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Short Note

On the Recent Development of Simple, Coupled Ocean-Atmosphere Models of ENSO*

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Abstract: Based on simple model results, I describe the recent progress in our understanding of physical mechanisms directly associated with El Niño-Southern Oscillation (ENSO) phenomena. In particular, I extract two extremes from recent simple coupled models in order to interprete two complementary phases of interannual ENSO events. I also discuss what is most necessary for more complete model studies useful to improve forecasting skills.

1. Introduction

As pointed out by Bjerkness (1966) just two decades ago, there is now no doubt that air-sea interaction plays a key role during the El Niño/ Southern Oscillation phenomena. Recent progress in our understanding of the phenomena has been much accelerated by simple modelling studies, and of course, by the best-observed 1982/83 event. Here briefly reviewing the state of the art of simple air-sea coupled models, I direct my attention to examination of underlying physical processes in order to find valid approximation for more complete model studies.

The content of the present concise note will be the following: First, I will review briefly some primary elements of a simple, coupled model which possibly simulates or at least explains major features of ENSO events. Then I will extract two extremes from simple, coupled models and examine their stability properties a little in detail. These two extreme models will serve for our understanding of the evolution of two complementary phases of ENSO, that is, warm phase and cold phase. Based on the argument on the stability properties, a simple interpretation of several numerical experiments reported so far and the actual ENSO phenomena will be given. Finally, summarizing my report, I will briefly discuss what is most necessary for further understanding of ENSO events.

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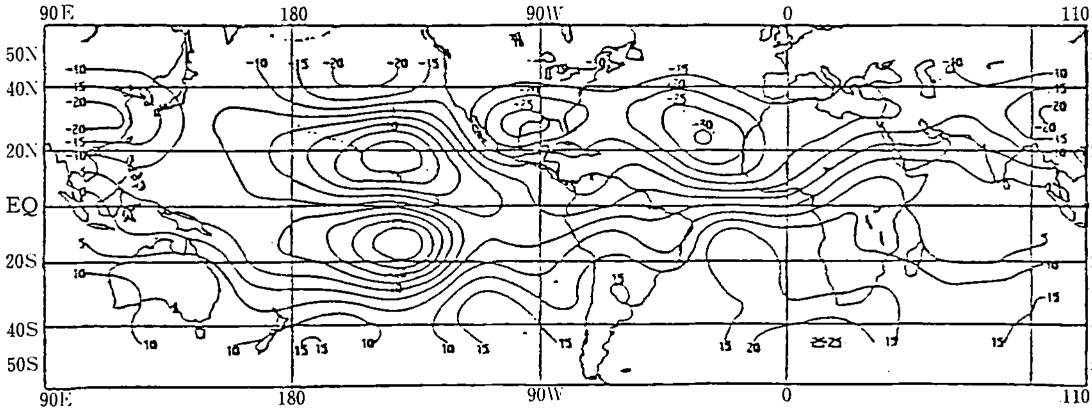
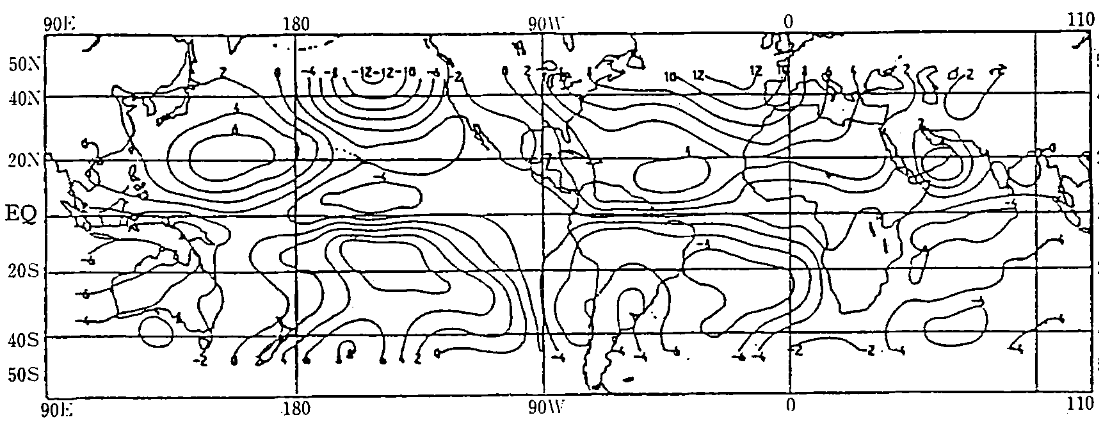
2. Primary elements of a simple, coupled model for modelling ENSO events

