Vacillations in a Coupled Ocean–Atmosphere Model

PAUL S. SCHOPF

Laboratory for Oceans, NASA Goddard Space Flight Center, Greenbelt, Maryland

MAX J. SUAREZ

Laboratory for Atmospheres, NASA Goddard Space Flight Center, Greenbelt, Maryland

(Manuscript received 24 March 1987, in final form 24 August 1987)

ABSTRACT

Results are presented from a 35-year integration of a coupled ocean–atmosphere model. Both ocean and atmosphere are two-level, nonlinear primitive equations models. The global atmospheric model is forced by a steady, zonally symmetric Newtonian heating. The ocean model is solved in a rectangular tropical basin. Heat fluxes between ocean and atmosphere are linear in air–sea temperature differences, and the interfacial stress is proportional to lower-level atmospheric winds.

The coupled models produce ENSO-like variability on time scales of 3 to 5 years. Since there is no external time-dependent forcing, these are self-sustained vacillations of the nonlinear system. It is argued that the energetics of the vacillations is that of unstable coupled modes and that the time scale is crucially dependent on the effects of ocean waves propagating in a closed basin.

1. Introduction

As a result of much recent interest in El Niño and the atmosphere’s Southern Oscillation, a large literature has developed on interactions between the tropical ocean and atmosphere. Attempts at explaining the large variance in tropical SST at periods from 2 to 5 years quickly discovered the need to invoke the coupling between ocean and atmosphere as a key feature in any theory for ENSO. It became apparent that the surface heat balance controlling the temperature of a tropical oceanic mixed layer implies very short damping time for SST anomalies. While ENSO has a time scale of 1000–2000 days, reasonable estimates of the mixed-layer depth and the sensitivity of surface heat flux to SST indicate that tropical SST anomalies would decay in ≲100 days if no positive feedback processes existed. Studies with atmospheric GCMs added the result that the source of these long-range variations cannot be ascribed simply to natural atmospheric variability: when the GCMs are run with steady SSTs, they produce far too little tropical variance at these low frequencies (Manabe and Hahn, 1981). When the SSTs vary in time, but in a prescribed fashion, the atmosphere shows more variability, but it is limited to the region of the spectrum where the SST has power. In short, there must be some strong feedback process that very nearly balances the damping associated with surface heat fluxes.

One possibility, explored by Schopf (1983), is that our estimates of the surface heating sensitivity to SST are in error because the atmospheric temperature can adjust rapidly to the SST. Schopf (1985) estimates that on the length scales of interest for El Niño this effect can double the decay times. While this is significant, the basic problem remains.

The resolution seems to be in the dynamical responses of the atmosphere and ocean to each other. Lau (1981), Philander et al. (1984), Yamagata (1985) and Hirst (1985) identified unstable coupled modes in linearized versions of the tropical ocean–atmosphere system. These theories draw on the basic coupling processes identified by Bjerknes (1966): warm SST anomalies along the equator induce westerly wind anomalies which drive currents that help maintain or enhance the warm water. The growth rates found by Philander et al. were in the range of 30–60 days—a convenient time scale for resolving the difficulty presented above. While these instabilities are interesting and fundamental to the development of an ENSO theory, they do not provide a ready explanation for the periodicity or quasi-periodicity of ENSO.

Two means of obtaining low-frequency variability were proposed in McWilliams and Gent (1978) and Vallis (1986). Based on very similar simple schematic models, but using different parameter values, the two studies reach different conclusions. McWilliams and Gent considered a system which could have become unstable, but argued that realistic parameter choices lead only to damped but oscillatory solutions. They then produce “low-frequency” variability in the least-damped modes by adding a random forcing.
If this were the proper paradigm for ENSO we would expect its spectrum to show the frequencies of the selected modes. Vallis' case is qualitatively different. His choice of parameters was such that the model produced spontaneous chaotic variability. In the model we present here the variability also arises spontaneously when dynamic coupling is included, but unlike Vallis', it selects a time scale.

Regular oscillations similar to ours have been found in other dynamical models. McCreary (1983) and McCreary and Anderson (1984) considered a model with ocean dynamics and a crude parameterization of surface wind response to SST changes. This system had sufficient feedback strength to bifurcate and possess multiple equilibria through abrupt transitions prescribed in the atmospheric response. A question raised in these calculations was that of explaining whether the system can oscillate between these states through some internal or external process. In the first, "ringing" and "overshooting" of the system following a transition from one state to the other led to a transition back to the original state. The second study included a mock "annual cycle" which played a crucial role.

Anderson and McCreary (1985) subsequently examined a system in which the atmosphere was modeled as in Gill (1980). The ocean dynamics was treated through a single-layer reduced gravity approach, and SST was predicted nonlinearly through horizontal advection, surface heating, a background entrainment and upwelling. The surface heat flux was linear in SST for positive anomalies, but insensitive to negative SST changes below a certain threshold. The role of these two nonlinear effects is not clear in their results. But the model produces a regular oscillation that appears unrelated to the mechanisms they had proposed previously. The authors do not identify the mechanism responsible for the frequency selection.

Zebiak (1984) and Cane and Zebiak (1985) have developed a coupled ocean-atmosphere model in which the ocean also evolves through linear reduced gravity equations for a single layer. The atmosphere is that of Gill (1980) and the relationship between SST and heating of the atmosphere allows for a temperature dependence of the coupling coefficient. As in Anderson and McCreary (1985), all the dynamics is linearized about a state of rest. The coupling, however, is dependent on oceanic temperatures predicted by a fully nonlinear thermodynamic equation. Ocean temperatures are thus affected by both climatological currents and linearly predicted perturbations. Through the prescribed climatology, they include effects of the seasonal cycle. Depending on the choice of parameters, their model produced a variety of low-frequency behavior (Zebiak, 1984), including long-period ENSO-like oscillations.

Here, we present calculations using nonlinear primitive equations models for both ocean and atmosphere, but coupling them through a simple linear relationship between SST and heating. Though greatly idealized, they are general circulation models which evolve their own mean states and produce their own natural high-frequency variability. We wish to see if low-frequency variability can arise internally when these models are coupled. To do this, we suppress all temporal variations in the forcing function (including the annual cycle). It is not our aim to simulate the ENSO phenomenon, but rather to see if it is possible to construct a simple dynamically consistent system. This captures its essence, and which can be more readily understood. The addition of more detail and refinement will come later, as this type of calculation is performed with more realistic GCMs.

