Behaviour of Coupled Modes in a Simple Nonlinear Air–Sea Interaction Model

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ABSTRACT

In this article, a model with the simplest framework is constructed and solved analytically. It is shown that intraseasonal wave coexists with interannual variation, and ENSO cycle arises from coupled Kelvin wave destabilisation. Its irregularity can be related to reasonable model parameters. External processes have some important effects on nonlinear air–sea interaction modes.

Key words: Nonlinear, Stationary coupled instability, External forcing, Multiscale

I. INTRODUCTION

The phenomenon of El Nino and the Southern Oscillation (ENSO) is an important climatic variation. It is a quasi–cyclic oscillation with a time scale of 2–7 years. Now it is generally understood that ENSO arises through interaction between the tropical atmosphere and ocean. It was based on a considerable number of researches, such as observation analyses (Bjerknes, 1966; Rassmusson and Carpenter, 1982; Wang, 1992), dynamical studies (McWillians and Gent, 1978; Lau, 1981; Philander et al., 1984; McCreary and Anderson, 1984, 1985, 1992; Hirst, 1986; Lau, 1988) and numerical experiments (Cane and Zebiak, 1987; Schopf and Suarez, 1988). Full coupled GCM simulations have also been widely adopted. Some researchers are able to predict ENSO events to some extent. Some relationships between ENSO and other phenomena, such as intraseasonal oscillation (ISO) and monsoon, are also indicated by some scientists (Li, 1990; Yasunari, 1990; Webster et al., 1992; Li and Zhou, 1994). Although great progress has been made in the last two decades, there are still many problems as yet to be fully understood, for example:

(1) What is the mechanism of ENSO cycle? One hypothesis suggests that oceanic waves and ocean boundary play an important role in determining the time scale of ENSO. Suarez and Schopf (1988) firstly proposed a possible conceptual model "delayed–oscillator" to explain the model results. Another hypothesis claims that ENSO is a self–excited oscillation in air–sea interaction (Munnich, et al., 1991). Therefore, it is important for understanding the mechanism of ENSO to study the behaviour of coupled modes in nonlinear air–sea interaction.

(2) Why is ENSO an irregular cycle with a time scale of 2–7 years? How about nonlinearity affecting the irregularity of the ENSO time scale? In the air–sea interaction system, there are different time scale modes. How do they behave in the coupled system?

Some clues to answer the above problems are suggested in this paper. The paper is arranged as: a linear anomaly atmospheric model and a nonlinear oceanic anomaly model are described in Section 2. The model results and their analyses are given in Section 3. Section 4 describes the characteristic of stable air–sea coupled waves. Section 5 is devoted to the summary and discussion.

II. MODEL AND ITS PHYSICS

According to the observed symmetrical features about the Equator in ENSO events and the dynamic characteristics of planetary scale fluid in the equatorial region, the simplest models of atmosphere and ocean are employed. The models are similar to those of Lau (1981) except nonlinear terms canot be neglected. The essence of this work is to reveal the characteristics of the nonlinear model and the interannual and intraseasonal variability in this model.

(1) The atmospheric model

The heating–excited tropical atmospheric circulation can be well defined by a model with only one–baroclinic layer (Gill, 1980). The equations of heat–induced linear equatorial Kelvin wave in the tropical atmosphere are

$$U_t + gH_x = -AU, \tag{1}$$

$$H_t + DU_x = -BH - Q, \tag{2}$$

where coordinates (t,x) measure time and distance in the eastward direction; U is the horizontal velocity in the x direction, H is the deviation from D, the climatic mean of equivalent depth of one–layer atmosphere. Q is the heating source function and A, B are the coefficients of Rayleigh friction and Newtonian cooling in the atmosphere, respectively.

(2) The oceanic model

The upper ocean can be reduced to a one–layer shallow water model driven by wind. Such wind–driven reduced model has been widely adopted in spite of its limit capability (Philander, et al., 1984)

The oceanic model equations are

$$u_t + uu_x + gh_x = -au + F, \tag{3}$$

$$h_t + uh_x + (d+h)u_x = -bh, \tag{4}$$

where F is the wind stress that atmosphere acts on ocean, d the average climatic value of mixed–equivalent depth. The other symbols are in common use.