Links necessary for air-sea interaction models of the tropics are already well-established, despite the variety of modelling efforts. So, I will summarize briefly some essential points. First, we must clarify how SST affects atmospheric winds. In reality, this process is highly complicated. For example, the relation between anomalous SST and evaporation anomalies is not straightforward. It is affected by some factors such as wind speed, air temperature and humidity. Even the relation between anomalous evaporation zones and anomalous condensation zones is not a simple matter. The advection and convergence of low-level winds play significant roles in determining the above relation. Furthermore, the effect of the anomalous latent heat release on the wind field itself must be clarified. However, there is no doubt that the anomalous SST is somehow strongly related to the heating in the atmosphere. It is clearly necessary that more elaborate future models should take these detailed links into account. Secondly, the process in which winds affect ocean currents, sea levels and depths of the mixed layer is necessary. Thirdly, the relation between SST and oceanic variables must be given. Once one knows all these relations, one can construct a closed set of mechanistic air-sea models.

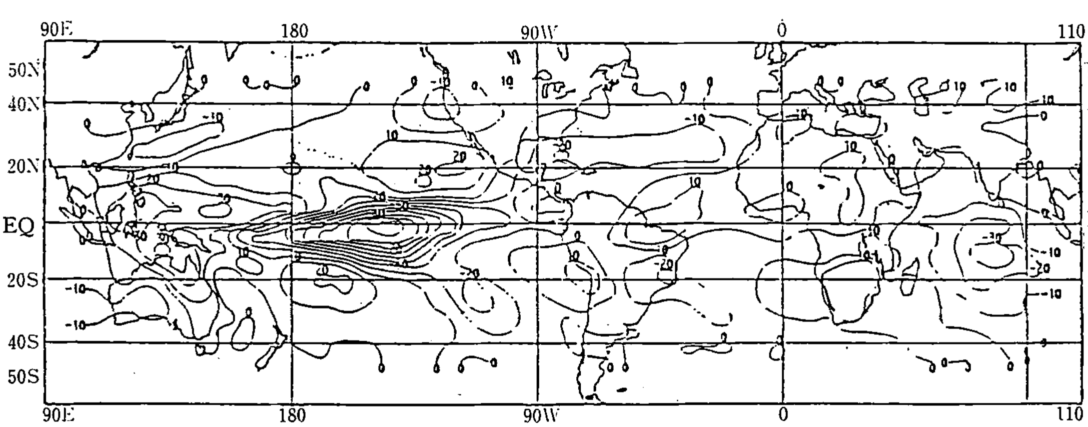
Now, returning to the first point, it is now well-established at the lowest order of approximation that the low-frequency dynamics of the



200mb Stream function anomaly DJF 1982-83



850mb Stream function anomaly DJF 1982-83



OLR anomalies DJF 1982-83

Fig. 1. Stream-function anomaly at 200 mb for the mature phase of 1982/83 event; stream-function anomaly at 850 mb; outgoing longwave radiation anomaly (from Arkin *et al.*, 1983) and low level winds obtained by use of a simple one mode linear model with forcing extending from 10°S to 10°N with the zonal extent of 80° (adapted fram Gill, 1980).

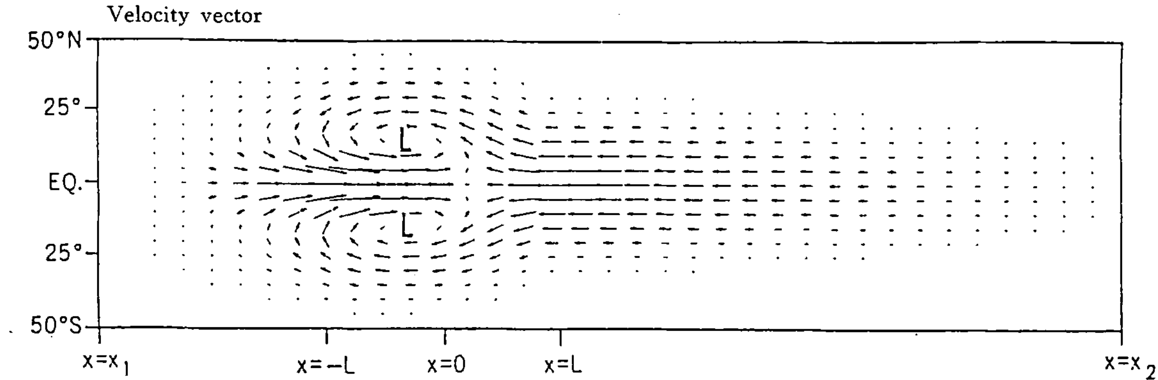


Fig. 1. (continued)