A detailed description of the models and a discussion of the major simplifications made is given in section 2. In section 3 we present the results of a 35-year integration. These show an irregular oscillation with SST anomalies reminiscent of those during El Niño events, and with most of the variance in the 36-54 month range. In section 4 we examine the behavior of the system when the atmospheric model is linearized, thereby removing variability generated by the atmosphere alone. Section 5 discusses the destabilization process which leads to the bifurcation from a stationary state and the selection of a preferred time scale.

2. The model

We felt a minimum model of ENSO would need to include only a tropical ocean in a simple basin, but one capable of producing a reasonable climatology of surface temperature, thermocline depth and upper-ocean currents when forced with realistic winds. We also wanted the coupled model to produce a consistent ocean-atmosphere climatology. So rather than driving the ocean with climatological winds and using an "anomaly" model for the atmosphere, we chose to predict the global atmospheric circulation. This way we could study the forcing of the system is the differential pole-to-equator heating of the atmosphere. A schematic of the model geography is shown in Fig. 1 (the continents are drawn only to give a sense of scale). The ocean basin occupies 160° of longitude and spans 40° of latitude. Figure 2 shows the vertical structure used in the models.

a. The atmosphere

The atmospheric model is based on the model described in Held and Suarez (1978). It is a two-level primitive equations model on a sphere. Winds and temperature are carried at both levels, but the vertically averaged flow is taken to be nondivergent. Water vapor is not included in the version used in this paper. We drive the general circulation by relaxing the mean atmospheric temperature toward a zonally symmetric state with a large pole-to-equator temperature difference. The only other thermal forcing of the atmosphere is heat exchange with the ocean, which we assume to
depend only on the air–sea temperature difference, and add directly to the atmosphere’s vertically averaged thermodynamic equations.

For computational efficiency, Held and Suarez used an unconventional discretization in the horizontal. The equations were differenced in the meridional direction and spectrally decomposed in the zonal direction, the intention being to use severe truncation in describing zonally asymmetric motions. Since this is not acceptable for our purposes, the mixed discretization was abandoned, and the equations were simply finite-differenced on a latitude–longitude grid. For completeness, the two-level equations are reproduced here.

For any variable $A$, let

$$
\dot{A} = \frac{1}{2} (A_1 + A_2)
$$

with subscripts 1 and 2 denoting values at the upper and lower levels. Also let $P = (p/p_0)^\kappa$, where $p$ is the pressure, $p_0 = 1000$ mb and $\kappa = R/c_p$. The $R$ is the gas constant and $c_p$ the heat capacity of air at constant pressure. The two levels are placed at $p_1 = 250$ mb and $p_2 = 750$ mb. With these definitions, the finite-difference hydrostatic equation is written

$$
\Phi = -\beta \hat{P} c_p
$$

where $\Phi$ is the geopotential and $\beta$ is the potential temperature.

As in the Held-Suarez model, the vertically integrated divergence is set to zero. Thus, the continuity equation is

$$
\nabla \cdot \vec{v} = 0
$$

(4)

$$
\nabla \cdot \vec{\theta} = -\omega
$$

(5)

where $\omega$ is the $p$-velocity divided by $2\beta$, and $\vec{v}$ is the horizontal velocity. The prognostic equations for the zonal and meridional velocity components, $u$ and $v$, are:

$$
\frac{\partial \hat{u}}{\partial t} = \left( f \hat{v} + \hat{v} \right) - \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} (\hat{\Phi} + \hat{K}) + \hat{F}_u
$$

(6)

$$
\frac{\partial \hat{v}}{\partial t} = \left( f \hat{u} + \hat{u} \right) - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\hat{\Phi} + \hat{K}) + \hat{F}_v
$$

(7)

$$
\frac{\partial \bar{u}}{\partial t} = \left( f \bar{v} + \bar{v} \right) - \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} (\bar{\Phi} + \bar{K}) + \bar{F}_u
$$

(8)

$$
\frac{\partial \bar{v}}{\partial t} = \left( f \bar{u} + \bar{u} \right) - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\bar{\Phi} + \bar{K}) + \bar{F}_v.
$$

(9)

In these, $f = 2\Omega \sin \phi$ is the Coriolis parameter, $a$ the radius of the sphere, $\phi$ and $\lambda$ the latitude and longitude,

$$
\hat{\xi} = \frac{1}{a \cos \phi} \left\{ \frac{\partial v}{\partial \lambda} \frac{\partial}{\partial \phi} (u \cos \phi) \right\}
$$

(10)

is the relative vorticity, and

$$
K = \frac{1}{2} (u^2 + v^2).
$$

(11)

The $F_u$ and $F_v$ represent frictional effects.

Because we assume $\nabla \cdot \vec{v} = 0$, the vertically averaged flow is governed by the vorticity equation alone,

$$
\frac{\partial \bar{\xi}}{\partial t} = -\frac{2\Omega}{a} \cos \phi \bar{v} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left[ \bar{\xi} u + \omega \hat{v} + \bar{F}_v \right] - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left[ \cos \phi (\bar{v} + \omega \hat{u} + \bar{F}_u) \right].
$$

(12)

Finally, the thermodynamic equations are

$$
\frac{\partial \vec{\theta}}{\partial t} = -\nabla \cdot (\bar{\theta} \vec{v}) + \bar{Q}
$$

(13)
\[
\frac{\partial \bar{\theta}}{\partial t} = -\nabla \cdot (\bar{v} \bar{\theta}) - \omega \bar{\theta} + \bar{Q}
\]

where \(\bar{Q}\) represents the diabatic heating.

As discussed, we use the following simple form for the heating:

\[
\bar{Q} = -K \delta (\bar{\theta} - \bar{\theta}(\phi)) + gH/p_0 \bar{P}_c
\]

(15)

\[
\bar{Q} = -K \delta (\bar{\theta} - \bar{\theta}_e)
\]

(16)

where \(1/K_\delta = 30\) days and the equilibrium temperature distribution in Kelvins is taken as

\[
\bar{\theta}_e(\phi) = 302 + 40 \cos(2\phi),
\]

(17)

\[
\bar{\theta}_e = 15.
\]

(18)

We emphasize that \(\bar{\theta}_e\) is independent of time and that this steady relaxation of the atmosphere to a prescribed radiative-convective equilibrium is the only "external" forcing of the coupled system. The second term on the right-hand side of (15) is the midtropospheric heating due to heat exchange with the ocean. The form of the "vertical heat flux", \(H\), and the justification for including this effect only in (15) are given below.