(3) The coupled model

The manner of coupling between atmosphere and ocean is taken from Philander, et al (1984) and Battisti, et al. (1989) where more detailed validity and physical background were discussed. Simple assumption is adopted so that the heat flux from the ocean to atmosphere is parameterized as

$$Q_1 = \alpha(h - \kappa h^3). \tag{5}$$

and external heating function is parameterized as

$$Q_2 = \alpha Q_{ext}. \tag{6}$$

The wind stress acts as a body force in the simple form

$$F = \gamma U. \tag{7}$$

Considering the adjusting time scale for atmosphere to ocean heating is much small, we neglect the time derisive term of atmospheric equations. Rayleigh friction and Newtonian cooling both in atmosphere and ocean are also ignored in order to solve the problem more conveniently.

The air–sea coupled equations we will study next are

$$DU_x = -\alpha(h - \kappa h^3) - \alpha Q_{ext} \,, \tag{8}$$

$$u_t + uu_x + gh_x = \gamma U \,, \tag{9}$$

$$h_t + uh_x + (d+h)u_x = 0 \,. \tag{10}$$

(4) Model physics

It is generally known that ENSO is a kind of internal variability in the atmosphere–ocean system, and nonlinear bifurcation resulting from coupled instability leads to a nonlinear oscillation whose period is the underlying time scale of ENSO. Our model is the simplest self–organized system with three independent variables.

The positive feedback process, which is formed by wind stress acting on ocean and the heating from ocean to atmosphere can be described as the following: positive (negative) h will cause atmospheric convergence (divergence) through term$-\alpha h$, in the meantime, oceanic convergence (divergence) will further intensify positive (negative) h. A negative feedback, which is formed by the gravitational restoring force connecting with the stratification of the ocean and the nonlinear term $\alpha\kappa h^3$ (Battisti, et al, 1989). The process can be described as next: in (9) the gravitational restoring force leads to produce divergence (convergence) in the maximum (minimum) value region of h, in (10) the divergence (convergence) leads to reduce the value of h; on the other hand, nonlinear term $\alpha\kappa h^3$ also plays a role in the negative feedback.

Now the linearized version of (8), (9) and (10) is

$$\begin{cases} DU_x + \alpha h = 0 \,, \\ u_t + gh_x = \gamma U \,, \\ h_t + du_x = 0 \,, \end{cases} \tag{11}$$

and the solution of (11) can be found in the form

$$\begin{cases} h = \mathrm{Re}\{h_0 \exp[i(kx - \omega t)]\} \,, \\ u = \mathrm{Re}\{\gamma_1 h_0 \exp[i(kx - \omega t)]\} \,, \\ U = \mathrm{Re}\{\gamma_2 h_0 \exp[i(kx - \omega t)]\} \,. \end{cases} \tag{12}$$

Substitution of (12) into (11) yields a dispersion relation

$$\omega_{1,2} = \pm kC_0 \left(1 - \frac{\alpha\gamma}{k^2 C_a^2}\right)^{\frac{1}{2}} \,. \tag{13}$$

which is the same as that of Lau (1981).

When

$$\alpha\gamma > k^2 C_a^2 \,, \tag{14}$$

the stationary instability occurs and the air–sea system is unstable due to the coupling process.

III. MODEL RESULTS AND ANALYSES

In order to obtain analytic solutions of Eqs.(8)–(10) we consider a case of critical instability. The critical value of the coupling strength is $(\alpha\gamma)_c = k^2 C_a^2$. If linear unstable condition (14) is satisfied as

$$\gamma = (1+\delta^2)\gamma_c \ , \tag{15}$$

$$k = \left(\frac{\alpha\gamma_c}{C_a}\right)^{\frac{1}{2}} \ , \ 0<\delta \ll 1 \ , \tag{16}$$

where δ is the small parameter, then the frequency of critical wave is

$$\omega_{1,2} = \pm i\delta k_c C_0 \ . \tag{17}$$

This gives a slow time scale for the amplitude development of the unstable mode. Because the growth is slow, it is possible for weak nonlinearity to counterbalance it and reach finite amplitude solutions by the multi-scale method. The slow time scale $T=\delta t$, where δ is the ratio of perturbation h to average value of equivalent depth of ocean.

