c_4'Q_I = \delta_A c_3 T_W - \delta_I c_4 T_I$$

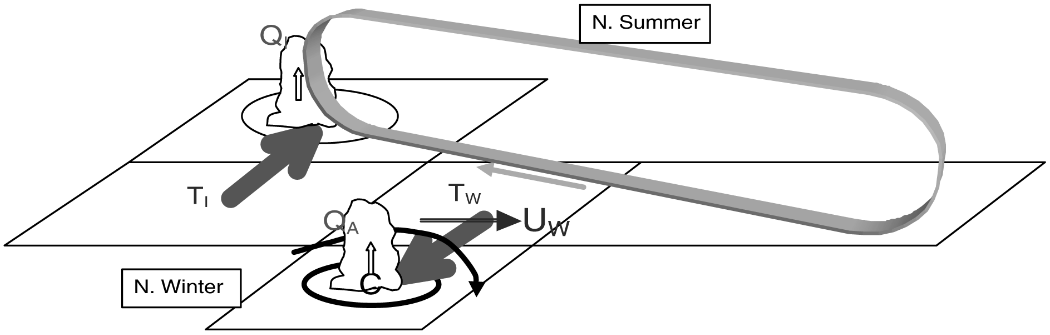

Fig. 1b. Schematic diagram showing the relationship between U_W and T_I and T_W in the TBO model**.** The western equatorial Pacific SST determines anomalous moisture convergence (green thick arrow) into Australia, which in turn determines the convective heating (Q_A) and cyclonic vorticity (***C***) of the southern summer monsoon, therefore relating T_W and U_W during northern winter only. The strength of the east-west circulation depends on the convective heating of the South Asian monsoon, thus it links the Indian Ocean SST with U_W during northern summer. (From Chang and Li 2000).

$$U_C = -c_5'Q_A - c_6'Q_I + \frac{A}{\varepsilon L_{EW}}(T_E - T_W)$$
$$= -\delta_A c_5 T_W - \delta_I c_6 T_I + \frac{A}{\varepsilon L_{EW}}(T_E - T_W)$$

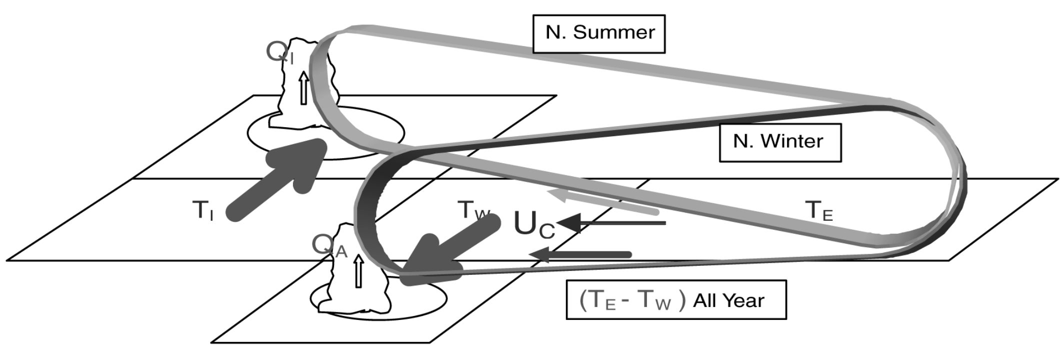

Fig. 1c. Schematic diagram showing the relationship between U_C and T_I , T_W and T_E in the TBO model**.** Through the strength of the east-west circulations, U_C is affected by the Indian Ocean SST during northern summer and the western equatorial Pacific SST during northern winter. In addition, the SST gradient between the eastern and western equatorial Pacific influences U_C throughout the year. (From Chang and Li 2000).