The friction terms in (4)–(7) include both surface and lateral friction:

\[
F_1 = D_1
\]

(19)

\[
F_2 = D_2 + 2g\tau/p_0
\]

(20)

where \(\tau\) is the surface stress. Lateral friction \((D_1)\) results from an eighth-order Shapiro filter applied to the two velocity components every two hours.

b. The ocean

The ocean model used is that developed in Schopf and Cane (1983), as modified by Schopf and Harrison (1983). It solves the primitive equations, including thermodynamic budgets, for two layers meant to represent the oceanic mixed layer and the upper thermocline. These overlie an abyssal layer of fixed temperature to which they are connected by diffusion. This abyssal layer is at rest and supports no horizontal pressure gradients. The model thus predicts sea-surface temperature, thermocline displacement, sea-level and currents. The salient climatological features of the tropical Pacific circulation—the east–west surface temperature gradients, the sloping thermocline, the surface easterly flow and the equatorial undercurrent—can be captured in such a model.

Entrainment at the base of the mixed layer is parameterized using a bulk turbulent kinetic-energy budget similar to that proposed by Kraus and Turner (1967). Except for the vertical diffusion of heat, the base of the lower layer is intended to represent a material surface with a mean depth of approximately 150 m. In certain cases, however, this condition must be relaxed. The occurrence of time-mean Ekman pumping in the solutions can produce locally strong shoaling of the lower interface. When this occurs, we take it to represent the appearance of abyssal water at the surface and try to model its effect on the SST. To do this, we allow a vertical mass flux across the interface when it reaches a minimum depth (equivalent to transforming locally from the material coordinate to a z-coordinate.) The water thus added to the upper layers is assumed to be at the abyssal temperature and at rest.

If this were the only departure made from the material coordinate, the upper layers would accumulate mass during the run until they became too deep to reasonably resolve the upper-ocean structure. To balance this effect, mass is returned to the abyssal layer (carrying the temperature and momentum of the second layer) whenever the interface becomes deeper than a prescribed depth. With this scheme the model has some freedom in determining its own mean thermocline depth, but can be maintained at a convenient depth for representing the upper ocean. This results in vertical advection of abyssal water into the active layers, particularly along the equator in the east.

Letting subscripts 1 and 2 denote the upper (mixed) and lower layers, the continuity, thermodynamic and momentum equations are

\[
\frac{\partial h_1}{\partial t} = -\nabla \cdot (h_1 V_1) + w_e
\]

(21)

\[
\frac{\partial h_2}{\partial t} = -\nabla \cdot (h_2 V_2) + w_d - w_e
\]

(22)

\[
\frac{\partial T_1}{\partial t} = -V_1 \cdot \nabla T_1 + \{w_2(T_2 - T_{2,t})
\]

(23)

\[
+ A(T_2 - T_{1}) - Q_{oa} - H/p_0 c_0)/h_1
\]

\[
\frac{\partial T_2}{\partial t} = -V_2 \cdot \nabla T_2 + \{w_d(T_d - T_2) - w_1(T_e - T_{2,t})
\]

(24)

\[
+ A(T_d - 2T_{2,t} + T_1) + Q_{oa})/h_2
\]

\[
\frac{\partial V_1}{\partial t} = -V_1 \cdot \nabla V_1 + f k \times V_1 + w_1(V_e - V_1)/h_1
\]

(25)

\[
- (\nabla \pi_1 - b_1 \nabla z_1) + \tau/\rho_0 h_1
\]

\[
\frac{\partial V_2}{\partial t} = -V_2 \cdot \nabla V_2 + f k \times V_2 + [w_d(V_d - V_2) - w_2(V_e - V_2)]/h_2
\]

(26)

\[
- (\nabla \pi_2 - b_2 \nabla z_2) - \nabla h_2 - (\nabla h_2/2).
\]

In these, \(h\) is the layer thickness, \(T\) is the temperature, and \(V\) is the horizontal velocity. A subscript \(e\) denotes quantities associated with entrainment at the base of the mixed layer, and a subscript \(d\), those associated with mass exchange across the bottom of the second layer when the material surface condition is relaxed. The \(T_d\) and \(V_d\) are computed in an upstream sense, based on an abyssal temperature of 14.5°C and zero velocity. The \(H\) and \(\tau\) are the fluxes of heat and momentum at the ocean surface. These also appear in the atmospheric equations (15) and (20). The \(A\) is a constant mixing coefficient, and \(k\) is the vertical unit vector. The \(Q_{oa}\) represents heat exchange by convective adjustment of the layers. The pressure \((\pi)\) and depth \((z)\) of the layers are

\[
\pi_1 = \frac{1}{2} b_1 h_1 + b_2 h_2
\]

(27)
\[ \pi_2 = \frac{1}{2} b_2 h_2 \]  
\[ z_1 = -\frac{1}{2} h_1 \]  
\[ z_2 = -\left( \frac{1}{2} h_1 + \frac{1}{2} h_2 \right) \]  

Equation (26) includes an ad hoc correction to the pressure gradient in the second layer (the \( \Gamma \) terms), which depends on the layer buoyancy and thickness. For a discussion of the terms involving \( \Gamma \), the reader is referred to Schof and Harrison (1983). The buoyancy, \( b \), is given by

\[ b = g\alpha(T - T_r) \]  

where \( \alpha \) is a constant coefficient of expansion and \( T_r \) a reference temperature. \( \rho_0 \) and \( c_0 \) are the density and heat capacity of sea water.

Details of the treatment of the mixed layer are given in Schof and Cane (1983). Briefly, the scheme is based on the integrated turbulence kinetic energy budget:

\[ w_e h_1 (b_1 - b_e) = S - 4h_1 + h_1 g\alpha H / \rho_0 c_0. \]  

The term on the left represents the work done by turbulent motions against stratification at the base of the mixed layer. The \( w_e \) is entrainment velocity, \( S \) is the surface production by wind mixing (assumed constant) and \( \epsilon \) is the turbulence dissipation per unit depth, also assumed constant. The last term is the buoyancy production due to surface heat flux. As discussed in Schof and Harrison (1983), \( (b_1 - b_e) \) is required to be positive. When the right-hand side of (32) is positive, \( w_e (>0) \) is computed directly. When it is negative, (32) would result in "detrainment". In this case \( h_1 \) is adjusted to make the right-hand side zero (the Monin-Obukhov equilibrium depth). In the calculations this adjustment is made in a single time step, and an effective \( w_e (<0) \) is obtained from the rate of change of \( h_1 \).