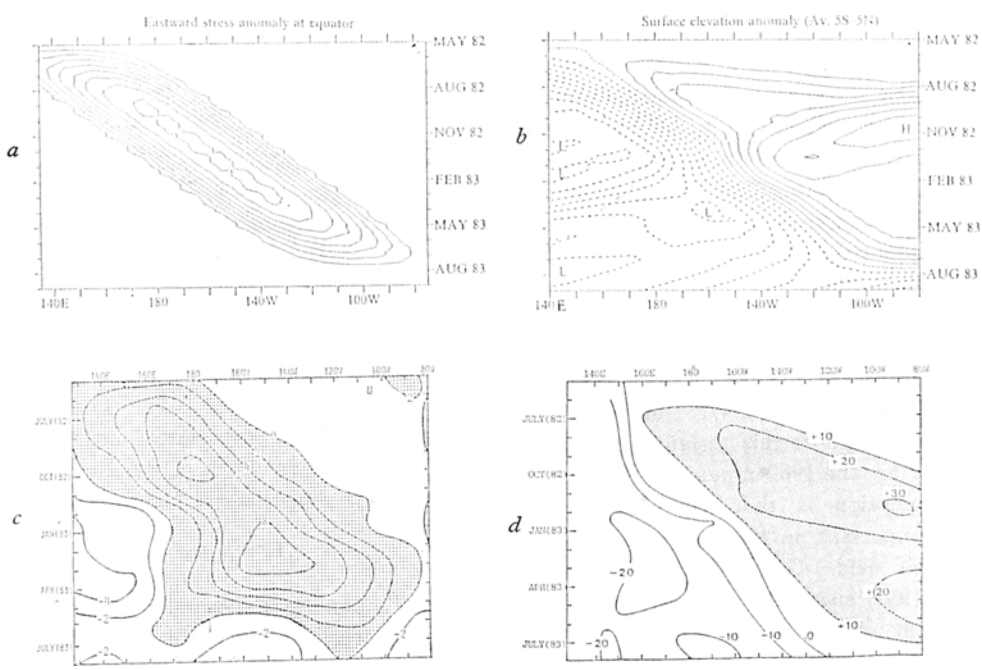


Fig. 2. Longitude-time plots along the equator of *a*, model eastward stress anomaly, *b*, observed eastward stress anomaly, *c*, surface elevation anomaly obtained by a model calculation, *d*, observed surface elevation anomaly (from Gill and Rasmusson, 1983 and Wyrtki, 1984).

tropical troposphere is basically described by use of a simple one-mode linear model when the heating is known (Gill, 1980; Zebiak, 1982). This is mainly because the equatorial heating has a single maximum at about 500 mb and the mixing of momentum associated with the cumulus activity dissipates waves of small vertical wavelength. The first illuminating effort in this meteorological aspect of ENSO studies can be traced back to Matsuno (1966). In Fig. 1, in addition to the simple model result, the stream function anomalies and the OLR anomaly are shown for the mature phase of the 1982–83 ENSO event. Two gyres above the heating anomaly are quite similar to the Matsuno circulation pattern.

As to the second point, there is also no doubt that the major cause of sea level changes and anomalous currents is anomalies in the surface winds. In particular, sea level changes are described well by use of a simple, adiabatic one-mode linear model when the forcing is known

(Wyrtki, 1975; Busalacchi and O'Brien, 1980; Busalacchi *et al.*, 1983). This is mainly because there is a clear-cut thermocline in the equatorial ocean and also because the sea level is an integrated quantity compared to other oceanic variables. The model results and observations for the 1982-83 event are shown in Fig. 2. It is clearly seen that the relation between the westward wind anomaly and the sea level anomaly is simulated by the simple ocean model more than qualitatively. Here, we must notice that time and space scales of the anomalous winds are very important in the adjustment of the upper ocean. The phase speed of the peak in sea level does not necessary coincide with the free oceanic wave speed as far as the forced problem is concerned.

