Scholarly Paper

Coupled Instability in the Equatorial Pacific Ocean

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Abstract

The El Niño-Southern Oscillation (ENSO) phenomena, due to its tremendous influence on global climate and society, have been studied comprehensively for several decades, especially during the TOGA decade. There are mainly two views of ENSO: one characterizes it as a stable system that needs an external trigger (McPhaden, et al., 1998; Kessler, 2002) and the other considers it as a self-sustained oscillation (Schopf & Suarez, 1987; Suarez & Schopf, 1988; Jin, 1997a,b). We hypothesize that whether it's a cycle or an event depends on the mean state of equatorial Pacific Ocean, which is influenced by mean depth of thermocline along the equator and the coupling coefficient in simple atmospheric models. A mean state which results in the coupling between ocean and atmosphere can support a self-sustained oscillation. An Ocean GCM coupled with simple atmospheric model is used to study the coupled instability problem in the equatorial Pacific ocean. The control run with a coupling coefficient 0.4 and relatively shallow thermocline results in a coupled system without external easterlies which leads to a zonally asymmetric mean state and an ENSO-like behavior similar to observations. This is consistent with the recently proposed theory proposed by Cai (2003) which proposed that coupling itself can explain the zonally asymmetric mean state and the ENSO phenomena. Sensitivity experiments show that the coupling in the ocean-atmosphere system depends crucially on two factors: thermocline depth along the equator and the coupling coefficient. The results from coupling coefficient experiments are compared to that in Cai (2003) and the differences with his idealized model are analyzed.

1. Introduction

Climatological state of the equatorial Pacific ocean consists of a western warm pool and an eastern cold tongue near the sea surface, and the thermocline tilting up toward east, which is driven by easterlies over the equatorial ocean. Associated with this asymmetric climatological state are the ENSO phenomena, which is the largest natural climate variability at internannual scales.

Due to its global influence on climate and the society, ENSO has been studied for a long time. Our understanding about ENSO has been advanced in the past several decades as a result of numerous studies, including observations and numerical simulations (for a comprehensive review, see JGR 1998). There are mainly two characterizations of ENSO: one is that ENSO is a stable system which needs an external trigger (McPhaden, et al., 1998; Kessler, 2002) and the other considers it as a cycle (Schopf & Suarez, 1987; Suarez & Schopf, 1988; Jin, 1997a,b) which is self-sustained oscillation. Although these two perspectives are different, they are both based on the Bjerknes feedback (Bjerknes, 1969) processes as far as the growth of the initial anomalies are concerned. In this process, positive sea surface temperature (SST) in the eastern tropical Pacific bring about the relaxation of the easterly trade winds, which makes the SST anomalies stronger leading to a growing mode. In recent 20 years, more and more specific feedback mechanisms are proposed as part of Bjerknes feedback, such as the upwelling feedback, advection feedback, and the thermocline feedback. These mechanisms refer to the oceanic response to atmospheric perturbations. In many coupled models whose atmospheric component is simplified, the coupling coefficient parameter **μ** is used to embody the strength of coupling. The general form is $\tau = \mu \frac{\partial T}{\partial x}$, where T is sea surface temperature and τ is the coupled zonal wind stress. This equation, basically a simplified representation of the lower branch of Walker circulation (Walker, 1924), determines the wind stress feedback to SST anomaly. The major differences between these two perspectives are in the origin of the anomaly and the decay of growing mode. The "cycle" theory attributes these to the negative feedback mechanism which resides in the dynamic system of tropical ocean and "event" theory attributes these to external perturbations such as the MJOs or the noise(e.g., Lau and Chan 1988; Weickmann 1991; Kessler et al. 1996) in the system. In the past twenty years, the "cycle" theory has been developed very well and there are two leading views: delayed oscillator theory and recharge-discharge paradigm. Although there are some differences between them, the oscillation in these two views both come from the delay or time lag between positive feedback and negative feedback. In delayed oscillator theory, the lag between wind excited waves and boundary reflected waves is responsible for the phase-shifting and in the latter the oscillation comes from the lag between thermocline tilting response to wind anomaly, which is the positive feedback, and the response of mean depth of thermocline along the equator to the integrated Sverdrup transport caused by wind stress curl, which is negative feedback. If ENSO is a cycle, then one would expect that to be advantageous in predicting this phenomena.

What we propose in this paper is that whether ENSO is an event or a cycle depends on the coupled atmosphere-ocean system in the equatorial Pacific ocean. If the system is coupled and can stay coupled, then ENSO could be a cycle, otherwise it is an event which needs an external trigger. For the system which stays coupled, the coupling itself can influence the property of ENSO, like the amplitude and period.

The first question here is: Is the equatorial ocean-atmosphere system always coupled? The paleoclimate records show that it is not. As far back as 3 million years ago (Philander, 2003), the SST of equatorial Pacific ocean was almost zonally uniform although the easterly trade winds existed over the sea surface. The mean state at that time was zonally symmetric and there was no interannual ENSO or one could say there was a permanent El Niño. This may suggest that the easterlies are not a necessary condition for the system to get coupled. The reason that the system wasn't coupled is explained as too deep a thermocline due to high-latitude warming resulting in a reduced heat uptake in the deep tropics, thus preventing the upwelling influence on SST. This means that the thermocline depth is a factor for determining if the system gets coupled or not. Another factor that will influence the coupling is the so-called coupling coefficient in simple atmospheric models as many previous studies have shown. (Cai, 2003; Dijkstra & Neelin, 1995) There may be many other factors that affect the coupling, but we only study these two factors, which can be considered as two switches which in a series. Only when two switches are both switched on, the system can be coupled. The switch in the atmospheric component is more explicit and the switch in oceanic component is rather implicit.