Selecting time scale $T=\frac{L}{C_0}$, where L is the basin width of the ocean, C_0 is the Kelvin wave speed. Nondimensional quantities are defined by

$$(H,h) = \frac{C_0^2}{g}(H',h') \ , \tag{18}$$

$$(U,u) = C_0(U',u') \ . \tag{19}$$

The dimensionless form of Eqs.(8)–(10) is

$$\begin{cases} U_x + \alpha_1 h = \alpha_1 \kappa h^3 - \alpha_2 Q_2 \ , \\ u_t + h_x - \gamma_c U + u u_x = 0 \ , \\ h_t + u_x + (uh)_x = 0 \ . \end{cases} \tag{20}$$

Asymptotic solutions of (20) are sought in the form

$$E = \delta(E_0 + \delta E_1 + \delta^2 E_2 + \ \cdots \ )E = \{U,u,h\} \ ,$$

where

$$\alpha_1 = \alpha T C_0^2 / C_a^2 \ ,$$
$$\alpha_2 = \alpha T \overline{Q} / D \ ,$$
$$\gamma_1 = \gamma T \ ,$$

and assuming $\alpha_2 \sim \delta^3$ and $Q_2^* = \mu F(T)\cos k_c x$, dropping the primes " ' " and stars " * ", the equations finally become

$$\begin{cases} U_x + \alpha h = \alpha\kappa h^3 - \delta^3 \mu F(T)\cos k_c x \ , \\ \left(\frac{\partial}{\partial t} + \delta\frac{\partial}{\partial T}\right)u + h_x - \gamma_c U + u u_x = \delta^2 \gamma_c U \ , \\ \left(\frac{\partial}{\partial t} + \delta\frac{\partial}{\partial T}\right)h + u_x + (uh)_x = 0 \ . \end{cases} \tag{21}$$

The solutions of $O(\delta)$ and $O(\delta^2)$ problems are

$$\begin{cases} u = \delta u_0 + \delta^2\left[c_1(T) - \frac{A}{k}\sin k_c x - u_0 A_0 \cos k_c x\right] + O(\delta^3) \ , \\ h = \delta A_0 \cos k_c x + \delta^2 A_1 \cos k_c x + O(\delta^3) \ , \\ U = -\delta\frac{k}{\gamma_c}(A_0 + \delta A_1)\sin k_c x + O(\delta^3) \ . \end{cases} \tag{22}$$

Equations of $O(\delta^3)$ problem are

$$\begin{cases} h_{2x} - \gamma_c U_2 = -u_0 u_{1x} + \gamma_c U_0 - U_{1T}, \\ u_{2x} = -u_0 h_{1x} - h_0 u_{1x} - h_{1T}, \\ U_{2x} + \alpha h_2 = \alpha \kappa h_0^3 - \mu F(T) \cos k_c x. \end{cases} \tag{23}$$

The restriction of removing resonance from (23) yields the Eq. of amplitude evolution

$$\frac{d^2 A_0}{dT^2} - k^2 A_0 + \frac{3}{4} \kappa k^2 A_0^3 - \mu F(T) = 0, \tag{24}$$

which is the famous forced Duffing equation. When there is no external forcing, the equilibrium of the amplitude is $\overline{A_0} = \left(\frac{3}{4}\kappa\right)^{\frac{-1}{2}}$. Let $A_0 = \overline{A}_0 A$, the equation of (24) is

$$\frac{d^2 A}{dT^2} - \kappa^2 A + \kappa^2 A^3 - G(T) = 0, \tag{25}$$

$G(T) = \mu \cos\left(\frac{2\pi}{\omega_f}\right) t$, $\omega_f = 1$ year or 2 years, which is external forcing. The external forcing processes consist of, for example, intraseasonal variation, annual cycle of intraseasonal oscillation and quasi–biennial variability of Asian Monsoon, et al. Numerical results of (25) are calculated by the four–order Rung–Kutta method. Standard parameters are listed in Table 1. Some fundamental results are shown as following:

Table 1. Basic Parameters and Their Values in the Model

Symbol	Parameter	Standard values
g	gravity parameter	9.8 ms^{-1}
d	equivalent depth of atmosphere	440 m
d	equivalent depth of ocean	20 cm
δ	small parameter	0.01
L	basin width of ocean	15000 km
κ	nondimensional parameter	0.1

(1) No external forcing

In this case, the air–sea interaction system is an self–organized system. It is shown from Fig.1 that the period of oscillation varies with the coupling strength. When $\alpha\gamma = 10 \times 10^{-10}$ s^{-2}, the period is 4 years or so. When $\alpha\gamma = 165 \times 10^{-10}$, the period becomes less than 3 years; when $\alpha\gamma = 165 \times 10^{-11}$, the period is 8 years. The period decreases with the coupling strength. The reason is that when the coupling strength increases, critical wavenumber k becomes larger, the recovering term $\alpha\kappa h^3$ in the amplitude evolution becomes more dominant, which leads to shorter period. In this case, the air–sea interaction system can be regarded as a "Pacific oscillator", which is suggested by some authors.

(2) Annual external forcing

In the air–sea interaction system, annual cycle and internal oscillation of the system coexist, though the mechanism of annual cycle is not clear now. In this case, we assume that the internal oscillation of air–sea interaction system is distinguished from annual cycle, the coupling strength is $\alpha\gamma = 7.875 \times 10^{-10}$ s^{-1}. Fig. 2 shows that the period of oscillation varies with the strength of external forcing. Fig. 2a is the case with no external forcing, the period is

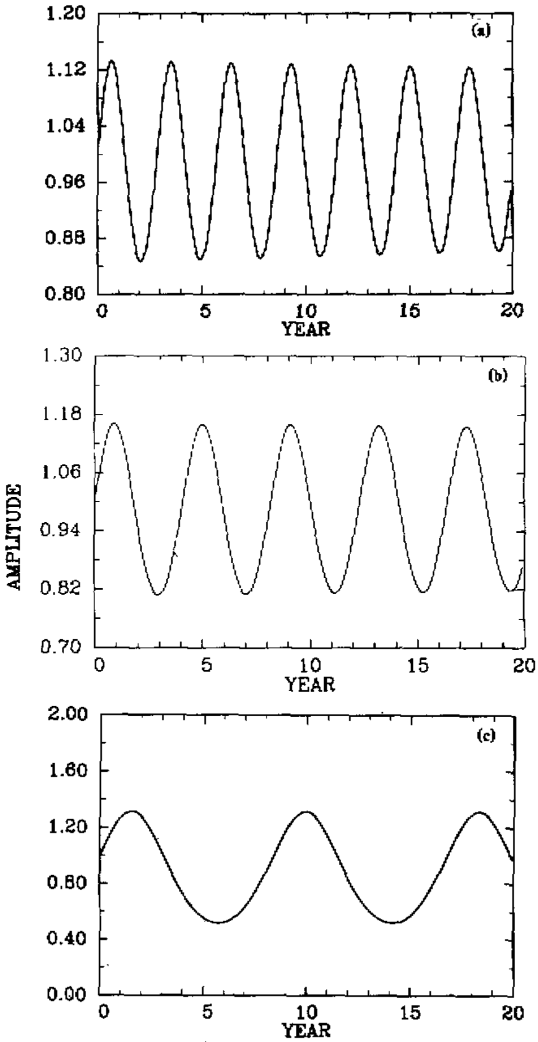

Fig. 1. The period of oscillation varies with coupled intensity in the cases for $\mu = 0.0$ and $\alpha\gamma = 1 \times 10^{-10}$ s^{-1} (a), $\alpha\gamma = 5 \times 10^{-11}$ s^{-1} (b), $\alpha\gamma = 1 \times 10^{-11}$ s^{-1} (c).

about one year. Fig. 2b shows the amplitude evolution in the case of $\mu = 3.0$, the period is the same of Fig. 2a, but the amplitude is modulated with a period of ten years or so. Figs. 2d–2f show irregular oscillation arising from the nonlinear subharmonic resonance between the air–sea interaction system mode and annual cycle. It is known that annual cycle and variability with a time scale of one year exist in the atmosphere and ocean. Some observational researches have pointed out that annual cycle is closely related to interannual variability and ENSO, for instance, annual cycle seems to play a role in modifying El Nino (Rasmussen and Carpenter, 1982). Our results further show that annual cycle forcing can influence significantly the air–sea coupled oscillation.