The third and the most controversial issue is what are the processes responsible for the anomalous SST in the tropics. It is very probable that different parameterizations are justified in different regions and under different conditions. The most conventional assumption is that the SST is always in local equilibrium by thermal mixing associated with upwelling and downwelling processes. The depth of the mixed layer is, then, assumed to be positively correlated with the SST. Donguy *et al.* (1984) showed some evidences which support this assumption in the central Pacific for the 1969 minor ENSO event. Another assumption is that the lateral SST advection is important in the generation of the anomalous SST field. Gill (1983) and Harrison and Schopf (1984) suggested the importance of this mechanism at least in the first few months of the 1972 and 1982-83 events. The third assumption is that the anomalous heating or cooling through the ocean surface generates the anomalous SST. Since anomalously high heating appears to be associated with cold water and anomalously high rate of cooling occurs when the SST is high, this mechanism is not the cause but the effect of the anomalous SST. However, it is still important because it provides one of the most important negative feedback mechanisms from a long-term, thermodynamical viewpoint of ENSO.

Thus, there are two conflicting ideas on the process responsible for generating anomalous SST. These two different ideas lead to two extremes of simple coupled models. The basic equations for the atmosphere are:

$$-fV+gH_x=-AU$$
$$fU+gH_y=-AV$$
$$D(U_x+V_y)=-BH-Q$$

where U, V, H and Q denote the zonal velocity, the meridional velocity, the depth and the mass source or sink (which corresponds to the heating rate). A and B are coefficients for Rayleigh friction and Newtonian cooling, respectively. The atmosphere has an equivalent depth D, perturbations to which are measured by H. For Case I the model equations for the ocean are written as

$$u_t-fv+gh_x=-au+\tau^x$$
$$v_t+fu+gh_y=-av+\tau^y$$
$$h_t+d(u_x+v_y)=-bh$$
$$T=\ kh$$

where u, v, h and (τ^x, τ^y) are the zonal velocity, the meridional velocity, the depth perturbation and a body force due to winds. The oceanic motion is damped by Rayleigh friction (the coefficient of which is a) and Newtonian cooling (the coefficient of which is b). The mixed layer temperature T is proportional to the depth of the mixed layer h in Case I, whereas the temperature equation for Case II written as

$$T_t+\bar{T}_x u=-cT$$

Here $\bar{T}_x$ denotes the zonal gradient of the mean temperature field. It turns out that these extremes serve for our understanding of the different aspects in the evolution of ENSO.

The manner of coupling between the atmosphere and ocean must be specified in order to complete the coupled model. We assume the following simple form for the dynamical coupling, that is

$$(\tau^x, \tau^y)=\gamma(U, V)$$

where γ is the coefficient of coupling. The thermodynamical coupling is also simply assumed to be

$$Q=\alpha T$$

where α is the coefficient of coupling.

3. Two extremes of simple, coupled models and their stabilily properties

Once the manner of coupling is specified, it is quite easy to derive a necessary condition for instability. An energy integral of the atmospheric equations leads to

$$\langle gHQ\rangle - A\langle D(U^2+V^2)\rangle - B\langle gH^2\rangle = 0$$

where $\langle\ \rangle$ denotes the integration with respect to x over a wavelength and y from $-\infty$ to ∞. Since we assume that the atmosphere is in equilibrium with the heating, only time-independent terms appear in the expression. A similar integral for the oceanic equations leads to

$$\frac{1}{2}\langle d(u^2+v^2)+gh^2\rangle_t = -a\langle d(u^2+v^2)\rangle - b\langle gh^2\rangle + \gamma\langle d(uU+vV)\rangle.$$

Since the first two terms on the right hand side are negative definite, the positive correlation between atmospheric winds and oceanic flows is necessary for the disturbance to grow. This holds true, irrespective of oceanic models.