Our experiments in this paper explore how these two switches — the coupling coefficient and the mean thermocline depth along the equator affect the coupling and hence influence the mean state and ENSO properties.

2 Model Description

The ocean GCM used is a reduced gravity, primitive equation, sigma-coordinate model

of Gent and Cane (1989) with modifications as described in Murtugudde et al. (1996). A variable depth of mixed layer in this model simulates the three major mechanisms of turbulent vertical mixing in the upper ocean: the entrainment/detrainment due to wind and buoyancy forcing, the gradient Richardson number mixing generated by the shear flow instability, and the convective mixing related to static instabilities in the water column (Chen et al. 1994a,b).

The vertical structure of the model ocean consists of discrete layers. The thickness of uppermost layer, representing the surface mixed-layer and the last layer are predicted by the model. The thicknesses of the layers below the mixed-layer are chosen according to a sigma coordinate, hence called sigma layers. The total depth of water column is the sum of mixed-layer and all the sigma layers. The depth of sigma layers is allowed to change but the ration of each sigma layer to the total water column below the mixed layer is fixed to its prescribed value. Once the depth of last layer is predicted, the depth of all sigma layers can be diagnosed using the specified sigma values.

The model domain covers the tropical and subtropical Pacific Ocean extending from

30^oS to 30^oN. The model has 15 sigma layers plus the surface mixed layer. The longitudinal resolution is constant, 1°. The meridional resolution is stretched from 0.3° at the equator to 1.5° at the meridional boundaries.

The OGCM is coupled to an advective atmospheric mixed layer model but here the thermal boundary condition is provided by restoring surface temperatures to an imposed temperature distribution (Haney, 1971).

$$Q = \lambda(T_{air} - T)$$

This formulation assumes that the heat capacity of atmosphere is infinite. The coefficient λ represents the restoration time. At the northern and southern boundaries along 30^oN and 30^oS, the temperature, salinity, and the layer thicknesses are restored to World Ocean Atlas (Boyer et al. 1998) annual mean data in order to prevent underestimating the polarward return flow caused by the meridional overturning circulation. (the WOA data is adjusted to make them symmetric about the equator to avoid introducing asymmetries). The ocean-atmosphere coupling in this model is realized by zonal SST gradients generating wind stresses, which is basically Walker circulation

$$\tau_x = \alpha \frac{\partial T}{\partial x}$$

Whether the wind stresses generated by SST gradients are able to drive SST variability depends on many implicit mechanisms. The fact that coupling coefficient α is greater than zero doesn't guarantee that the ocean-atmosphere system will get coupled.

3. Results

3.1 The control run

The first question we are going to answer is, can the tropical atmosphere-ocean system be coupled given the appropriate mean thermocline depth along the equator and intermediate coupling coefficient without external easterlies? If it can be, does it lead to the zonal asymmetry and ENSO properties similar to observations?

To answer this question, we need to design a standard experiment, with intermediate coupling coefficient and starting with appropriate mean thermocline depth along the equator. In order to get appropriate mean thermocline depth, we need to shoal the thermocline depth along the equator without tilting it. A study about the effect of off-equatorial winds on thermocline depth along the equator (Wang et al. 2003) showed that off-equatorial winds can change thermocline depth in equatorial ocean by Sverdrup transport without significantly changing the slope. Based on this study, we imposed off-equatorial winds poleward of 5S-5N as in fig.1 . After running the model for 30 years initialized with a flat density structure and zero currents and no coupling to the atmosphere, we got the steady state with a shallow, level thermocline depth along the equator, around 60m, and very weak currents in equatorial region, which can be considered nearly as a rest state. Then starting with this initial state and kicked by easterlies in the eastern equatorial ocean for 3 months, with coupling coefficient 0.4, the model was run for 100 years.

a) Simulation of the long time mean state

Fig.2 presents last 80-year mean simulated SST (shaded) and zonal wind (contour).

The fundemental features, namely, the western Pacific warm pool and eastern equatorial cold tongue, easterlies over equatorial ocean, are captured adequately, but the core of cold tongue is too far east compared to the observations. This is an artifact of the simple atmospheric model employed here. Besides being simply proportional to the local zonal SST gradient, the zonal wind is easterly over the whole equatorial basin in the model whereas in the real world, the monsoonal winds penetrate into the western Pacific warm pool. In the real world and in more sophisticated numerical models, when a negative SST anomaly is placed in central or eastern-central equatorial pacific, easterlies will occur to the west of the anomaly and westerlies will occur to the east of the anomaly (Gill 1980). In our model, the kick-up winds placed around 120^{o}E, induce cooling of SSTs due to upwelling, which in turn cause easterlies to the east of the perturbation instead of westerly as expected from the simple relation between zonal wind stress and the local zonal SST gradient. This basin-wide easterly structure leads to a large slope of the thermocline as shown in Fig.3 Another restriction of simple atmospheric model here is the prescribed meridional structure. The net effect is a larger amplitude along the equator and the the curl of wind stress is also larger. Without external easterlies, the coupled model easterlies will always lead to equatorward sverdrup transport and hence deepen thermocline along the equator. These two restrictions give rise to a relatively deep thermocline with a large slope. Another feature seen in Fig.3 is that the vertical scale of the Equatorial Undercurrent (EUC) is large and the thermocline is much more diffuse than that in the real world. The diffuse thermocline is a result of the deep thermocline. As is well known the stability of the ocean can be influenced by two factors: the depth of thermocline and the temperature gradient across the thermocline. The depth of thermocline here is forced by the prescribed meridional coupled wind structure. If the ocean be made unstable, then the vertical temperature gradient has to decrease to compensate the stability caused by deep thermocline. In all of our results based on this simple atmospheric model, oscillating solution is always associated with a deep, diffuse thermocline and the steady solution is a shallow, sharp thermocline. In the real world, there are easterlies due to the Hadley circulation (Hadley, 1735). The external easterlies combined with coupled Walked circulation will lead to a different meridional wind structure, which generates polarward Sverdrup transprot and shoals the thermocline along the equator. In this case, the vertical temperature gradient does not need to be small to compensate the effect of the depth of the thermocline. The solution we got here is thus expected to be modified by the external easterlies. The study about the effect of external easterlies on coupled solution (Dijkstra & Neelin, 1995) showed that the asymmetric seed by the external easterlies will affect both the position of core of cold tongue and amplitude of cold tongue.