(3) Biennial external forcing

For the couple coefficient, $\alpha\gamma = 1.97 \times 10^{-10}$ s^{-1}, it is learnt from Fig. 3 that biennial external forcing can excite the oscillation, of which period is larger than two years, and showing the air–sea system a multi–scale system (Wu and Anderson, 1993). The observed SSTA also has such multi–scale variability (Fig. 4). We can therefore suggest that ENSO cycle is related to the nonlinear interaction among QB component, annual cycle and very low–frequency variability. Our model results provide an alternative explanation of the spectrum of ENSO.

IV. COUPLE MODES IN THE STABLE CONDITION

In the condition that (14) is not satisfied, the air–sea system is stable. Next we will study

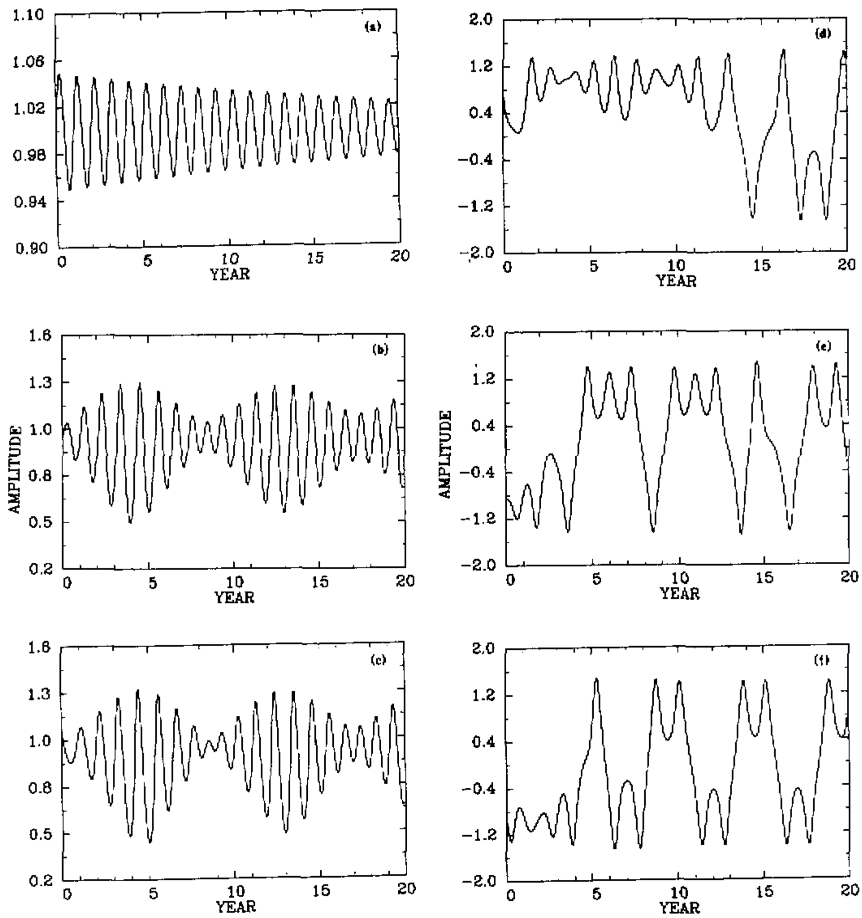

Fig. 2. When $\omega_f = 1$ year, the period of oscillation varies with the intensity of external forcing in the case of different μ. The Figs. 2a–2f show respectively the different cases with $\mu = 0.0$, 3.0, 3.25, 5.0, 5.25 and 5.5.

the characteristics of the system under this condition. External forcing is neglected.