Equations (21)–(26) are solved in a rectangular basin using Cartesian coordinates \( x = a\lambda \) and \( y = a\phi \). The basin has rigid walls at which we apply no-slip boundary conditions. It extends over \(-20^\circ < \phi < 20^\circ\) and \(160^\circ\) of longitude. Values of the parameters used are given in Table 1.

**Table 1. Coefficient values used in the model.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( A )</td>
<td>( 10^{-4} \text{ m s}^{-1} )</td>
</tr>
<tr>
<td>( \alpha )</td>
<td>( 2 \times 10^{-4} \text{ C}^{-1} )</td>
</tr>
<tr>
<td>( T_r )</td>
<td>( 10.5^\circ \text{C} )</td>
</tr>
<tr>
<td>( S )</td>
<td>( 1.25 \times 10^{-6} \text{ m}^2 \text{ s}^{-3} )</td>
</tr>
<tr>
<td>( \epsilon )</td>
<td>( 7.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-3} )</td>
</tr>
<tr>
<td>( \min(h_1) )</td>
<td>20 m</td>
</tr>
<tr>
<td>( \max(h_1) )</td>
<td>100 m</td>
</tr>
<tr>
<td>( \min(h_2) )</td>
<td>50 m</td>
</tr>
<tr>
<td>( \max(h_2) )</td>
<td>100 m</td>
</tr>
<tr>
<td>( \rho_0 )</td>
<td>( 10^3 \text{ kg m}^{-3} )</td>
</tr>
<tr>
<td>( c_0 )</td>
<td>( 4 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1} )</td>
</tr>
</tbody>
</table>

**c. The coupling**

The vertical heat flux between the ocean and atmosphere, \( H \), is assumed to depend only on the air–sea temperature difference:

\[ H = -\gamma(T_s - T_1) \]  

where \( T_1 \) is the ocean mixed-layer temperature, \( T_s \) is the atmospheric temperature at the surface, obtained by extrapolating \( T \) linearly in \( \log(p) \),

\[ T_s = 0.986 \bar{\theta} - 1.337 \bar{\theta} \partial (\text{K}) \]  

and \( \gamma = 40 \text{ W m}^{-2} \text{ K}^{-1} \). The \( H \) is used in Eqs. (15), (23) and (32).

The mechanical coupling is also very simple. The surface stress is assumed to be proportional to the atmosphere’s lower-level velocity:

\[ \tau = \rho_0 K_e v_2 / 2g, \]  

with \( (1/K_e) = 5 \text{ days} \), which is roughly equivalent to using a 6.5 m s\(^{-1}\) wind speed with a drag coefficient of \( 1.4 \times 10^{-3} \). This stress is applied to both the ocean and atmosphere, as shown (20) and (25). Because the ocean grid is much finer than the atmospheric grid, we linearly interpolate the surface air temperature and wind stress to the ocean grid. Heat fluxes are then computed on the ocean grid, and averaged horizontally before being passed to the atmosphere, thereby ensuring that the total heat leaving the ocean enters the atmosphere.

The choice for the mechanical coupling (35) is straightforward, but the choice of thermal coupling is a bit more subtle and deserves further comment. Both observational and modeling studies have shown that the anomalous atmospheric thermal forcing during El Niño events is due to shifts in the regions of strong convective precipitation; that is, to changes in the horizontal transport of water vapor in the lower troposphere, rather than to direct heating by anomalous vertical heat fluxes at the surface. At the same time most of the anomalous heat loss by the ocean is in latent heat of evaporation. So although the main damping of SST anomalies (evaporation) is closely linked to the main anomalous forcing of the atmosphere (precipitation), the relationship is far from local.

Clearly, with a “dry” atmosphere model we could not simulate this important aspect of the coupled problem. We thus had to be satisfied with the local coupling we have described. This idealization, further, makes it difficult to choose a value for the coupling parameter. Since most of the anomalous precipitation is due to water vapor transport, peak anomalous atmospheric heating rates must be several times larger than peak evaporation rates. If we think of \( H \) as the evaporation [as it effectively appears in (23)] we choose a much lower value for \( \gamma \) than if we think of it as the atmospheric heating from deep convective precipitation [as it appears in (15)]. The choice made is intended to be a compromise between these two. For the ocean
this value corresponds to a damping time \((=\rho_0 h_1 c_0/\gamma)\) of \(\sim 60\) days, while for the atmosphere, it corresponds to relaxing \(\bar{\theta}\) to \((T_i + 1.337\bar{\theta})\) with a time scale \((=\rho_0 F_\rho / g\gamma)\) of \(\sim 3\) days.

3. Results

The coupled models were started from rest and integrated for 35 years. In about 100 days the atmosphere adjusted to the differential pole-to-equator heating, producing a general circulation with lower-level easterlies in the tropics. These spun up the ocean's circulation. We allowed 5 years for the system to equilibrate, and analyzed the last 30 years of the run.

a. Climatology

The three panels in Fig. 3 show 30-year averages of sea surface temperature, dynamic height of the ocean, and zonal surface wind over the basin. The mixed layer is cooler than the surface air over most of the basin. This implies a net heat flux into the ocean that is balanced by mixing with abyssal water. There is an SST minimum along the equator, with a “cold tongue” found on the equator in the east. In this region the mixed layer is shallow. In the central and western portion of the basin SST gradients are small and the mixed layer is quite deep. The interface with the abyssal layer has a zonal tilt along the equator characteristic of the thermocline in the Pacific, with 15°C water very near the surface in the east. However, cold surface waters do not appear along the eastern boundary.

This is consistent with the surface wind climatology. Easterlies are found over much of the equator, but at the eastern end there is a small region of westerly stresses. This westerly stress is not seen in the Pacific, and accounts for the unrealistically warm, deep mixed layer the model produces near the eastern boundary. The air–sea heat flux \(H\) is parameterized as in (33) over water only, and is from the atmosphere to the ocean over most of the basin. Over the land, \(H = 0\). Viewed in terms of a departure from zonal mean surface heating, the land areas appear as surface heat sources for the atmosphere. Westerly winds just to the west of the continent, as appear here, are consistent with the linear atmospheric response to such heating (Gill, 1980).

b. Low-frequency behavior

As shown in Figs. 4–10, the model has produced a large amount of internal variability. Figure 4a shows the time–longitude behavior of monthly mean SST anomalies along the equator. The largest anomalies are \(\sim 2^\circ\)C, somewhat smaller than those of observed El Niño events, but well organized and with multiyear time scales.