b) Simulation of ENSO

The SST anomalies are calculated by removing long time mean from the simulated values. The SSTAs, Wind stress anomalies, MLD anomalies along the equator are shown in Fig.4 for the last 10 years. The SSTAs in the eastern Pacific show standing oscillation and weak westward propagation in central and west equatorial Pacific. The zonal wind stress anomalies show no propagation. The MLD anomalies show weak eastward propagation. Compared to the observations, the maximum anomaly of SST is confined to far the eastern boundary. So we use the averaged SSTA over the region 120W-76W and 5S-5N as the index for the SSTA variability instead of Nino3.4. The zonal wind anomalies are also confined to the east around 120^{o}W where the kick-up wind was placed. In the observations, wind stress anomalies occur in the western Pacific due to warm SSTs which favor convection and there are other high frequency varibilities such as the Madden-Julian Oscillation and the westerly wind bursts (Moore & Kleeman, 1999; Zavala-Garay &Moore, 2005; Belamari et al. 2003). The anomalies of MLD in the western Pacific lead SSTA in eastern Pacific by about half of the cycle, which suggest that the western Pacific is the source of SSTA in the east. This can be confirmed in Fig4. Figure 4a is the peak of La Niña as seen by the cold eastern Pacific, but there is a warm anomaly at a depth of 300m in the western Pacific. This anomaly is brought eastward by the EUC, almost zonally untill it reaches 120^{o}W where the upwelling is the strongest due to strong surface zonal winds. The anomaly thus reaches the sea surface at the eastern Pacific due to EUC and upwelling. During the eastward journey of the warm anomaly, the amplitude is reduced by about a half because of mixing, but this does not have much impact because the role of this anomaly is to provide a small perturbation, which can grow under the local positive SST-wind feedback.

The spectral analysis of SSTA in 120^{o}W-76^{o}W, 5^{o}S-5^{o}N shows that the period is around 3-years, which is relatively short. This is caused by the deep thermocline due to the reasons explained in previous section. This deep thermocline leading to short period is opposite to Fedrov(2003) who showed that deeper thermcoline favors remote processes and hence longer periods. The reasons for our contrary results are partially attributable to the simplistic zonal wind structure and a lack of meridional winds.

3.2 Sensitivity studies

After demonstrating the coupled system itself can lead to an asymmetric mean state and ENSO, we are going to explore how the two factors, the thermocline depth along the equator and the coupling coefficient, affect the coupling and hence change the mean state and ENSO properties.

3.2.1 Thermocline depth along the equator

To test the effect of the equatorial thermocline depth, we simulated several states with different thermocline depth. The starting process is the same as that in standard run, i.e, we change the magnitude of the off-equatorial winds, and therefore change the curl and the mean depth along the equator. Varying from the maximum amplitude from 0.2 dyn/cm^2 to 0.8 dyn/cm^2, we created four different states. But only the states with 0.2 dyn/cm^2 and 0.4 dyn/cm^2 are in steady state, the ones with 0.6 and 0.8 lead to an oscillation due to the meridional inflow in the sponge layers and the equatorial thermcoline has an obvious tilt. Here we just compare the experiment with 0.2 dyn/cm^2 and 0.4 dyn/cm2 off-equatorial winds.

The results from the thermocline sensitivity experiments show two different solutions: the one with relatively shallow thermocline (control run) shows zonally asymmetric mean state associated with self-sustained oscillation around 3 year period;

the other one with relatively deep thermocline shows zonally level mean state without any oscillation. With the similar background conditions, the ocean with different thermocline depths responding so differently to the same kick-up wind stress suggest that the processes involving thermocline depths are very crucial for the initial growth of the anomaly. As we mentioned above, that indicates the important role of the thermocline feedback. In fact, this process is emphasized both in the delayed oscillator theory and the recharge-discharge paradigm. But in these two theories, the roles are stressed more in the decay of the anomaly and the phase shifting. In our results, the processes responsible for the growth of initial anomaly is far more important than the thermocline feedback, which is different from anomaly models that are used to study ENSO. As is known from observations and extensive model studies, the wind anomalies occur at western equatorial pacific because of the convective anomalies over the warm waters and other intraseasonal variabilities mentioned before. In most anomaly models, a wind anomaly is imposed over western Pacific, and it excites eastward propagating Kelvin waves, which shoal or deepen thermocline, change subsurface temperature and imprint an expression on SSTs with the help of the mean upwelling, and cause wind anomalies at the surface. In our model, which is a full OGCM, the changed subsurface temperature does not have a strong expression on SSTs because there is no strong background mean upwelling. So the wind anomaly has to be imposed in the east Pacific to generate local upwelling, combine with initial vertical temperature gradient, to influence SSTs at the largest extent. This is the main reason why the anomaly has to be placed over the east Pacific and the deep thermcoline makes it difficult for the model SSTs to be perturbed . When the depth is large, the vertical temperature gradient at the base of mixed layer, where the subsurface water is entrained into mixed layer and influence SST, tend to be small. Compared to surface heat fluxes which try to equilibrate the SSTs to the air temperature above it and advection that tend to make SSTs homogeneous, the vertical advection is not strong enough to compete with those two effects and the ocean will always stay homogeneous and stable as in the initial state.

