NOTES AND CORRESPONDENCE

Interannual Positive Feedbacks in a Simple Extratropical Air–Sea Coupling System

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ABSTRACT

A simple theoretical analysis identified three possible interannual positive feedbacks in the extratropics: the upwelling mode, the SST–Sverdrup mode, and the SST–evaporation mode. The upwelling mode becomes unstable when the atmosphere responses to a warm SST anomaly predominantly with a high surface pressure. In contrast, the SST–Sverdrup mode is destabilized when the atmosphere responses to a warm SST with a low pressure. In the region of mean westerly wind, the SST–evaporation mode is unstable when the atmospheric response to a warm SST is a quarter-wavelength to the south. The upwelling mode seems to favor low-latitude regions, while the two SST modes seem to favor midhigh latitudes. It is suggested that the relative position of the stationary atmospheric response to anomalous SST is of crucial importance for the extratropical ocean–atmosphere interaction.

1. Introduction

Large-scale persistent atmospheric and SST variability has been observed in the extratropics three decades ago (e.g., Namias 1965, 1972; Bjerknes 1964). This variability is extensively explored in many observational and numerical works in the last decade (Palmer and Sun 1985; Wallace and Jiang 1987; Wallace et al. 1990; Pitcher et al. 1988; Lau and Nath 1990; Miller 1992). It is found that the variability has time scales from interannual to decadal and exhibits the most salient feature in winter times. Furthermore, the atmospheric variability is well correlated to anomalous SST in the extratropics as well as in the tropics.

Two approaches have been proposed to study the persistent climatological anomaly. The first suggests that the ocean acts to rectify short time-scale (synoptic) atmospheric forcing, generating SST anomalies whose variance grows with time (Hasselmann 1976; Frankignoul and Hasselmann 1977).

The second emphasizes the positive feedback in the coupled ocean–atmosphere system (Namias 1965, 1972; Bjerknes 1964; Palmer and Sun 1985; Wallace et al. 1990). One reason for this argument is the fact that the observed climate variability has large-scale coherent structures, which seem unlikely to be entirely caused by random processes.

Furthermore, observations show that some extratropical atmospheric variability is correlated to extratropical SST anomalies more than to tropical SST anomalies, particularly for those with time scales longer than El Niño (Namias et al. 1989; Lau and Nath 1990; Wallace et al. 1990). Recent GCM simulations also suggest that SST anomaly in the tropics cannot explain the extratropical variability satisfactorily unless extratropical SST anomaly is also considered (Pitcher et al. 1988). In addition, there is evidence that SST anomalies in the extratropics can be forced by extratropical atmospheric variability (Davis 1976, 1978). All of these suggest the existence of air–sea interaction in the extratropics, which is not directly related to the tropical air–sea coupling, and which may contribute significantly to the climate variability found in the extratropics.

In the extratropics, although positive feedbacks have been proposed long ago (Bjerknes 1964; Palmer and Sun 1985; Wallace et al. 1990), there has been little theoretical study. In fact, some works suggest the absence of positive feedbacks in the extratropics (e.g., Philander et al. 1984; Yamagata 1985; Frankignoul 1985).

In this paper, we attempt to present a preliminary theoretical analysis of some positive feedbacks in the extratropics. The questions that we are particularly interested in are the following: Is there any positive feedback in the extratropical atmosphere–ocean system? What is the physical mechanism? What is the typical time scale?

The paper is arranged as follows. The next section will introduce the model. Section 3 discusses the basic physics of three types of positive feedbacks: the upwelling mode, the SST–Sverdrup mode, and the SST–evaporation mode. These feedbacks occur for different atmospheric responses to SST anomaly. The last section summarizes the results and explores some further aspects of extratropical air–sea interaction.