The strongest anomalies are confined to that region in the eastern half of the basin where the thermocline is shallow and mixed-layer temperatures are low, (i.e.,

where the climatological state has large horizontal and vertical temperature gradients). There is practically no variability in the SST in the western equatorial region, or, as we will see below, anywhere off the equator.

Most of the variance is in the 3 to 5 year range. Although the model appears to have strongly selected these time scales, the timing and intensity of events are irregular, with a strong hint of episodic behavior. The 15 years shown have 3 cold and 3 warm “events”, each of about 1 year duration, separated by quiet periods. It is also interesting to note that there appears to be almost no asymmetry between “cold” and “warm” events.

Figure 4b shows the oceanic surface height field along the equator. Since the pressure gradient below the
thermocline is assumed to vanish, the surface height is equal to the dynamic height relative to the abyss:

\[ \eta = (b_1 h_1 + b_2 h_2)/g. \]  

(36)

This is effectively the heat content of the upper ocean. The models of Lau (1981), Philander et al., (1984) and Yamagata (1985) treat SST as proportional to the depth of the thermocline (or surface height). In our model, the three quantities are independent, and while the large SST anomalies in the east are strongly correlated with surface height perturbations, in the western half of the basin large surface height anomalies occur with little or no associated SST anomalies. In fact, SST and surface height are out of phase in the west.

Figure 4c gives the equatorial time-longitude behavior of the zonal surface wind field, which can be seen to exhibit a strong temporal correlation with the SST anomalies. Westerly surface winds are associated with warm SSTs and easterlies with cool. Maximum wind anomalies occur \( \sim 30^\circ \) west of the maximum SST anomalies. The strongest SST anomalies are around 130\(^\circ\)W, the maximum wind anomalies are close to the center of the basin, around 150\(^\circ\)-170\(^\circ\)W.

Although not shown here, the surface air temperature shows very little change, what changes exist being very uniform zonally. This is because the heat flux from the ocean surface is put into the midtroposphere as discussed in section 2, affecting the vertical mean temperature but not the static stability. As a result, the surface heating anomalies will follow the SST anomalies very closely, and the correlation between SST and wind and the zonal separation of the two are equivalent to correlation and separation of wind and heating anomalies.

It is these wind anomalies in the central/western portion of the basin that drive the equatorial surface height anomalies seen in Fig. 4b, with westerly winds in the center of the basin associated with positive height anomalies in the eastern half. Small height anomalies also appear to emerge from the western boundary substantially before the SST warms in the east and the wind anomalies arise, implying the existence of a signal reflected from the western boundary. As discussed in Schopf and Harrison (1983), wind anomalies drive changes in the model’s surface height through the gravest baroclinic mode. While details of entrainment and surface heating will affect the SST balance, the surface height anomalies behave very much as would be predicted by linear shallow water equations, and we consider the surface height as a good indicator of the ocean’s dynamic response to the wind field.

Figure 5a shows the time-longitude behavior of the surface height averaged between 5\(^\circ\) and 7\(^\circ\) latitude. In contrast to the equatorial values, the height signals in Fig. 5a have westward phase propagation. From linear theory we expect localized equatorial wind anomalies to drive height perturbations that propagate eastward and westward from the forcing region. Those propagating eastward will have peak amplitude on the equa-
Figure 5. Time–longitude history of model anomalies averaged between 5° and 7° of both hemispheres for the first 15 years of the run. (a) Ocean surface height ($\eta_r$). Contour interval = 1 cm. (b) SST. Contour interval = 0.25°C. Negative anomalies are hachured.

Figure 6c shows the correlation between $\eta_r$ and surface wind anomalies in the center of the basin. Here the strongest correlations occur after the event, west of the wind maximum.

In Fig. 7 we show the horizontal structure of the variations through maps of monthly mean SST and surface wind anomalies for a warm period (month 171 of the run) and a cold period (month 81). By presenting a map for a short (monthly) mean, we show not only the low-frequency, organized relationships between the SST and wind, but also certain effects of “weather” which can intrude on the system from other sources. Overall, however, the relationship can be viewed as predominantly westerly wind anomalies associated with warm equatorial SSTs, winds which are largely trapped near the equator. Easterlies are associated with cold events.

Figure 8 shows time–longitude plots of the principal terms in the SST budget (23) averaged from 2°N to 2°S. The low-frequency response appears to be a balance between vertical advective cooling and surface heating. The contribution from zonal advection is secondary, and generally has the same sense as the vertical term: to the west of the maximum temperature anomaly zonal advection causes warming at the same time that vertical motion anomaly is downward, consistent with the kinematics of a convergent Kelvin wave in a negative temperature gradient. The zonal advection changes sign, however, to the east of the maximum anomaly, primarily because of the change in sign of the temperature gradient, a change which exists in both the mean field and the anomaly fields. The zonal advective term, discussed in Harrison and Schoepf (1984) as a dominant term in the 1982/3 El Niño, is not as strong here. Finally we note that the tendency term is small so that the ocean model’s equatorial thermodynamics is a quasi-static balance between the “fast” terms: upwelling and thermal damping.

In Fig. 9 we present a series of lagged covariance maps of the surface height and wind fields. The quantity plotted is

$$\sigma = \frac{\langle \tau(x_0, y_0, t) \eta(x, y, t+\delta t) \rangle}{\langle \tau^2(x_0, y_0, t) \rangle^{1/2}},$$

chosen to give a sense of the relative amplitudes of the surface height field in space and time. As the center of action for the wind appears to be the center of the basin, we have computed the covariance of $\eta$ at all grid points with the zonal equatorial wind at $x_0 = 160^\circ W$. The features seen in Fig. 9 are generally consistent with the view presented above.

In the east the dominant signal appears to be a strong positive covariance trapped on the equator, with height perturbations nearly coincident with the wind changes. We have argued that height changes in this region are
strongly related to SST anomalies and consequent thermal forcing of the atmosphere which drives wind changes almost simultaneously. This meridional structure is similar to that of equatorial Kelvin waves. When we examined the correlations (rather than the covariances shown) between the wind and height, it was apparent that Rossby wave energy is reflected from the eastern boundary, with its maximum amplitude off the equator, but the data seen in Fig. 9 indicate that the amplitude of such reflected waves is small in comparison with the activity on the equator.

To the west there is a quite different situation. We
see most of the activity off the equator, with maxima around 5°–7° latitude. The covariances are negative, consistent with the changes that would be induced in the ocean through Rossby wave dynamics.