3.2.2 Coupling coefficient

a. Mean state

The mean state with 0.2 dyn/cm^2 coupling coefficient returns to the initial homogeneous state after a decaying oscillatory adjustment in the first 16 years. When the coupling coefficient is increased to 0.4, the mean state becomes zonally asymmetric and this basic feature does not change when the coupling coefficient is increased but some specific properties change. Figur 6 shows that when coupling coefficient is changed from 0.4 to 1.0, the thermocline depth in the western pacific gets deeper and deeper and the tilt of thermocline gets larger and larger. These two monotonical changes make the zonal asymmetry at the surface non-monotical as we see from Fig. 7. Also, the EUC gets stronger and stronger when the coupling coefficient increases. This is consistent with two existing explanations for EUC: potential vorticity dynamical reason (Fonfonoff & Montgomery, 1955; Pedlosky, 1996) and pressure gradient reason (Philander, 1990). In P.V. dynamical explanation, the strength of EUC is related to the equatorward geostrophic flow which feed EUC; in pressure gradient explanation, EUC is related to thermocline tilt along the equator which produce PGF and drive EUC. In our results, when α increase, tilt of thermocline increases as the result of integrated coupled zonal wind stress; equatorward geostrophic transport increase in the wake of increased wind stress curl due to prescribed meridional wind stress pattern, which drives increased Sverdrup transport. Note that even as the thermocline depth changes the temperature gradient across the thermocline does not change significantly. This makes the ocean get more stable when the coupling coefficient increases, which is consistent with the conclusion of Cai(2003) that the restoring force increases when the coupling coefficient increases after it passes the second critical value. This change in mean state will lead to some change of the oscillation as we will see in the next section.

b. ENSO

The time series of SST anomalies in eastern Pacific show both the period and amplitude change with the changing of coupling coefficient. When coupling coefficient is 0.2, there is no oscillating solution. When it increase to 0.4, the oscillation occurs with a 3 year period and $6^{o}C$ amplitude. Both the period and amplitude show monotonic relation with coupling coefficient for values larger than 0.4. The decreasing period is explained as the changes in wave propagation speeds associated with the thermocline changes. When coupling coefficient increases, the thermcoline depth along the equator gets deeper and hence the Kelvin wave and first mode of Rossby wave travel faster. In the perspective of delayed oscillator theory, this will lead to a shorter oscillation period. Also, the thermocline depth increasing significantly but temperature gradient across the thermocline not changing much lead to a more stable ocean, which has stronger restoring force and hence a smaller oscillation amplitude. In this trend, there may exist another steady state with zonal asymmetry when the coupling coefficient is large enough.

4 Comparion with other studies

Investigations of the mean state and ENSO have been conducted for a long time and there are some studies that are similar to our own results presented here. We obtained some different results compared to previous studies. We will focus on Cai(2003) and discuss the differences of his results with our results. First, we present a quick review of Cai(2003). In Cai (2003), a linear, reduced-gravity, shallow-water model, combined with a stripped-down equatorial SST equation governed by the Ekman layer and thermocline feedbacks, are coupled to a simple atmospheric model. Starting with an at rest zonally homogeneous state, kick started by basin wide easterly for 3 months, the model was run for 100 years. With the coupling coefficient varying between 0 (uncoupled) and 1 (fully coupled), there are two critical coupling coefficeints found, 0.575 and 0.61. When $\alpha < 0.575$, the model shows a damped oscillation around the zonally symmetric mean state and finally decays back to the initial state. When $0.575 < \alpha < 0.61$, the model show damped oscillation around zonally asymmetric mean state; when $\alpha > 0.61$, the model show self-sustained oscillation around zonally asymmetric mean state. When the coupling coefficient varies from 0.575 to 1, the zonal asymmetry at the surface (T_w-T_e) and subsurface (H_w-H_e) both increase with increasing coupling coefficient. The amplitude increases and period decreases with the coupling coefficient $\alpha > 0.575$. When doing comparison, we mainly focus on the range from 0.61 to 1, where the models show self-sustained oscillation around zonally asymmetric mean state. The common feature in Cai (2003) and our result are zonal asymmetry in the subsurface (H_w-H_e) and oscillation period. The different result is about the zonally asymmetry in the surface (T_w-T_e) and oscillation amplitude. To understand the differences in the results, we have to consider the differences in the models. Although these two models are both hybrid coupled models, the mean depth of thermocline is fixed in Cai (2003). In our model, in addition to the full nonlinear physics including the salinity effects for the evolution of the thermocline, the mean depth of thermocline is controlled by imposing temporally constant off equatorial wind stresses and mostly of the mean depth comes from the evolution of ocean responding to the coupling in equatorial region. In our model, the mean depth of thermocline increases with coupling coefficient. Combined with the tilting of the thermocline, this make the zonal asymmetry (T_w-T_e) on the sea surface uncertain trend. In Cai (2003), with fixed thermocline depth, the increased slope will lead to a larger T_w-T_e. The difference of amplitude relation with coupling coefficient also has some bearing on this fixed thermocline depth. When the thermocline depth increases and the vertical temperature gradient across it does not change much, this leads to stable mean state which favors smaller amplitude of oscillation. Furthermore, the fixed thermocline and variable thermocline make Cai (2003) closer to the delayed oscillator theory and our model simulations closer to the recharge-discharge paradigm. The explanation is as following. The thermocline along the equator responds to the coupled winds in two ways: the tilt response to integrated zonal wind stress along the equator and the mean depth response to Sverdrup transport due to the wind stress curl. The time scale of these two responses are different. The tilt is in quasi-balance with the wind stresses along the equator but the adjustment of mean depth of the thermocline along the equator is slower than the tilt response. The coupled winds with a Gaussian distributed easterlies across the equator lead to opposite effects in the eastern equatorial Pacific for these two responses. The different time scales between these two opposite effects provide the memory for the phase shift in the oscillation. In Cai (2003), there is only one response allowed – the tilt response. Therefore, the recharge-discharge mechanism is excluded from his model. Thus the oscillation from Cai (2003) is close to delayed-oscillator and the oscillation from our model is close to recharge-discharge oscillator.