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2. The model

To highlight the physics of positive feedback, a simple, linear, coupled ocean–atmosphere model will be used.

a. The atmosphere model

Following Neelin (1991), we write the general form of a linear, steady surface atmospheric response to a SST anomaly $T(x, y, t)$ as

$$\Psi(x, y, t) = AT(x, y, t),$$
$$\text{and} \quad U = -\partial y\Psi, \quad V = \partial_x\Psi. \quad (2.1)$$

Here, the complex $A = A_r + iA_i$ is the thermal coupling coefficient, which determines the relative position of low-level atmospheric response to SST anomaly. The relative position of atmospheric response to SST will be seen to be of crucial importance in the instability analysis. Unlike in the tropics, where the atmosphere response to a warm SST is dominated by low pressure (i.e., $A \approx A_r < 0$), the extratropical atmospheric response to anomalous SST is extremely complicated. A somewhat detailed discussion will be presented in section 4. In my opinion, the parameterization of stationary atmospheric response to SST anomaly is the major obstacle for studying extratropical air–sea coupling. Considering our present stage of understanding, it seems wise to adopt a general atmosphere model (2.1), allowing any relative position to the SST.

Due to the lack of a reliable atmosphere model, it seems unlikely that we can determine the magnitude of the thermal coupling coefficient A from typical atmospheric parameters. Instead, here we diagnostically estimate A from observation. Geostrophy gives the relation between the streamfunction in (2.1) and the height of isobar as $H = \Psi g/f$, where f and g are the Coriolis parameter and the gravitational acceleration. Thus, we have $\delta H = Ag\delta T/f$, or $A = \delta Hf/g\delta T$. Observations show that a SST anomaly of $\delta T = 1°C$ forces an atmospheric response with typically $\delta H = 10$ m on an isobar surface in the low troposphere. This gives the estimate in Table 1 as $O(A) = 10^6$ m^2/°C s.

b. The ocean model

The ocean is a 1.5-layer quasigeostrophic model. The anomalous wind stress will be linearly proportional to the wind speed as $(\tau_x, \tau_y) = \gamma(U, V) = \gamma(-\partial_y\Psi, \partial_x\Psi)$, where γ is the momentum coupling coefficient. Thus, the potential vorticity equation for planetary-scale motion reads

$$(\partial_t + K_m)\left(-\frac{\psi}{L_0^2}\right) + \beta\partial_x\psi = \gamma\nabla^2\Psi, \quad (2.2)$$

where ψ is the streamfunction for geostrophic oceanic current; K_m represents the vorticity damping; $L_0^2 = g'h_0/f^2$ is the oceanic deformation radius, with g' and h_0 being the reduced gravity and the mean thermocline depth.

TABLE 1. Standard parameters used in scaling analysis.

Parameter	Meaning	Value
$\bar{T}_y$	Mean SST gradient	1°C/100 km
$\bar{Q}_E$	Mean evaporation	200 W m^{-2}
$\bar{U}$	Mean zonal wind	10 m s^{-1}
h_m	Mean mixed-layer depth	100 m
h_0	Mean thermocline depth	500 m
L	Meridional scale of the disturbance	1000 km
c_p	Heat content	4. × 10^3 J (kg)$^{-1}$ K^{-1}
ρ_0	Density of water	1. × 10^3 (kg) m^{-3}
g	Gravity acceleration	10 m s^{-2}
f	Coriolis parameter	10^{-4} s^{-1}
β	df/dy	1.2 × 10^{-11} (m s)$^{-1}$
K_h/K_T		1°C/30 m
γ	Momentum coupling coefficient	1/100 days
A	Thermal coupling coefficient	10^6 m^2/°Cs

Note: A is estimated at the end of section 2a; $\bar{T}_y$ is for the maximum frontal regions such as the Gulf Stream; $\bar{Q}_E$ is estimated from Cayan (1990); K_h/K_T is estimated at the end of section 3a.