These results can be summarized as follows: Large SST anomalies are confined to the eastern portion of the basin (see Figs. 4a, b and 5a, b), where they can be maintained by anomalous upwelling (Fig. 8). It is only in this region then, that the model's ocean and atmosphere can be strongly coupled through the joint actions of winds driving ocean circulation changes which alter the SST which modifies the heating which further modifies the winds. In the results all the components in this cycle are of the right sense to produce positive feedback (Fig. 6a and 6b). Precursory, small-amplitude equatorial height signals can be traced from the region of maximum SST anomalies all the way back to the western boundary (Fig. 6b). These height signals have only weak SST anomalies associated with them, until they reach the active region in the East. We hypothesize that these precursory equatorial height signals can be traced further back to westward-propagating height anomalies found at higher latitudes and that their origin is at the wind anomalies in the central portion of the basin, and is associated with the previous phase of the cycle.

Figure 10 gives a synthesis of this cycle in a complicated time–longitude display. At the left (10a) we show the surface height anomalies averaged from 5 to 7° ($\eta_R$), but with 80°W on the left and 120°E on the right. Westward propagation is then a tilting of phase toward the upper right. In this panel we see the origin of substantial height perturbations in the middle of the basin, not a propagation from the eastern boundary. Adjacent to this (10b) we show the equatorial height field ($\eta_R$) between 120°E and 140°W with west to the left. Fortunately, the difference in meridional structure of the westward-propagating and eastward-propagating signals allows us to see clearly the reflection at the western boundary by the continuity of phase propagation to the upper right. The equatorial signals are relatively weak in the west, but much stronger in the east. The next panel over (10c) shows the zonal wind anomalies between 180° and 140°W, which are highly correlated with the surface height.

In Fig. 10d we have plotted $\eta_R$ again, but only from the center of the wind anomalies out to the western boundary. As before, east is to the left, but here the stippling is reversed. This shows the strong relationship between positive wind anomalies and the generation of negative $\eta_R$ anomalies. Finally, a repetition of Fig. 10b, but with reversed stippling completes the cycle.

Throughout the series of figures, propagation of phase from lower left to upper right indicates a loop of westward propagation (visible off the equator), a phase-preserving reflection at the western boundary into eastward propagation (enhanced along the equator) and an interesting phase-reversing reflection in the center of the basin through the action of the coupling process.

It is this phase-reversing reflection (what we shall term a “coupled reflection”) which allows the system
to undergo this cycle. The coupled reflection operates through the instability process: when eastward-propagating height disturbances reach a region where they can cause sufficient warming for the feedback of Lau (1981) and Philander et al. (1984) to be unstable, they will grow. A key element of this growth is the generation of wind anomalies in the center of the basin, and it is these winds which generate thermocline perturbations of opposite sense which propagate westward from the wind. Eastward-moving perturbations of one sense thereby give rise to westward-propagating ones of the opposite sense.

In short, the envisioned cycle is as follows. Warm SST anomalies in the east generate westerly winds in the center of the basin. These winds cause thinning of the thermocline on the western side of the wind, thinning signals which propagate westward to the boundary. At the western boundary, these signals reflect and propagate back to the east. When they arrive in the east, the thinning produces a cooling tendency. This cycle then repeats, but with opposite sense. Cooling drives easterlies in the center of the basin, and therefore thickening of the thermocline in the west. A transit of these thickening signals to the western boundary and back to the east completes the cycle by generating warming. Thus, the original warm SST conditions are restored after two excursions of the signal from the winds to the western boundary and back to the region of SST anomalies. In this simple view the period would be twice the transit time. In the full model, additional effects can cause the period to be longer, but not shorter than this.

4. Additional calculations

The results of the previous section suggest that coupled feedbacks play a crucial role in maintaining the model’s low-frequency variability, and that, although
solutions are somewhat chaotic in appearance, the model has selected a time scale of about three to four years.

Our interpretation of the results is that at some sufficiently strong coupling the steady state solution (or solutions) becomes unstable through processes similar to those discussed by Lau (1981), Philander et al. (1984), and others. When the coupling is stronger than this critical value, a stable time-dependent solution appears which in our case is periodic or nearly so. To confirm the critical importance of the coupling and to explore the nature of the low-frequency variability, we performed several additional calculations with a still simpler model.

First, we considered the possibility (admittedly small) that the observed variability could be generated by the ocean alone under steady atmospheric forcing. We experimented by making a run in which the ocean was forced with the time-averaged winds and atmospheric temperatures from the original run (shown in Fig. 3), and was initialized from the end of that run. The model quickly settled to a nearly steady state, indicating that coupling, or at least atmospheric variability, is needed to explain the results. We note in passing that in this run temperatures in the cold tongue region in the east fall, and the east–west temperature gradient increases. So the low-frequency “El Niño” signal is tending to offset the cooling due to time-mean upwelling in the east.

Although the atmospheric model used is “dry”, there is still some tropical variability, presumably the result of extratropical phenomena. Therefore, a second possible explanation of the results is that natural variability in the atmosphere, having nothing to do with the coupling, is sufficient to excite an organized low-frequency response in the ocean. This would occur if, for example, the ocean alone had a neutral or weakly damped mode at the appropriate frequency. The obtained behavior could then be thought of as the excitation of an oceanic resonance. To test this “random forcing” hypothesis.
we made another calculation in which we ran only the atmosphere, forcing it with the ocean's time-mean results from the original calculation—just the reverse of the previous test. We then took the full history of winds and temperatures from this run and used them to force the ocean. Thus the test includes no coupling, but does force the ocean with atmospheric variability typical of an uncoupled case. The result was that oceanic variability increased, but only to about 0.1°C, an order of magnitude less than in the coupled run. Maximum SST variability occurs in the same region in the eastern half of the basin—as one would expect since the shallow thermocline makes the ocean most sensitive there. These results also showed no sign of a preferred time scale. At periods beyond a year the spectrum is weak and white.

We are thus led to the conclusion that the results depend on the destabilization of the system by coupling. We would then like to consider two further possibilities. One is that no bifurcation has taken place but the system is near one, and that a weakly stable mode of the desired frequency (in this case a mode of the coupled system) is being excited by atmospheric noise (weather). The alternative is that the system has undergone a bifurcation into a time-dependent state, through a coupled instability of the stationary state.

To distinguish between these two possibilities, we need to remove the atmospheric noise while preserving the coupling. We think we can effectively do this by using a linear atmospheric model which faithfully reproduces the primitive equations model's stationary tropical response of surface winds to heating while excluding extratropical synoptic-scale transients.