Let $\theta = kx - \omega t$, the solutions of Eqs.(8)–(10) are sought in the form

$$E = \delta[E_0(\theta) + \delta E_1(\theta) + \delta^2 E_2(\theta) + \cdots] , \quad 0 < \delta \ll 1 , \tag{26}$$

where

$$E = \{u, h, H\} .$$

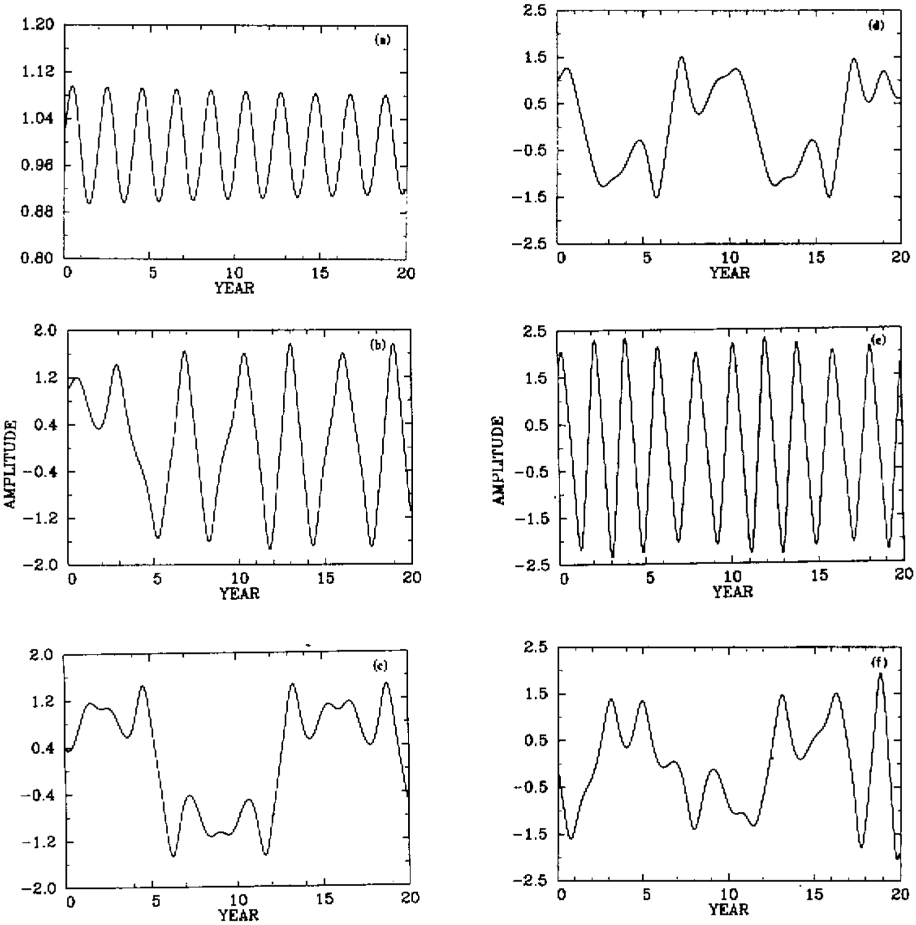

Fig. 3. When ω_f = 2 year, the periods of oscillation vary with the intensity of external forcing in the cases of different μ. The Figs.2a–2f show respectively the different cases with μ = 0.0,3.0,3.4,5.0,5.3 and 6.0.