Taking into account the evaporation cooling effect, the linearized Kraus–Turner mixed-layer model yields the perturbation SST equation (Xie 1992, personal communication)

$$(\partial_t + K_T)T + v\bar{T}_y + E\bar{U}U - K_hh = 0,$$
$$\text{with} \quad E = 2\bar{Q}_E/\rho_0c_ph_m\bar{U}^2 > 0. \quad (2.3)$$

Here, $\bar{T}_y$, $\bar{U}$, and $\bar{Q}_E$ are, respectively, the mean meridional SST gradient, mean wind, and mean evaporation; K_T is the thermal damping coefficient; $K_h > 0$ is a coefficient representing the strength of upwelling. Other parameters are explained in Table 1. Equation (2.3) states that SST can be changed by advection, evaporation, and upwelling. Here, only the meridional advection is retained because in the extratropics the mean zonal SST gradient is much weaker than the meridional component (Levitus 1982). In the evaporation term, the mean meridional wind is neglected. Without loss of generality, hereafter we consider the Northern Hemisphere. Since the mean temperature decreases northward ($\bar{T}_y < 0$), a northward (southward) advection produces a warming (cooling). In the presence of a mean westerly $\bar{U} > 0$, a westerly (easterly) anomaly will cool (warm) the SST by enhancing (reducing) evaporation. In the last term, h is the perturbation thickness of the upper layer. This term parameterizes the local upwelling–downwelling effect. An upwelling $h < 0$ (downwelling $h > 0$) causes a cooling (warming).

The current in (2.3) consists of two parts: a geostrophic $v_g = \partial_x\psi$ and an Ekman drift $v_{\text{Ekman}} = -\tau_x/f$. Therefore, the temperature advection is caused by both the geostrophic and Ekman current. Furthermore, the thermocline depth is geostrophically related to the oceanic streamfunction by $h = f\psi/g'$. Thus, (2.3) can be rewritten as

$$(\partial_t + K_T)T + \partial_x\psi\bar{T}_y - \left(E\bar{U} - \frac{\gamma\bar{T}_y}{f}\right)\partial_y\Psi - \frac{fK_h}{g}\psi = 0. \quad (2.4)$$

3. Unstable coupled modes in the extratropics

Equations (2.1), (2.2), and (2.4) form our coupled system for unknowns Ψ, T, and ψ. The system is linear with constant coefficients (if $\bar{T}_y$ is assumed constant) and therefore can be solved easily. The coupled system generally has two eigenvalues that correspond to two types of air–sea coupling modes. One is the the SST mode (Neelin 1991) caused by the derivative in the SST equation (2.4). The other is due to the ocean dynamics in (2.2). In general, these two modes are coupled with each other. For a clear understanding of the physics of each mode, we simplify the equations further such that each mode can be studied separately.

a. Dynamic mode—The upwelling mode

Now, we consider the mode originating from the oceanic dynamics. Since our focus is the basic physics of the positive feedback, for clarity, we set $\partial_t T$, the advection term, and the evaporation term in the SST equation zero. Thus, (2.4) reduces to

$$K_T T = K_h h = fK_h\psi/g' \quad (3.1)$$

(Philander et al. 1984; Yamagata 1985; Hirst 1986). Equation (3.1) is perhaps proper in the southern part of subtropical gyres, where the mixed layer is shallow and therefore upwelling is more efficient in affecting SST. Furthermore, to illustrate the essence of the physics of the positive feedback, we neglect the Rossby wave (or the β effect) in the oceanic dynamics. Thus, (2.2) becomes

$$(\partial_t + K_m)\psi = -L_0^2\gamma\nabla^2\Psi. \quad (3.2)$$

This simply states that a cyclonic (anticyclonic) wind stress shallows (deepens) the thermocline due to local upwelling (downwelling). Substituting (2.1) and (3.1) into (3.2) results in the eigenvalue equation for SST as

$$(\partial_t + K_m)T = -\frac{\gamma h_0 K_h A}{fK_T}\nabla^2 T. \quad (3.3)$$

With a disturbance of the form $T \sim e^{i(kx+ly)+\lambda t}$, the eigenvalue is obtained from (3.3) as

$$\sigma \equiv \lambda + K_m = \frac{\gamma h_0 K_h K^2}{fK_T}A, \quad (3.4a)$$

where σ is the eigenvalue in the absence of dissipation; $K = \sqrt{k^2 + l^2}$ is the total wavenumber. The growth rate is simply

$$\sigma_r \equiv \lambda_r + K_m = \frac{\gamma h_0 K_h K^2}{fK_T}A_r. \quad (3.4b)$$