The model Case I was first proposed by Philander *et al.* (1984) in connection with the 1982–83 warm event. They found that the oceanic Kelvin wave is destabilized by the air-sea coupling as far as the traditional assumption that the SST is proportional to the depth of the mixed layer is adopted. This instability is due to the asymmetric response of the tropical atmosphere to the heating and also to the asymmetric response of the ocean to the wind forcing. The positive correlation between the winds and currents is possible only for the oceanic Kelvin wave modulated by the coupling (Yamagata, 1985). Therefore, the unstable mode propagates eastward. The growth rate and the dispersion relation are shown in Fig. 3. It can be seen clearly that the coupling between the oceanic Kelvin wave and the Rossby wave leads to the unstable Kelvin wave and the damped Rossby wave. Figure 5 of Philander *et al.* (1984) demonstrates the eastward expansion of the warm anomaly and the positive correlation between the oceanic flows and atmospheric winds.

The model Case II was proposed by Rennick (1983) and Gill (1985). They assumed the lateral advection of SST is important in the generation of the anomalous SST. In this model, the oceanic Rossby wave is destabilized. The advection produces the positive SST anomaly ahead of the oceanic Rossby wave crest at the equator. Thus, the positive correlation between

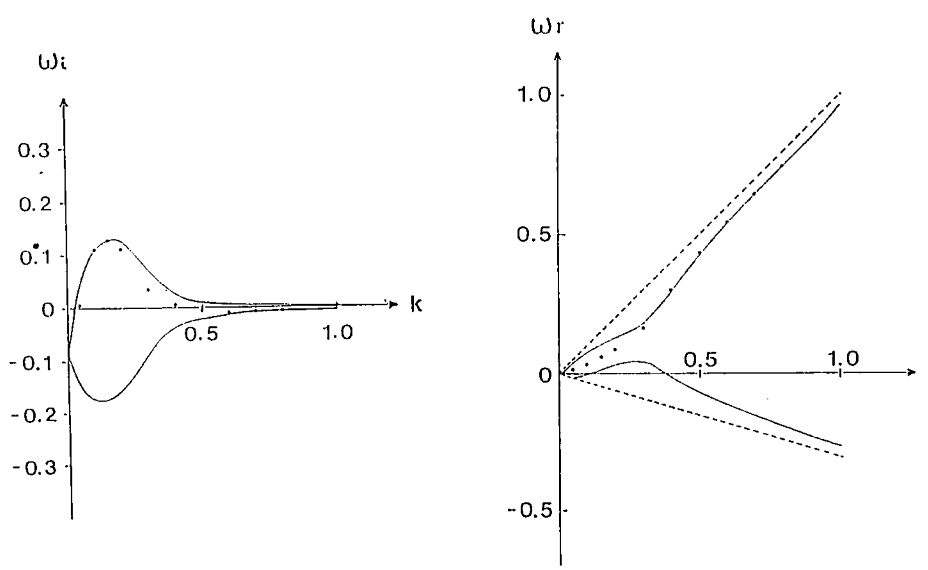


Fig. 3. Real (ω_r) and imaginary (ω_i) parts of frequencies as a function of the wavenumber k for $\alpha\gamma=A=B=0.04$ and $a=b=0$. The oceanic equatorial Rossby radius $\sqrt{C_0/(2\beta)}$ is used as a length scale and $1/\sqrt{2\beta C_0}$ as a time scale, where $C_0(=\sqrt{gd})$ is the oceanic long-wave speed. Dots correspond to numerical results. The broken lines denote the oceanic Kelvin wave and the gravest Rossby wave (from Yamagata, 1985).

the atmospheric response and the oceanic westward motion is possible. Recently, Hirst (1986) calculated the dispersion relation and the growth rate for some realistic values of coupling coefficients. His result confirms the above picture of Case II instability. It should be noticed that the asymmetric response of the atmosphere is not essential in this mechanism. This suggests the oceanic disturbance even with a small meridional extent can grow by virtue of this mechanism. Also one should note the westward phase propagation is not only due to the Rossby wave dynamics but also due to the zonal gradient of the mean SST. As a natural consequence of these arguments, it is possible to derive the simplest form of Case II instability. This can be accomplished by neglecting the meridional velocity and the effect of earth's rotation. The similar argument was done by Rennick and Haney (1986).

Recently, the instability mechanism which is intermediate between the Case I and the Case II was discussed by Hirst (1986). By including the lateral advection of SST and the thermal mixing associated with upwelling and downwelling processes, he found that the strongly coupled, oceanic mode which consists of Kelvin and Rossby waves is destabilized. Because of the strong coupling among oceanic waves, the unstable wave is almost stationary.