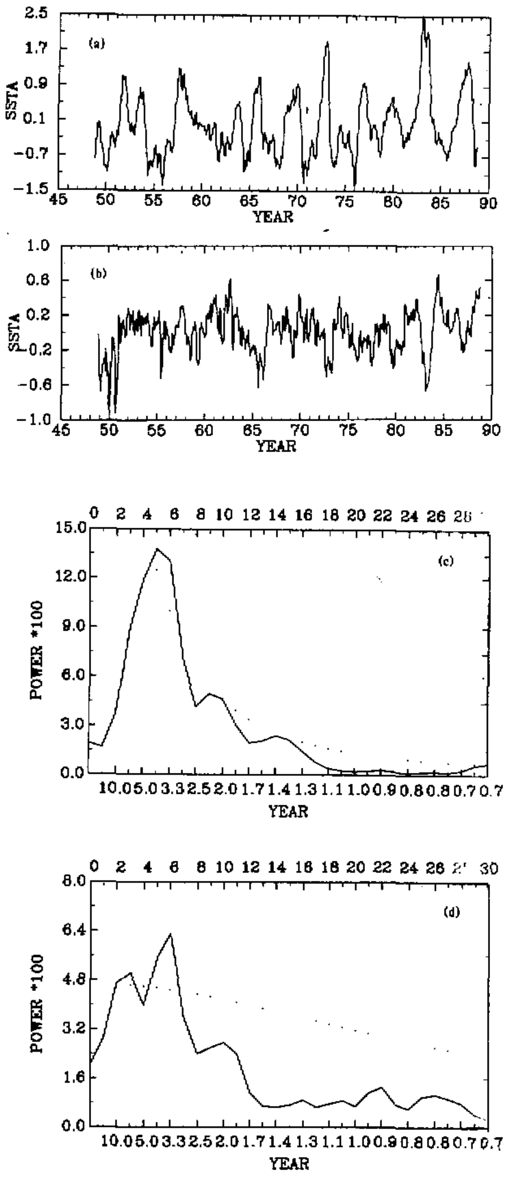

Fig. 4. The monthly SSTA evolution and power spectrum for Nino3 region (a,c) and Eq.–W. Pacific (b,d) from 1948–1988.

The solutions of order $O(\delta)$ and $O(\delta^2)$ are

$$\begin{cases} h = \delta(d+\delta A_1)\cos\theta - \delta^2(K^2-1)d\cos(2\theta) + o(\delta^3), \\ u = \delta(d+\delta A_1)\frac{\omega_0}{kd}\cos\theta - \delta^2(K^2-\frac{1}{2})\frac{\omega_0}{k}\cos(2\theta) + o(\delta^3), \\ U = -\delta\left(d+\frac{\delta A_1}{d}\right)\frac{\alpha d}{Dk}\sin\theta + \delta^2(K^2-1)\frac{\alpha d}{2kD}\sin(2\theta) + o(\delta^3), \end{cases} \tag{27}$$

where $\omega_0 = \frac{C_0}{C_a}[\alpha\gamma(K^2-1)]^{\frac{1}{2}}$, $K = kC_a(\alpha\gamma)^{-\frac{1}{2}}$, A_1 may be taken as zero here.

Problem on order $O(\delta^3)$ is

$$\begin{cases} DU_{2x} + \alpha h_2 = \alpha\kappa h_0^3, \\ u_{2t} + gh_{2x} - \gamma U_2 = -d(u_0u_1)_x - \frac{\omega_2\omega_0}{k}\sin\theta, \\ h_{2t} + du_{2x} = -(u_0h_1+u_1h_0)_x - d\omega_2\sin\theta. \end{cases} \tag{28}$$

After series of complicated but standard algebra processes, we obtain

$$\omega_2 = -\frac{3}{4}\left(K^2-\frac{2}{3}\right)\omega_0 + \frac{3d^3\alpha\gamma\kappa}{8D\omega_0}, \tag{29}$$

where $T_1 = t$, $T_2 = \delta^2 t$.

There are oscillations on the primitive time scale and second slow time scale. The order of periods can be defined as $\frac{2\pi}{\omega_0}$ and $\frac{2\pi}{\delta^2\omega_2}$. The phase speeds on the two time scales are defined as

$$C_0 = \frac{\omega_0}{k}, \tag{30}$$

$$C_2 = -\frac{3}{4}(K^2-\frac{2}{3})C_0 + \frac{3d^3\alpha\gamma\kappa}{8D\omega_0 k} \tag{31}$$

Fig. 5 depicts T_0 (solid line, scaled by day) and T_2 (dashed lined, scaled by month) varying with coupled intensity. Fig. 6 depicts C_0 and C_2 (varying with coupled intensity). In this air–sea coupled system, there is multiscale behaviour under a rich variety of parameters: large–scale intraseasonal oscillation on the primitive time scale and 2–10 years interannual oscillation on the second slow time scale. For the first time scale, C_0 is almost in agreement with gravity wave speed, which is eastward propagating. On the second slow time scale, a kind of slow westward motion exists, its phase speed is comparable with the observed SSTA westward development in the stage of relaxing of El Nino events. The physical diagram is disturbance propagating eastward on the ordinary time scale at the speed of 1.0—1.4 ms^{-1}, but evolves westward at the speed ranging from 4×10^{-3} to 1×10^{-3} ms^{-1}.