Thus, the positive feedback occurs when the atmosphere responds to a warm (cold) SST anomaly with a high (low) pressure, that is,

$$A_r > 0. \quad (3.5)$$

The physics of this instability is explained as in Fig. 1a. A warm SST causes a high surface pressure with anticyclonic wind stress, which forces downwelling and therefore warms the SST further. Since this positive feedback depends crucially on upwelling and downwelling, it will simply be called the upwelling mode. Obviously, if the atmospheric response is dominated by a low over a warm SST, one finds a decaying upwelling mode as shown in Fig. 1b. The upwelling mode can be inferred directly from the work of Philander et al. (1984).

To estimate the growth rate, we estimate K_h/K_T from (3.1). Assuming that a thermocline anomaly of 30 m produces a SST anomaly of 1°C, (3.1) gives $K_h/K_T = 1/30°\text{C m}^{-1}$. Thus, using other parameters in Table 1, with $K^2 \sim 1/L^2$, we obtain $\sigma_r \sim (\gamma h_0 K_h/fK_T L^2)A_r \sim 10^{-8}\,\text{s}^{-1} \sim 1/(2\text{ years})$. In the extratropics, the momentum damping in the thermocline (K_m) is usually several years. Hence, it is possible for an unstable upwelling mode to overcome dissipation (i.e., $\lambda_r > 0$).

b. SST modes—The Sverdrup mode and the evaporation mode

To separate the SST mode from the temporal ocean dynamics, we neglect the upwelling effect in (2.6). This is perhaps not a bad approximation except in the southern part of subtropical gyres. Because the mixed layer is deep, one might expect a weak impact from upwelling and downwelling. In particular, near the frontal regions such as the Gulf Stream and Kuroshio (after their separation from the coast), the strong meridional temperature gradient and mean evaporation enhances the effect of temperature advection and evaporation. Thus, (2.6) is reduced to

$$(\partial_t + K_T)T + \partial_x\psi\bar{T}_y - \left(E\bar{U} - \frac{\gamma\bar{T}_y}{f}\right)\partial_y\Psi = 0. \quad (3.6)$$

In addition, we assume that the oceanic planetary wave is much faster (the fast-wave limit of Neelin 1991) or much slower than the time scale of interest. This assumption is not intended for realistic climate variability, because the oceanic Rossby wave has the time scale from interannual to decadal. Nevertheless, it seems likely that the Rossby wave will significantly change the frequency rather than the growth rate of the coupling modes, at least in the weak coupling limit. Thus, (2.2) reduces to the Sverdrup balance (Pedlosky 1975)