4. A simple interpretation of several numerical experiments and the 1982-83 event

Now it is of interest to interprete several, existing model results by use of the simplified idea on the elements of unstable air-sea interaction. The first successful one which follows Philander *et al.* (1984)'s pioneering work is Anderson and McCreary (1985)'s coupled model results. Their model includes the oceanic mixed layer physics and the equation for the temperature of the layer. Therefore it is similar to the intermediate model discussed by Hirst (1986). However, the model results are more easily interpreted in terms of the Case I instability originally proposed by Philander *et al.* (1984) and have much in common with the 1982-83 event, since the temperature of the mixed layer is a monotonically increasing function of the mixed layer depth h. For example, the unstable mode propagates eastward with the propagation speed smaller than the oceanic Kelvin wave. This appears to be due to the weak zonal SST gradient they assumed.

On the other hand, Rennick (1983) and Gill (1985) included only the SST advetion mechanism in the temperature equation. Their models correspond directly to the Case II model in Section 3. Therefore, they found the unstable mode which propagates westward. The relation between the SST anomaly and the mixed layer depth anomaly is consistent with the Case II instability examined by Hirst (1986). Although the authors claim that the model Case II explains some aspects of the warm event, it appears to explain the westward evolution of the cold phase such as the reduction of convection zones rather than the evolution of the warm phase. Very recently, Zebiak and Cane (1985) developed a more elaborate coupled model. Their mode includes the oceanic frictional surface layer and the SST equation. In addition, monthly mean climatologies of SST, surface winds and thermocline depths are prescribed as a basic state. Thus, this is the only model which takes the mean annual cycle into account. The model result shows the interannual evolution of the warm anomaly in the central Pacific. This warm anomaly is almost stationary and has several features in common with the intermediate mode discussed by Hirst (1986). Their model also shows westward propagating disturbances during the cold phase. This disturbance may be ex-

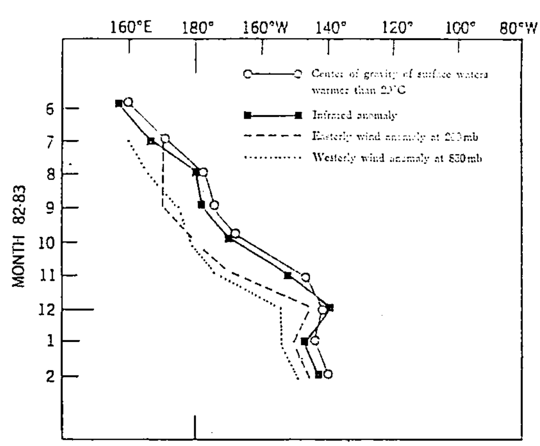


Fig. 4. Eastward expansion of anomalous conditions in the equatorial Pacific (5°N to 5°S) during the 1982/83 event (from Philander *et al.*, 1984).

plained in terms of the Case II instability because the mean zonal temperature gradient takes a large, negative value during the cold phase.

Now the problem is what happens in reality. The best observed 1982-83 event nicely shows us the relationship among the wind anomaly, SST, OLR anomaly and the sea level anomaly (Figs. 2 and 4). The longitude-time plots of these variables suggest that the positive feedback mechanism proposed by Philander *et al.* (1984) actually played an important role in the 1982-83 warm event.

5. What triggers ENSO?

Then, the natural question is what triggers each one of the two complementary phases of ENSO. This question is closely related to the predictability of utmost concern (Cane *et al.*, 1986; Philander, 1986).

As to the warm phase, the right time of seeding in the western Pacific must be when the ITCZ moves across the equator in the late spring and fall (Philander, 1983). This is because the net low-level convergence is necessary for the disturbance to grow. If this disturbance exists initially on the atmospheric side, the burst of westerly wind anomaly associated with the intraseasonal oscillations can be a plausible candidate (*cf.* Luther *et al.*, 1983; Nitta, 1986). This is because the structure of such oscillations is quite similar to the atmospheric component of ENSO except for the time scale (Yamagata and Hayashi, 1984). Therefore the ENSO can be aperiodic and regulated by the seasonal transition of the Asian monsoon. This also partly explains why the simple, phenomenological model such as Nicolls (1984) can reproduce the observed statistical behavior of atmospheric pressure and oceanic SST in the western Pacific. Lau (1985)'s simple stochastic-dynamical model also captures the above aspect of ENSO and suggests that the predictability can be limited by the atmospheric side. On the other hand, if the initial disturbance lies in the western Pacific Ocean, one may trace the signal back much further (possibly one year or even more) because of the long memory on the oceanic side (McCreary, 1983). White *et al.* (1985) suggests the latter was the case in the 1982-83 event by analyzing the heat content in the extratropics. More recently, Inoue *et al.* (1985) have succeeded, to some extent, in hindcasting the observed sequence of events reported by White *et al.* (1985) by use of a one-mode linear model.