5. Conclusion

An Ocean GCM coupled to a simple atmospheric model is used to study the coupled instability problem in equatorial Pacific Ocean. Two factors, mean depth along the equator and the coupling coefficient, are chosen the candidates that influence the coupled behavior in the equatorial ocean-atmosphere system. This is because the vertical velocity has a nonlinear relationship with depth. For certain wind stresses over the sea surface, there is a certain depth where the vetical velocity is maximum. If the mean depth of equatorial thermocline is placed around that certain depth, then it can be guranteed that strong upwelling can bring subsurface cold water to the sea surface and hence influence the SSTs and enable coupling. This situation can be changed in two ways: either the mean depth of the thermocline or the upwelling magnitude /the depth of maximum vertical velocity. The standard run with coupling coefficient 0.4 and relatively shallow thermocline showed that the system can be coupled in this condition without external easterlies and the coupled system leads to zonally asymmetric mean state and an ENSO as that in observations. This suggests that the current mean state and ENSO phenomena are favored by β -effect and two meridional boundaries. The results from the sensitivity experiments showed the oceanic response to surface wind stresses are very sensitive to the mean thermocline depth along the equator as we expected. The coupling coefficient tests show that when the coupling coefficient increases after it reaches a critical value where the model can attain a self-sustained oscillation, the zonally asymmetry in the subsurface (H_w-H_e) increase but zonally asymmetry in the surface (T_w-T_e) doesn't show monotonical relation with the coupling coefficient. The oscillation period and amplitude both decrease with coupling coefficient increases as a result of the increase of thermocline depth along the equator due to increased equatorward Sverdrup transport.

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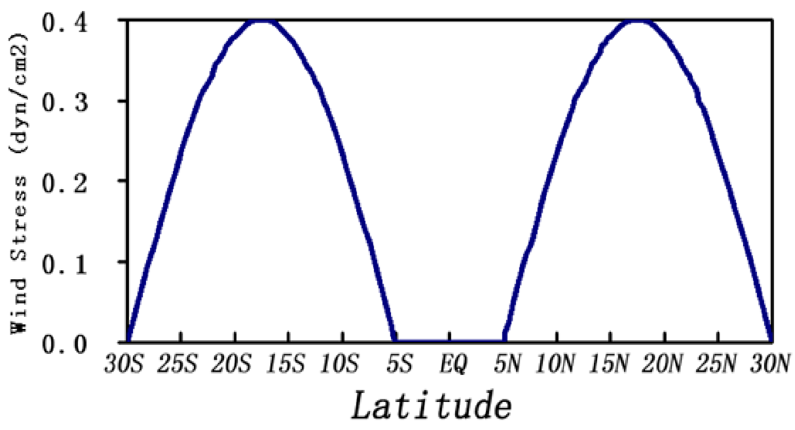


Fig.1 off-equatorial wind stress

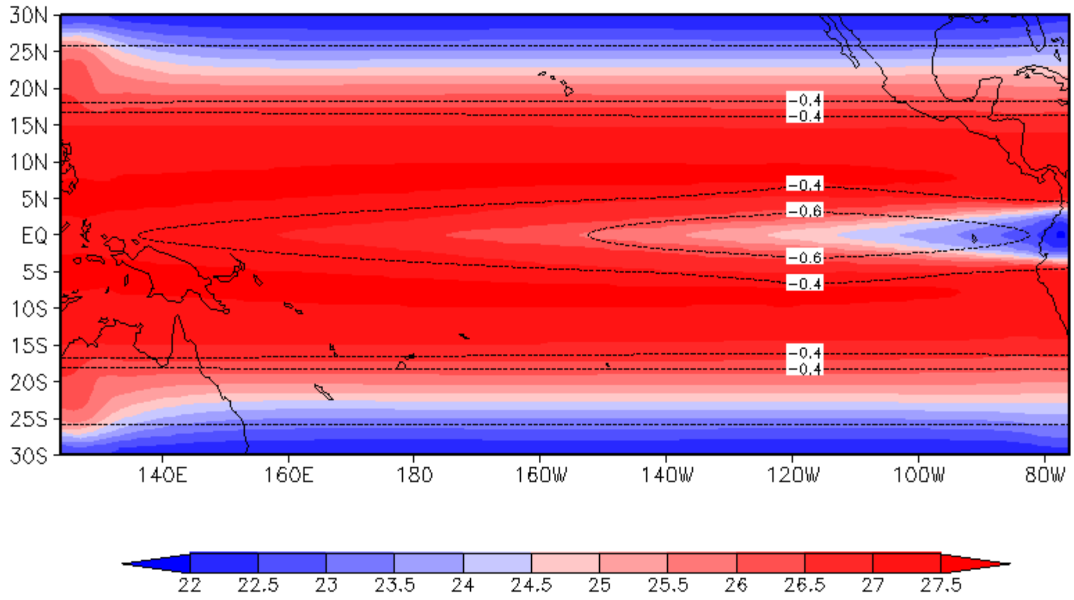


fig.2 long time mean of SST (shaded) and zonal wind stress (contour) from standard run