$$\beta\partial_x\psi = \gamma\nabla^2\Psi. \quad (3.7)$$

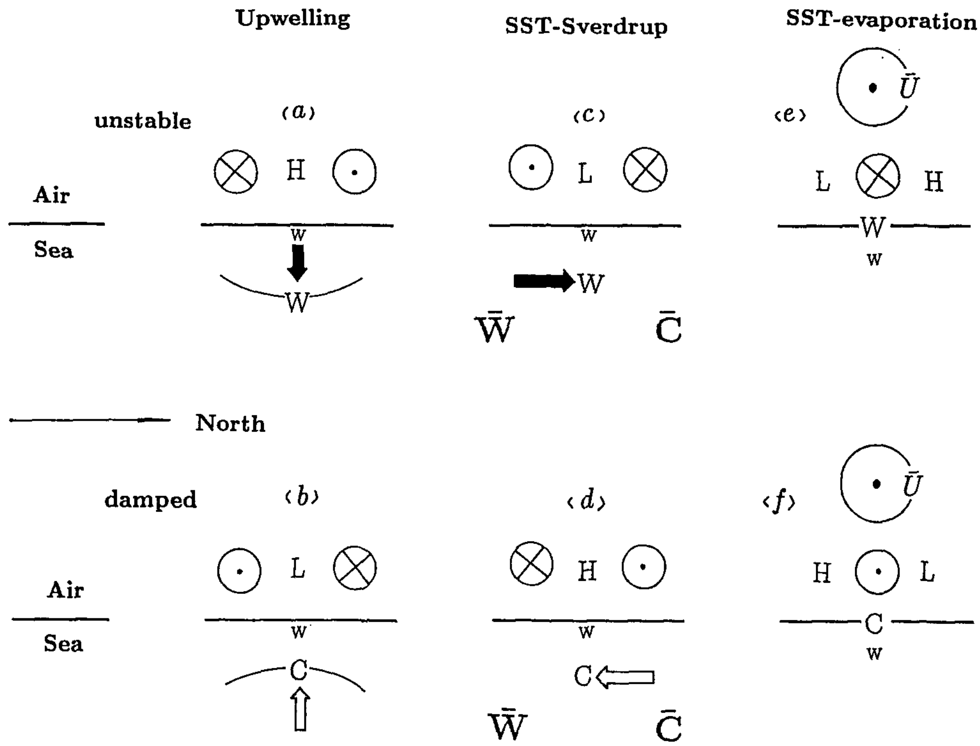

FIG. 1. This schematic figure shows meridional sections of (a) the unstable upwelling mode, (b) damped upwelling mode, (c) unstable SST–Sverdrup mode, (d) the damped SST–Sverdrup mode, (e) the unstable SST–evaporation mode, (f) the damped SST–evaporation mode. In all cases, the initial disturbances are a warm SST anomaly, denoted by a small W; the perturbation temperature due to coupling is denoted by midsized W. The arrows represent current forced by anomalous SST. The $\bar{W}$ and $\bar{C}$ in (c) and (d) represent the mean temperature; $\bar{U}$ is the mean wind. See the text for explanation.

Substituting (2.1) and (3.7) into (3.6), we have the eigenvalue equation for SST as

$$(\partial_t + K_T)T + \frac{\gamma\bar{T}_y}{\beta}A\nabla^2 T - \left(E\bar{U} - \frac{\gamma\bar{T}_y}{f}\right)A\partial_y T = 0. \tag{3.8}$$

The eigenvalue is then

$$\sigma \equiv \lambda + K_T = \frac{\gamma\bar{T}_y K^2}{\beta}A + il\left(E\bar{U} - \frac{\gamma\bar{T}_y}{f}\right)A. \tag{3.9a}$$

The growth rate is given by the real part as

$$\sigma_r \equiv \lambda_r + K_T = \sigma_S + \sigma_E, \quad \text{where} \quad \sigma_S \equiv \frac{\gamma\bar{T}_y K^2}{\beta}A_r,$$

$$\sigma_E \equiv -l\left(E\bar{U} - \frac{\gamma\bar{T}_y}{f}\right)A_i. \tag{3.9b}$$

Thus, the growth rate of SST mode has two parts: σ_S due to Sverdrup advection, and σ_E caused by evaporation and Ekman drift advection. Using parameters in Table 1, one can show that $(\gamma\bar{T}_y/f)/E\bar{U} \sim 10^{-1} \ll 1$. Thus, in σ_E, evaporation dominates Ekman drift.

In (3.9b), σ_S is similar to the upwelling mode in (3.4), but with the opposite sign. This is because the mean temperature decreases northward ($\bar{T}_y < 0$). Thus, a positive feedback occurs due to the Sverdrup advection if the atmosphere responds to a warm (cold) SST with a low (high) pressure, that is, $A_r < 0$, opposite to the unstable upwelling mode in (3.5). This unstable mode can be observed in Pedlosky's work (1975) if eddies are neglected. The physics of this positive feedback is explained as in Fig. 1c. An initial warm SST disturbance produces a low pressure with cyclonic wind stress, which forces a northward anomalous Sverdrup ocean current. Since the mean SST is warmer on the southern side, this northward current warms the SST anomaly further. By the same argument, one can show that the feedback is negative if the atmosphere responds to a warm (cold) SST with a high (low), as illustrated in Fig. 1d. Because this positive feedback depends crucially on the SST mode, the mean SST gradient, and the Sverdrup current, it will be called the SST–Sverdrup mode. Equation (3.9b) shows that the growth rate increases with the mean SST gradient, because an increased mean SST gradient enhances the temperature advection. This suggests that the potential regions for this mode might be those strong temperature frontal zones such as the Gulf Stream, the Kuroshio, and the Antarctic Circumpolar Current.