In the eastern Pacific, the warm phase occurs when the seasonal migration of the ITCZ takes it to the lowest latitude early in the calender year (Philander, 1983) and when the event in the western Pacific reached the South American coast (Wyrtki, 1975). Thus, in the eastern Pacific, all ENSO appears as an intensification of seasonal phenomenon. Some difference in the evolution of anomalous conditions may be attributed to the difference of arrival time of the event originated in the western Pacific. As suggested in Section 3, the travelling speed may be a complicated function of various variables, especially the depth of the mixed layer and the coefficients of ocean-atmosphere coupling.

As to the cold phase, it must be triggered when the seasonal movement of the convergence zone in the western Pacific revives the heat source over the maritime continent in spite of the mature phase of El Niño. In the eastern Pacific, it is when the reduction in the intensity of the westerly, associated with the revived heat source in the far western Pacific, causes the lower SST. Thus, the cold phase is also initiated and regulated by the strong seasonal signal. Recently Philander (1985) pointed out that the major properties of the cold phase are (1) the westward expansion, (2) spatial scales much smaller than those of El Niño and (3) an intensification of the cold season. The growing mechanism of this cold phase appears to be given by the Case II instability proposed by Rennick (1983) and examined by Hirst (1986). However we need more analyses to pinpoint basic physical mechanisms of this cold phase.

6. Summary and discussion

The stability of air-sea coupled models is especially dependent on the parameterization of SST. One extreme proposed by Philander *et al.* (1984) assumes that the effect of upwelling/downwelling is parameterized in terms of the mixed layer depth. The salient feature is that it gives rise to the unstable, oceanic Kelvin wave associated with Matsuno's circulation pattern in the atmosphere. This model explains many features of the observed warm events. The other extreme proposed by Rennick (1983) and

Gill (1985) assumes that the anomalous SST is determined by the lateral advection of the mean climatology of SST. This latter model gives rise to the unstable oceanic Rossby wave modulated by the air-sea coupling and seems to explain the evolution of the cold phase such as the westward reduction of convection zones spread over the whole equatorial Pacific at the mature phase of the warm event, rather than the evolution of the warm event.

The seasonal movement of the atmospheric convergence zones, which depends on the seasonal heating of the continents, regulates the positive feedback of the air-sea coupled system. This suggests that future models should resolve the seasonal cycle itself (Philander and Rasmusson, 1984). In particular, air-sea-land interaction models are required.

The burst of westerly anomaly which lasts several weeks west of the dateline may initiate the warm events (*cf.* Luther *et al.*, 1983; Nitta, 1986). This suggests that the role played by, say, 30-50 day oscillation as well as the anomalous conditions of the western Pacific Ocean prior to the ENSO events should be clarified in order to explain a detailed triggering mechanism (*cf.* Madden and Julian, 1972; Yasunari, 1980).

During the warm phase the equatorial ocean loses heat to the extra-tropics and to the atmosphere, but during the cold phase the ocean gains heat. As demonstrated by Wyrtki (1984) in this context, the net warm water disappeared from the equatorial Pacific after the 1982-83 event, whereas the slope of the sea level along the equator recovered almost completely. In addition, Yamagata *et al.* (1985) showed that the Kuroshio Extension and the Tsushima Current advect more heat than normal about 1.5 years after the ENSO events. It seems necessary to construct a model which is consistent not only dynamically but also thermodynamically by including roles played by the extratropics.

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エンソの簡単な大気－海洋結合モデルの最近の発展について

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要旨: 簡単なモデルの結果に基づいて，エンソの物理機構の解釈がどの程度まで進展しているかを論じている．特にこれまでに報告された大気海洋結合モデルから二つの極端なケースを取り出し，経年的に変動するエンソの相互に補完的な二つの相を解釈する．加えて，予報技術の改良にも使えるような，より完全なモデル研究に必要とされるものは何かについて議論する．

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