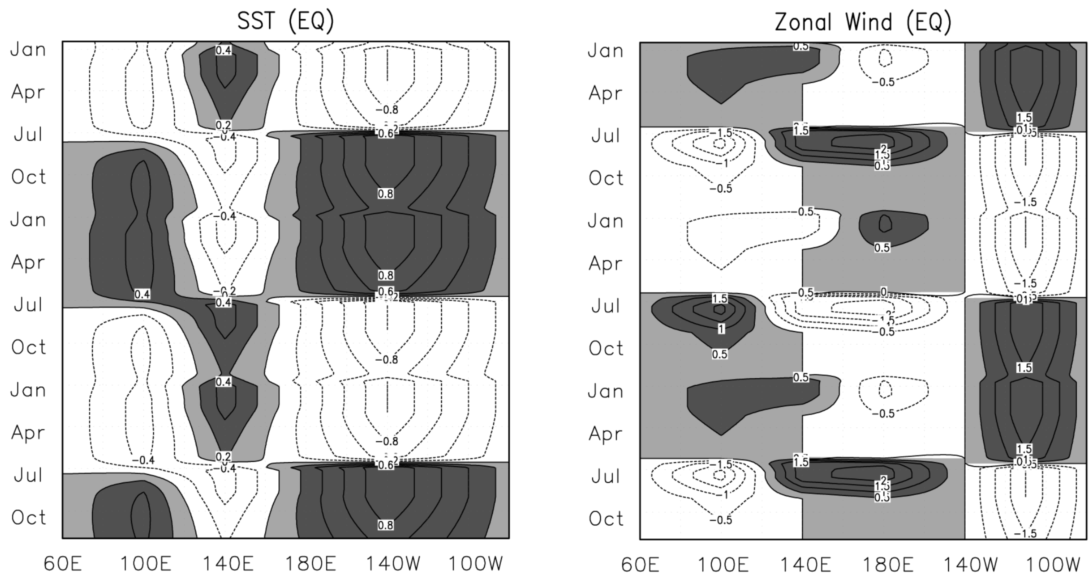

Fig. 2. Time evolution of the model SST (K, Intervals 0.2) and zonal wind (m s^{-1}, Intervals 0.5) anomalies, positive areas shaded. The eastern Pacific zonal wind is not a model variable. The wind anomalies east of 140°W are computed from zonal SST gradient anomalies due to T_E. (From Chang and Li 2000).