As shown in (3.9), the effect of evaporation differs dramatically from the SST–Sverdrup mode. For a positive feedback to occur, the atmosphere has to respond to the SST with a 90° phase difference. In the midlatitude where mean westerly prevails ($\bar{U} > 0$), the evaporation effect produces a positive feedback if the atmosphere responds to a warm (cold) SST with a low (high) a quarter-wavelength to the south, that is, $A_i < 0$. The physics is explained as in Fig. 1e. An easterly anomalous wind $U < 0$ is produced at the surface of the warm anomaly, which reduces the mean westerly and therefore reduces the evaporation cooling. This will intensify the warm anomaly further. The opposite case produces a damped mode as shown in Fig. 1f. This positive feedback has been proposed previously according to observations (e.g., Wallace et al. 1990).

The Ekman drift advection is similar to the evaporation effect in that positive feedback occurs only when the atmosphere responds to SST anomaly with 90° phase difference. More specifically, in the presence of a northward decreasing mean SST gradient $\bar{T}_y < 0$, the Ekman drift reinforces the evaporation feedback in the midlatitude region where mean westerly prevails, but opposes the evaporation in the polar region or the low latitude where mean easterly dominates. Due to the dominance of evaporation, the mode due to σ_E will be called the SST–evaporation mode.

Using parameters in Table 1, the growth rates of the SST–Sverdrup mode and SST–evaporation mode can be estimated from (3.9b) as $\sigma_S \sim (\gamma \bar{T}_y / \beta L^2) A_r \sim 10^{-7}\ \mathrm{s}^{-1} \sim 1/(4\ \text{months})$ and $\sigma_E \sim l E \bar{U} A_i \sim 10^{-7}\ \mathrm{s}^{-1} \sim 1/(4\ \text{months})$. These growth rates are faster than that of the upwelling mode. Therefore, although the thermal damping (K_T) is on the order of several months (larger than the momentum damping K_m), the two SST modes may be able to overcome the damping to grow. In comparison, the upwelling mode seems to favor the low-latitude region where the thermocline is shallow; the SST–Sverdrup mode favors the Gulf Stream or Kuroshio regions where the SST gradient is strong; and the SST–evaporation mode in general favors middle and high latitudes where the mean evaporation $\bar{Q}_E$ is large[1] (see Crayn 1990).

The upwelling mode and the SST–Sverdrup mode [see (3.5) and (3.9b)] have growth rates proportional to the square of the total wavenumber, due to the wind stress curl and the resultant Sverdrup current or upwelling–downwelling, while the SST–evaporation mode has a growth rate linearly proportional to the meridional wavenumber. These features seem to suggest that the most unstable wave is in the short-wave regime. This superficial conclusion is valid only when the atmospheric response is insensitive to total wavenumber, which is usually not true. Indeed, using a linear atmospheric model including relative vorticity and Ekman layer near the surface, one can show $A \sim 1/K^2$ for short waves ($K \to \infty$). Thus, at short-wave limit, the growth rate approaches a finite value for the upwelling mode and the SST–Sverdrup mode, and approaches zero for the SST–evaporation mode. Hence, the most unstable waves for the three modes are in the long-wave or planetary wave regime.