The linear model follows closely that of Gill (1980), and may be viewed as the two-level model linearized about a state of rest. In our notation

\[
\frac{\partial \bar{u}}{\partial t} = f \bar{v'} - \frac{\partial \bar{\phi'}}{\partial x} - k \bar{u}' \tag{38a}
\]

\[
\frac{\partial \bar{v}}{\partial t} = -f \bar{u'} - \frac{\partial \bar{\phi'}}{\partial y} - k \bar{v}' \tag{38b}
\]

\[
\frac{\partial \bar{\phi}}{\partial t} = -\hat{\delta} [\partial \bar{u}'/\partial x + \partial \bar{v}'/\partial y] - k \bar{\phi} + \bar{Q} \tag{38c}
\]

\[
\bar{\psi} = c_p \bar{\theta} \bar{P} \tag{38d}
\]

x and y are Cartesian coordinates on an equatorial β-plane, \( f = \beta y \), and \( \bar{\theta} \) is a fixed constant (=15°). The rest of the notation is as before. As in Gill and numerous other studies using this set we take a rather large value for \( k \) (1/k = 5 days). Equations (38) are the
simplest model for the atmosphere’s response to a deep heating in the tropics. They describe an overturning, with equal and opposite motions in the two layers.

Before using (38) in the coupled problem, we made a simple test to see how well this system can mimic the full model’s response. We took \( \dot{Q} \) to be the heating anomalies from the coupled run. These we stored as monthly means and smoothed in time with a 1-2-1 filter to retain only the low-frequency forcing. Equations (38) were then finite differenced on a 400 km \( \times \) 500 km grid and integrated forward in time. At each step the forcing was interpolated between the smoothed monthly values. Figure 11 shows the result for \( -\tilde{u} \), the lower-level (“surface”) wind, at the equator. Evidently the tropical low-frequency response of our full primitive equation model is simulated very well by these strongly damped linear equations. One suspects very little has been gained from the nonlinear atmospheric calculation, other than the satisfaction of producing the spontaneous low-frequency variability in a “complete” system, capable of attaining a consistent coupled climatology.

To use (38) in the coupled problem we need to interpret the solution as the perturbation response to an anomalous forcing dependent on predicted ocean temperatures, and use it in turn to modify the wind-stress forcing the ocean. This is essentially the problem done by Zebiak (1984) and Cane and Zebiak (1985) with a somewhat simpler ocean, more complicated coupling, and for a variety of prescribed oceanic basic states, including seasonally varying cases. We will use our own predicted climatology as the “basic” state and do the coupling as follows.

Let \( T_0(x, y) \) be the distribution of SST obtained in the “steady forcing” run already discussed in this section. We will make \( Q' \) proportional to deviations of the predicted ocean temperatures from \( T_0 \), and continue to force the ocean with the steady winds which produced \( T_0 \) (the climatological winds from the original run), but modify them by \( -\tilde{u}' \) and \( -\tilde{v}' \) as produced by (38). We will also continue to remove (or add) heat from the ocean as given by Eq. (33), but with \( T_0 \) fixed at the full run’s climatology (that is, for damping SST perturbations we ignore changes in \( \bar{\theta} \), and \( \bar{\theta} \) is already fixed).

This somewhat complicated way of doing the “anomaly coupling” has the nice property that, ignoring the very small natural variability of the ocean alone, \( \tilde{u} = \tilde{u} = \tilde{v} = 0 \) is a stationary state of the nonlinear coupled system, since the ocean is then forced by the winds that produced \( T_0 \), and \( T = T_0 \) implies \( Q' = 0 \). Obviously such an equilibrium state can always be constructed from a steady wind distribution. Simply spin up the ocean with the given winds and then couple only anomalies on both sides. The stability of such a constructed state depends both on the exchange coefficients and the choice of winds, which determines the basic state.

Coupling Gill’s model and the full ocean in this way, we made a 15-year run whose results are summarized in Figs. 12 and 13. Initially the model is at the constructed equilibrium state, but it quickly leaves it and settles to an oscillatory solution with a period of 30 months. This is somewhat shorter than the typical time scale one might estimate from the much more irregular history in Fig. 4. But the character of the oscillation is very similar to the behavior obtained before. This is illustrated in Fig. 13, which shows the same collage of the oscillation as was presented in Fig. 10 for the full run. In the eastern half of the basin, temperature and equatorial height perturbations are strongly correlated (Figs. 12 and 13b); winds driven by the resulting heating are, as before, in the sense to give a positive feedback with the upwelling/downwelling SST perturbations; and the propagation to the western boundary and back to the region of SST perturbations again gives the sense (stronger in this periodic solution) of a cycle whose period is twice the transit time or more.

The main result of this final calculation is the absence of chaotic behavior. Our hypothesis is thus that the original results can be explained as a bifurcation of the stationary state into a purely oscillatory solution (or limit cycle). Obviously, understanding the system’s behavior near this bifurcation, in particular the frequency of linearly unstable solutions, is of primary importance.

We have not undertaken such a linear analysis, and in fact have only dabbled in attempts to vary parameters to bring the system near to the bifurcation. We can report, however, on a set of somewhat technical
and produces the main features of the zonally averaged circulation when forced with differential heating between pole and equator. The ocean is a two-layer, primitive equations model for the upper ocean in an idealized tropical basin the size of the Pacific. The model was designed to study the low-frequency behavior of the coupled system. It represents an extension of previous models and simplification of the real world in several aspects:

1) Meridional boundaries are present in very simple geometry. The work of Lau (1981) and Philander et al. (1984) did not consider the reflection of wave energy from such boundaries and its impact on the resulting solutions. As in the work of Anderson and McCreary and Cane and Zebiak, the role of the western boundary seems to be crucial in our results.

2) The coupling is based on an explicitly computed SST. Indeed, as in the work of Cane and Zebiak, temperature advection in the ocean is the main nonlinearity. Although this is done with very limited vertical resolution, the prognostic equations for the surface temperature allow for important distinctions between upwelling and downwelling and for horizontal structure in the sensitivity of SST to surface winds.

3) There is no seasonal cycle or any other externally applied variability. This simplification was not made because we thought effects associated with the seasonal cycle were unimportant, but rather because we wanted to see if they were crucial.

4) Although the models are highly idealized, the calculations are in the spirit of coupled GCM calculations, with the model determining its own climatology.

Most of the results presented are from a single 35-year integration. In this run, the system was forced by relaxing the atmosphere to a prescribed zonally symmetric “radiative” equilibrium, to simulate annual mean conditions. This drove an atmospheric general circulation, which in turn spunup the ocean circulation. Analyzing the last thirty years of the run, we found that in spite of the steady external forcing, the ocean and the tropical atmosphere underwent ENSO-like irregular oscillations with a time scale of several years (Fig. 4).