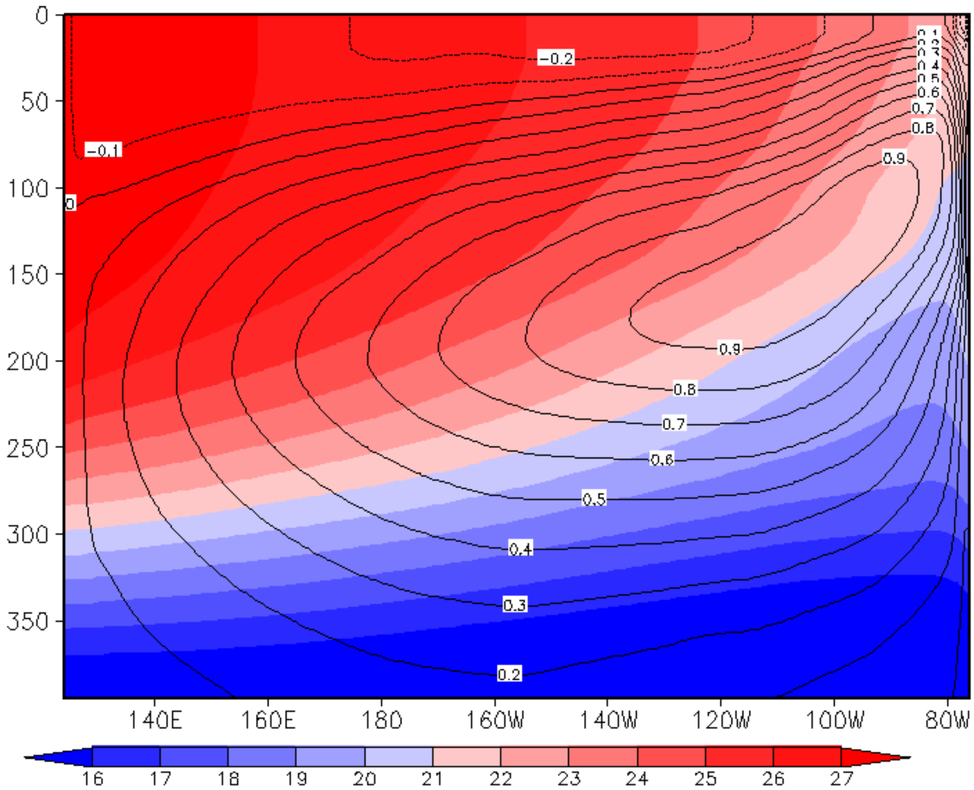


fig.3 long time mean temperature profile (shaded) and zonal velocity profile (contour) along the equator from standard run

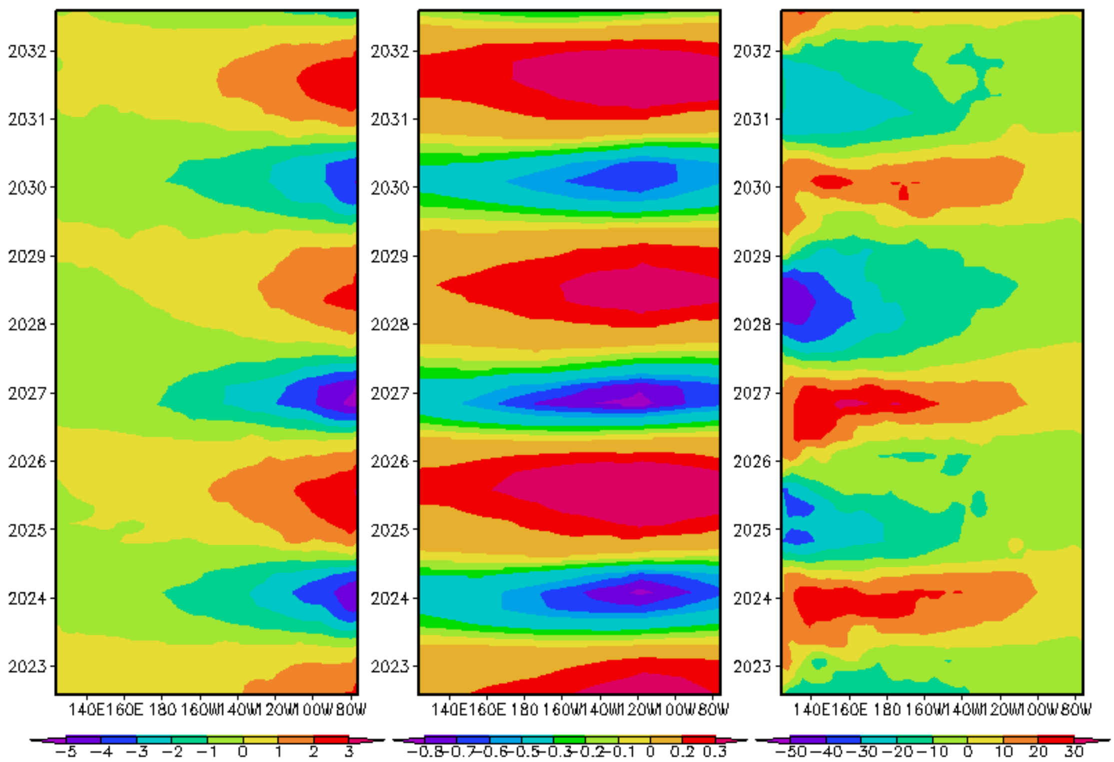


fig.4 SSTA, zonal wind stress anomaly and MLD anomaly along the equator from standard run