4. Summary and discussion

A simple theoretical analysis shows that positive feedbacks may exist in the extratropics. Furthermore, their time scales seem to favor interannual variability. Three positive feedbacks have been identified: the upwelling mode, the SST–Sverdrup mode, and the SST–evaporation mode. The upwelling mode becomes unstable when the atmosphere responds to a warm SST anomaly predominantly with a high surface pressure. In contrast, the SST–Sverdrup mode is destabilized when the atmosphere responds to a warm SST with a low pressure. In the region of mean westerly (easterly) wind, the SST–evaporation mode is unstable when the atmospheric response to a warm SST is a quarter-wavelength to the south (north). The upwelling mode seems to favor low-latitude regions, while the two SST modes seem to favor middle to high latitudes.

Furthermore, scaling analysis has shown that the Ekman flow plays a secondary role in positive feedbacks in the extratropics. On the other hand, the effect of Ekman flow is similar to the effect of evaporation. In some previous works (Bjerknes 1964; Namias 1965; Palmer and Sun 1985), the Ekman flow is frequently used to explain some correlated ocean–atmospheric variability. The analysis here seems to suggest that it is likely that the evaporation plays a more important role.

At the present stage, it seems premature to apply the simple model directly to observations and GCM simulations. The first difficulty comes from the atmosphere. We have seen that the positive feedback depends crucially on the response of the atmosphere to SST anomaly. However, observations and GCM simulations of extratropical atmospheric variability are not understood at all. The climatology or seasonal variation of the basic state suggest that a warm SST tends to produce a low pressure and vice versa. This can be well explained by most linear theories, which suggest that the lower-layer atmosphere should respond to a warm SST with a low of baroclinic vertical structure downstream (Hoskins and Karoly 1981; Webster 1981; Held

[1] It should be pointed out that the evaporation effect can be reinforced by the wind mixing effect on the mixed layer. A strong total wind also helps to stir the mixed layer and therefore cools the surface, reinforcing the evaporation cooling. However, in middle and high latitudes, the wind mixing effect on the mixed layer is not strong.

1983). This explanation even seems to agree with the experiments with an idealized GCM, in which the initial climate is taken as zonally uniform basic flow (Ting and Held 1990; Ting 1991). However, observations of the climate anomaly suggest almost the opposite—a cold SST anomaly is correlated with low surface pressure anomaly and a warm SST is correlated to high pressure (Namias 1972; Palmer and Sun 1985; Wallace and Jiang 1987; Wallace et al. 1990). In addition, the atmospheric response has a strong equivalent barotropic structure. GCM simulations with realistic mean climatology seem to produce most results of the observations. For example, a cold SST anomaly produces strong low surface pressure anomaly with equivalent barotropic structure. However, GCM experiments find that the atmospheric response is highly nonlinear. Responding to a warm SST anomaly, GCM results seem to still produce a low surface pressure anomaly (Pitcher et al. 1988; Kushnir and Lau 1992). Therefore, we have several different situations: the basic state and linear theories seem to suggest that the SST–Sverdrup mode is unstable[2]; observations seem to favor the unstable upwelling mode; GCM simulations appear to favor the unstable SST–Sverdrup mode during the warm SST half-period and favor unstable upwelling mode during the cold SST half-period. The intermediate phase requirement for the unstable SST–evaporation mode makes the analysis even more subtle. Indeed, past observations and numerical simulations can also be interpreted as favoring the unstable SST–evaporation mode.

Second, the ocean model is crude. In particular, the neglection of oceanic planetary waves is a bad assumption for time scales of our interest, because these waves have natural periods from years to decades. Therefore, inclusion of oceanic planetary waves will change the frequencies of the coupling modes substantially. In addition, these waves may also change the spatial phase of the favorable atmospheric response for a particular positive feedback. Moreover, the coupling between the SST and upwelling mode will be of great interest. Both modes have similar time scales and therefore might be expected to have a strong coupling. These results are in progress. Clearly, much effort is needed on both the atmospheric and oceanic sides.

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[2] A two-layer atmospheric model linearized on mean zonal winds has been used to replace (2.1a). The results show that over most of the parameter region, the dominant unstable mode is the SST–Sverdrup mode. Near the resonant wavenumber region, however, both the upwelling mode and the SST–evaporation mode can be unstable.

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