V. SUMMARY AND DISCUSSION

In this article, we have discussed the behaviour of coupled mode in a simple nonlinear air–sea interaction model. Some results can be summarized as follows:

(1) Stationary coupled instability can explain the spatial and temporal characteristics of the onset of ENSO. The bifurcation resulting from "stationary coupled instability" leads to

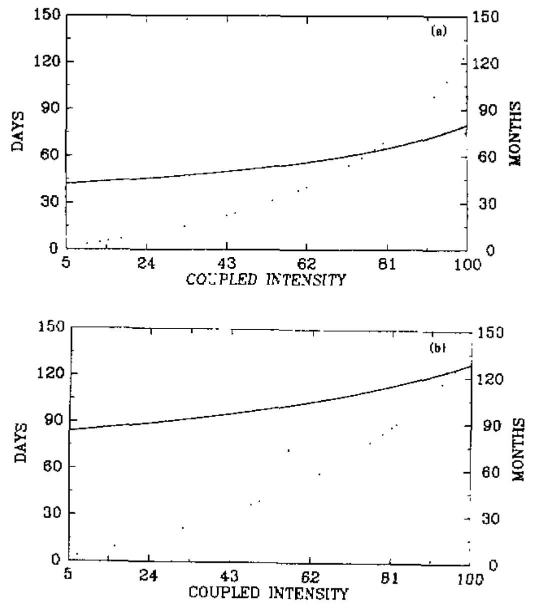

Fig. 5. T_u (solid line, scaled by day) and T_2 (dashed line, scaled by month) vary with coupled intensity for L = 5000 km and coupled intensity scaled by $5\times10^{-11}s^{-1}$ (a); for L = 10000 km and coupled intensity scaled by $1\times10^{-11}s^{-1}$ (b)

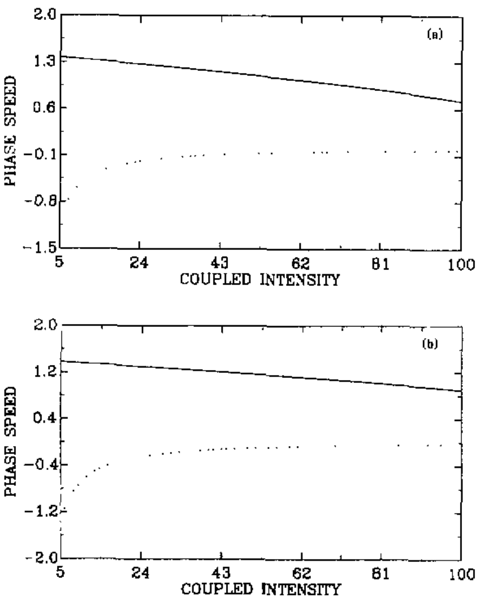

Fig. 6. C_u (solid line, unit: ms^{-1}) and C_2 (dashed line, scaled by 100 ms^{-1}) vary with coupled intensity for L = 5000 km and coupled intensity scaled by $5\times10^{-11}s^{-1}$ (a); for L = 10000 km, coupled intensity scaled by $1\times10^{-11}s^{-1}$ (b).

nonlinear oscillation. The time scale of oscillation is determined by coupled intensity, which is related to atmospheric and oceanic states. But the self–excited oscillation is different from ENSO cycle, although its period is consistent with the average time scale of ENSO.

(2) External forcing with low–frequency time scale(such as annual cycle or QBO) can clearly modify the nonlinear air–sea coupled modes and leads it to be irregular as well as ENSO cycle, especially as the observed SSTA series. Therefore, apart from the air–sea coupled interaction , external forcing is also important to cause the ENSO cycle. The result in this paper is partly consistent with that in Masumoto and Yamagata (1991) in which external forcing heating is emphasized.

(3) In the stable case, the coupled model results also show the existence of the multi–scale oscillations . There is a kind of nonlinear oscillation mode on the second slow time scale, which can be related to ENSO cycle and westward expansion of disturbance in the recovering stage of ENSO. For the primitive time scale, intraseasonal mode coexists in the system.

The results here are limited for model's simplicity, model with more detailed physical processes should be developed to test the results mentioned above.

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