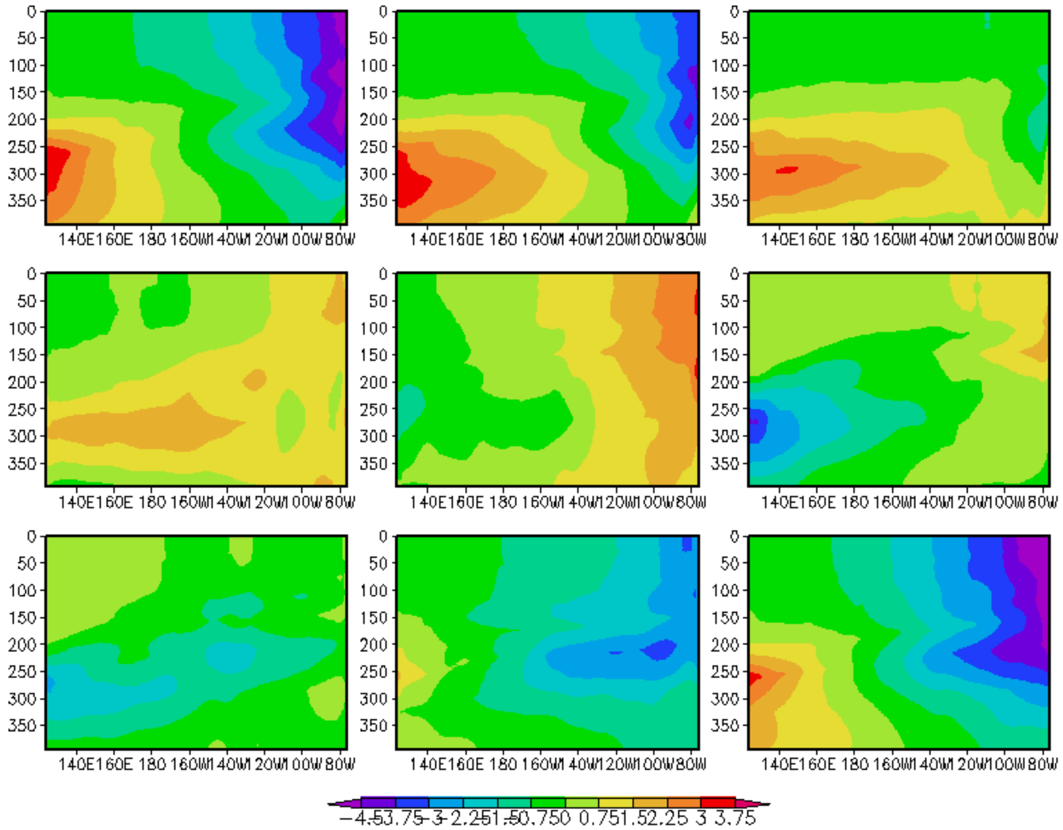


fig.5 SSTA profile along the equator in a ENSO cycle from standard run

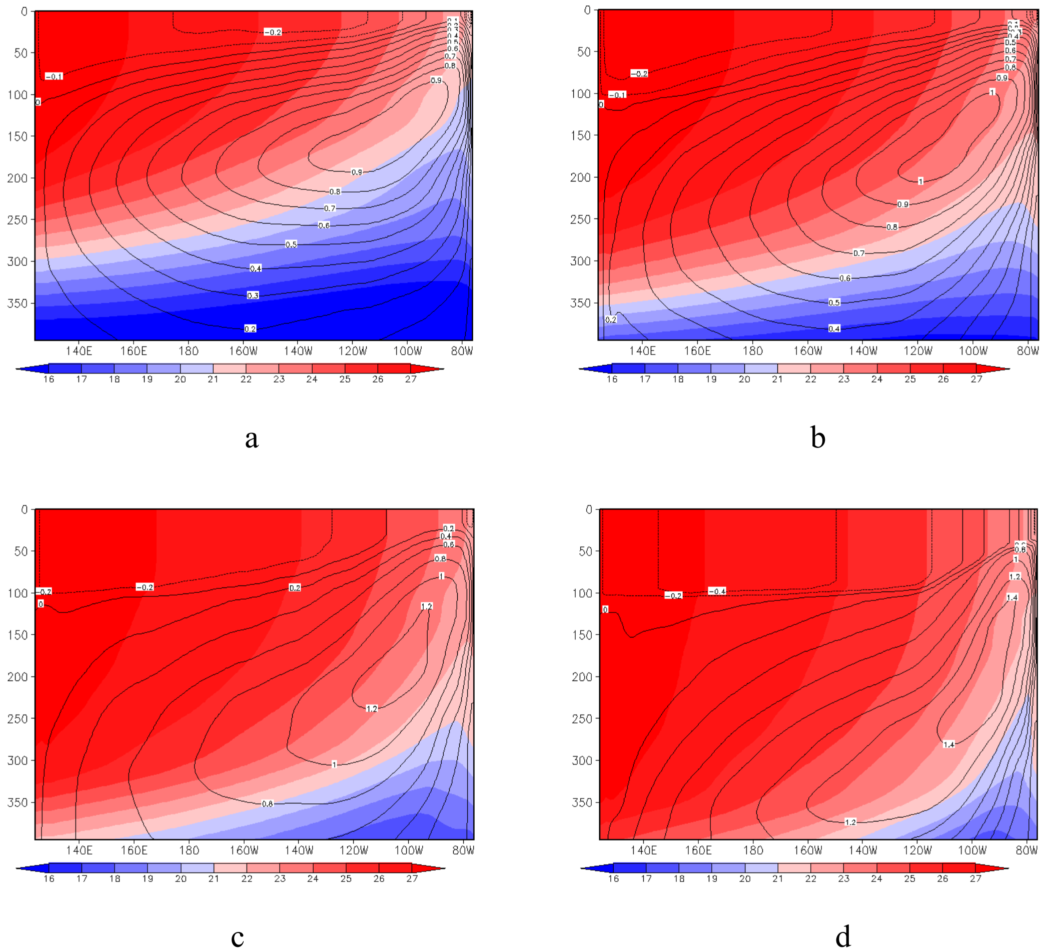
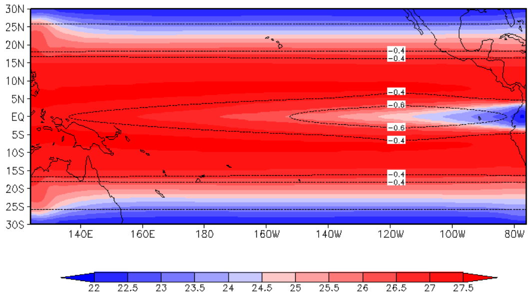
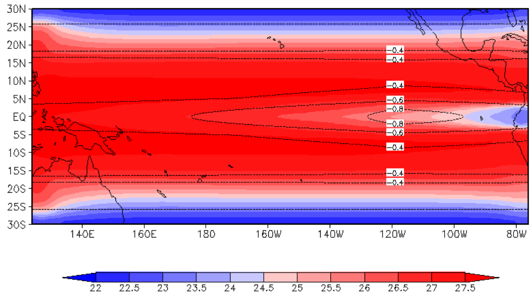