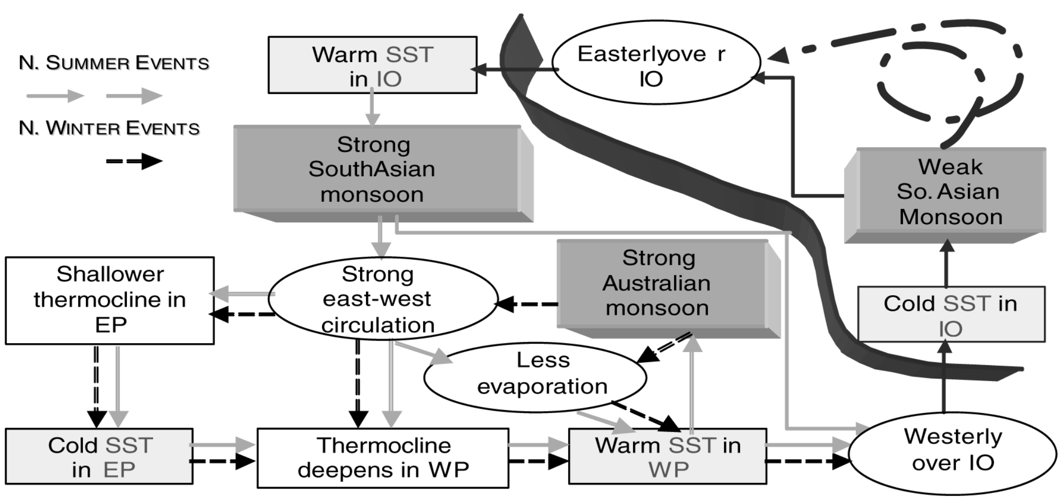

Fig. 3. Schematic diagram indicating the interactive processes leading to the TBO. Land and oceans regions are shaded. Non-shaded boxes indicate atmospheric (oval) and oceanic (rectangular) processes. The strong monsoon phase starts with warm Indian Ocean SST leading to a strong South Asian summer monsoon. In addition to a surface westerly anomaly during summer (thin arrow) that cools the Indian Ocean SST, complex interactive processes involving the tropical western Pacific and the Australian monsoon are required to reinvigorate this westerly anomaly until next summer. The processes also lead to a strong winter monsoon that follows the strong summer monsoon. The reversed phase is sketched in the upper right side. (From Chang and Li 2000).

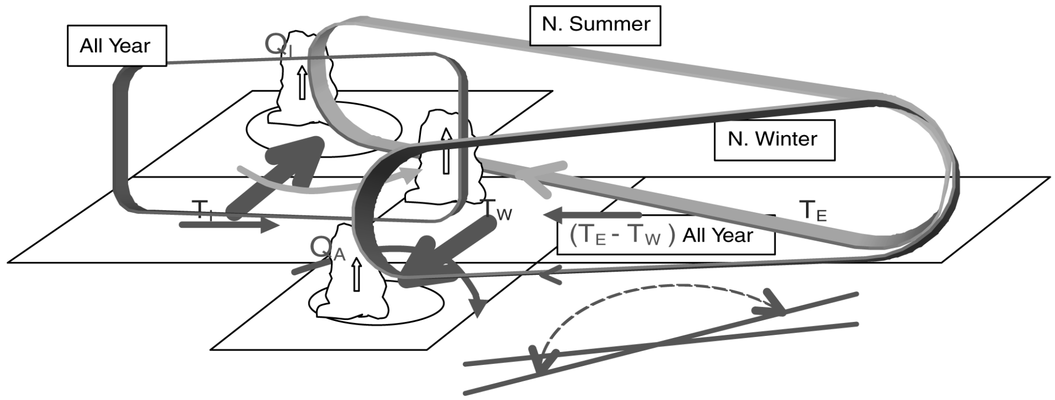

Fig. 4. Schematic diagram summarizing the important interactions. The primary SST change mechanism is through wind-evaporation for the Indian Ocean, thermocline tilting for the eastern equatorial Pacific, and both for the western equatorial Pacific. Notice the key roles played by the western equatorial Pacific both in allowing the northern summer monsoon to influence the northern winter monsoon, and in the maintenance of the Indian Ocean surface wind anomalies until the next monsoon by linking the feedback of the Australian monsoon to the Indian Ocean. (From Chang and Li 2000).

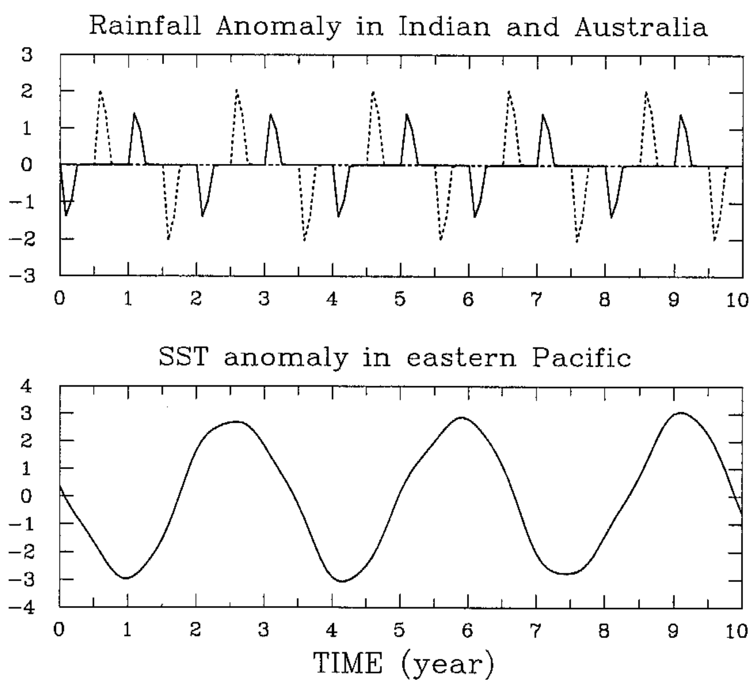

Fig. 5 .Time series of anomalous rainfall rate (mm/day) over India (dotted line) and Australia (solid line) and anomalous SST (°C) in the eastern Pacific (bottom panel) in the case of no TBO-ENSO interactions ($\delta_{ET} = \delta_{TE} = 0$).

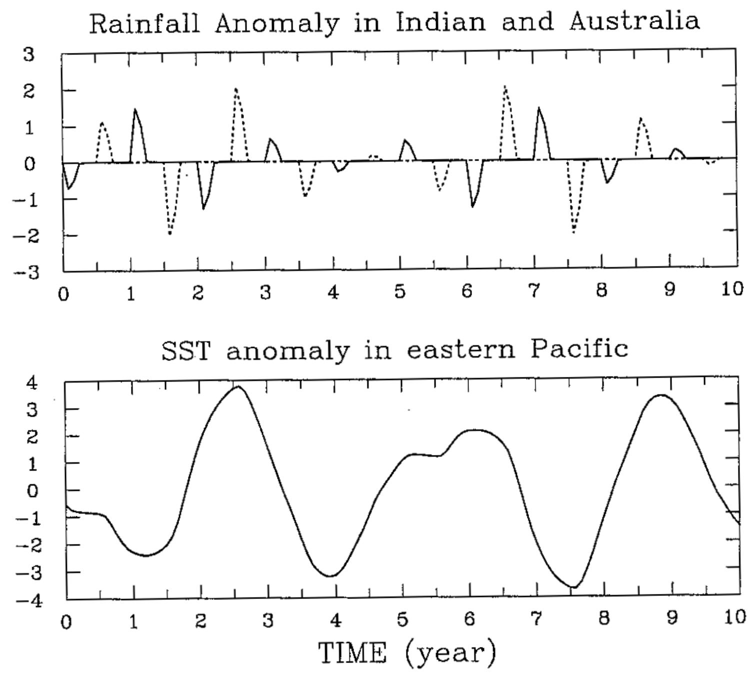

Fig. 6 Time series of anomalous rainfall rate (mm/day) over India (dotted line) when only ENSO influences TBO, and Australia (solid line) and anomalous SST (°C) in the eastern Pacific (bottom panel) when only TBO influences ENSO.

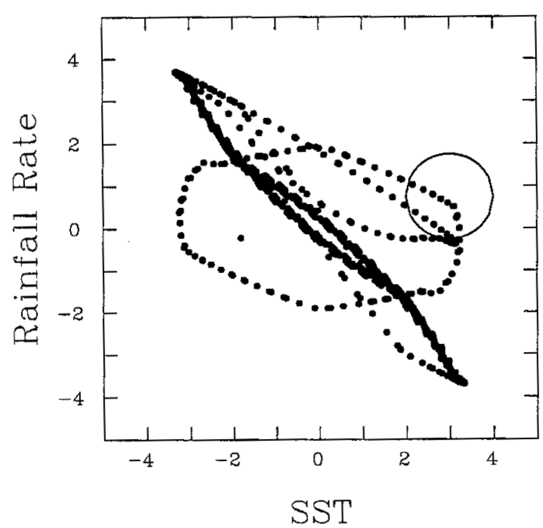

Fig. 7. Scatter diagram of December Australian rainfall anomaly (mm/day) versus eastern Pacific SST anomaly (°C), obtained from a 1000-year model integration under nonlinear two-way interactions ($\delta_{ET} = 0.1$, $\delta_{TE} = 0.1$). A circle denotes a possible phase regime for the 97-98 El Nino.