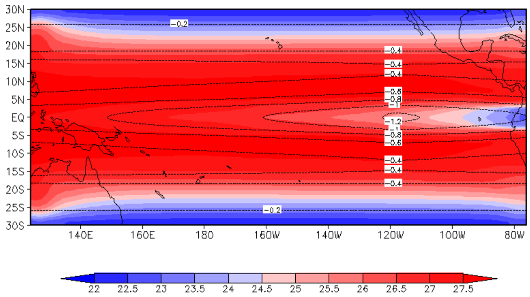
fig.6 long time mean temperature profile (shaded) and zonal velocity profile (contour) along the equator; a) 0.2; b) 0.4; c) 0.6; d) 0.8



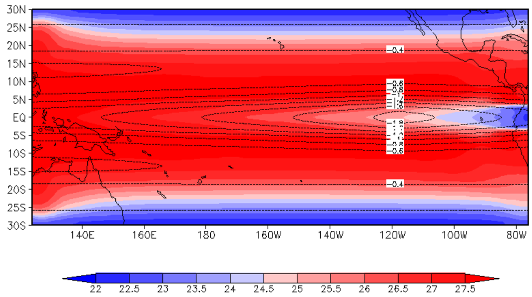
a



b



c



d

fig. 7 long time mean of SST (shaded) and zonal wind stress (contour) a) 0.2; b) 0.4; c) 0.6; d) 0.8

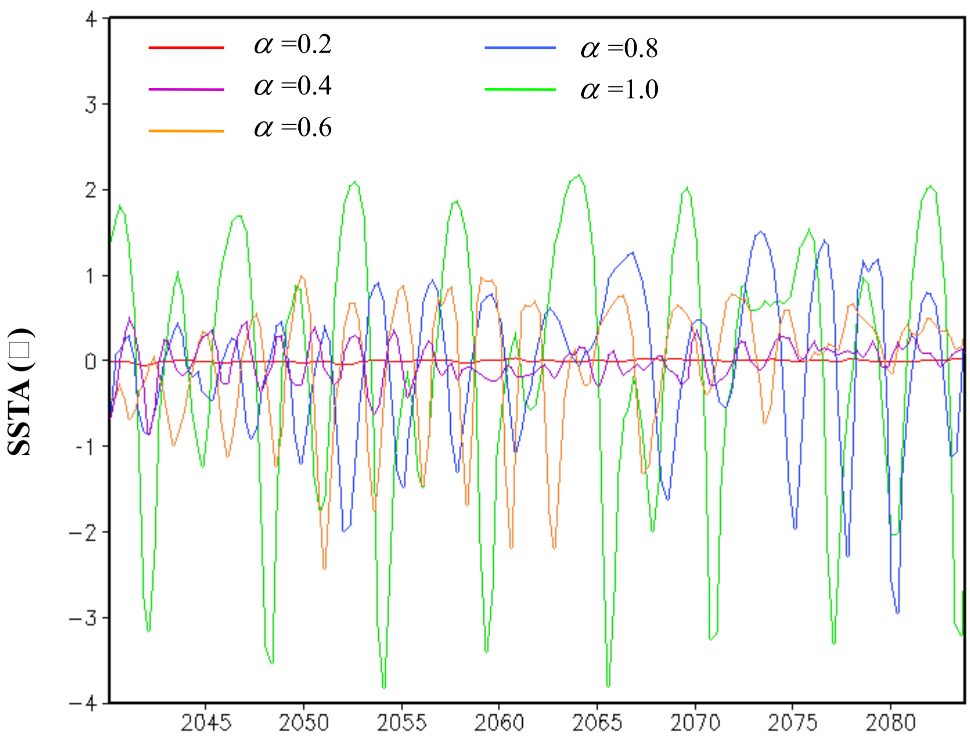


Fig. 8 SSTA averaged over 120W-70W, 5S-5N