We also did two additional calculations in which the coupling was suppressed, but neither showed low-frequency variability comparable to that of the original run.

Finally, we did a coupled calculation in which the atmospheric model was replaced by Gill’s model and the coupling was done only on anomalies about the 30-year climatology of the original run. In this case low-frequency variability was once again present (Fig. 12), but the solution was oscillatory with a period somewhat shorter than the apparent preferred time scale of the irregular oscillation. Otherwise, it was very similar to the full calculation.

5. Summary and discussion

We have analyzed the natural variability of a coupled ocean–atmosphere model. Based on the primitive equations of motion, the atmospheric model is global sensitivity experiments using the Gill model, in which we gradually increased the maximum depth the ocean’s two layers are allowed to reach before water is removed (“downwelled” to the abyssal layer). This parameter affects the mean depth of the “thermocline” (the base of our lower layer). With east–west pressure gradients more or less unchanged to balance (to first order) the mean trades, this strongly influences thermocline depth in the cold tongue region, and thus the sensitivity of SST and the strength of the coupling. Reducing the coupling by increasing this value we find that the amplitude of the oscillation diminishes until a warm stable stationary state appears. Interestingly, however, the transition from unstable to stable conditions appears to occur with a finite frequency.
Two key questions presented by these results are: where does the system obtain the energy to maintain the vacillations, and how is the time scale determined? Both revolve around the destabilization of the system from a stationary equilibrium state into time-dependent solutions.

a. Destabilization

A primary finding of this study is that the coupling between ocean and atmosphere is fundamental to its low-frequency variability. Not only does an analysis of the time scales of the oceanic SST balance indicate that some such feedback must be present, the calculations with uncoupled but otherwise identical models confirmed that there is no inherent oceanic mode which provides the low-frequency energy, nor does there appear to be a near-neutral mode excited by random atmospheric forcing. The removal of the annual cycle eliminates a priori the hypothesis of a nonlinear subharmonic response to the annual cycle.

The mechanism which operates here to provide instability is essentially that discussed in Bjerknes (1966), Lau (1981), Philander et al. (1984) and Yamagata (1985): unstable coupled Kelvin modes. Maximum zonal wind anomalies were found over the center of the basin—just west of the SST anomalies—and were of the sense needed to maintain and enhance the SST anomalies by tilting the thermocline. Checking on the necessary condition for instability described by Yamagata (1985), we found that the product of surface stress anomaly and surface current anomaly was almost always positive in our runs. The SST budget shows dominance of vertical advective cooling over horizontal advection. As discussed by Hirst (1985), this will favor the Kelvin mode of unstable growth over the Rossby.

Finally, as in the stability analysis of the linear models, a critical value of the strength of the coupling (some product of the sensitivity of SST to winds and winds to SST) must be exceeded before the steady solution gives way to the vacillating one. In runs with the parameters modified to make the ocean layers thicker, the SST response to wind perturbations is diminished and solutions settled into nonvacillating, warm states, akin to a permanent El Niño.

b. Time scale

The most surprising result of these calculations, as well as those of Anderson and McCreary (1985) and Cane and Zebiak (1985), is the very long time scale the models have selected. Our simplest expectation is that the system choose a time comparable to the transit time of waves back and forth across the basin. At these
low frequencies, the eastward leg of this transit must be at Kelvin wave speed, and the westward leg (discounting edge Kelvin waves, which are damped in the model) at the speed of the first few Rossby modes. For free neutral waves (using the first Rossby mode and our parameters) this transit time is roughly 350 days. Coupled unstable modes are a bit slower, but only the Kelvin leg would be affected and so significantly longer periods would not be expected.

We have tried to reconcile this discrepancy by arguing that the shortest time one might expect for coupled solutions in a closed basin like ours should be twice a transit time, not between meridional boundaries, but between a relatively localized “active” region and the western boundary. Several elements enter this argument.

First, east of the SST anomalies there is very little oceanic response to the wind forcing (Fig. 9). Thus the eastern boundary does not appear to play a role.

Second, SST anomalies in the western half of the basin—that is, west of the maximum wind stress anomalies—are very small. In nature, one might argue that even these small anomalies in the western Pacific can have a strong effect on the atmosphere. However, this is not the case in our model, in which the atmospheric heating is roughly proportional to the SST anomaly. The oceanic variability in the western half of the basin is thus forced, but uncoupled, response to wind anomalies produced by anomalies in atmospheric heating over the eastern half.

Third, the oceanic response west of the “active” region consists of a westward-propagating part emanating directly from the wind anomalies and an eastward-propagating part reflected from the western boundary.

Finally, the reflection of the signal at the eastern end occurs through the atmosphere. The incident signal grows with SST anomalies whose wind response forces the westward propagating signal. As we argued in section 3 this “coupled reflection” is phase reversing, unlike the phase preserving reflection at rigid boundaries. This phase reversal lengthens the period by at least a factor of two and helps account for the longer time scale.

All of these features are present in both the full model set and that with the atmospheric equations linearized. In the latter case, the system clearly exhibits oscillatory behavior. The full system appears somewhat irregular, with a broader spectrum about slightly lower frequencies. It is our opinion that the full system chooses its time scale for essentially the same reason as the oscillatory one, but is disturbed from the oscillation by the perturbations associated with the atmospheric variability.

Vallis (1986) presents a simpler system which exhibits chaotic behavior and has no preferred time scale. While all of the physical processes invoked by Vallis are modeled (with greater fidelity) in our study and that of Cane and Zebiak (1985), these results show selection of a preferred time scale, not a chaotic behavior.

c. Predictability

If the system is oscillatory, it is predictable. Knowledge of the state at any one time in Fig. 13, for instance, will allow for very good prediction many decades in the future. Our results with the full model are not nearly so periodic. Small anomalies in surface height in the western part of the basin often precede large SST and wind anomalies (see Fig. 10a and 10c); conversely, large height anomalies sometimes fail to lead to large wind changes. We have argued that the mechanism which acts in the full model is the same that produces regular oscillations with the linearized atmospheric dynamics and that the more irregular solutions of the full model are due to “weather” perturbations of the regular cycle. In nature, with a broader spectrum of tropical phenomena and a superposed annual cycle, such external perturbations would be even greater and further reduce predictability. Thus, the distinction between chaotic behavior (implying a lack of predictability) and this view of an easily perturbed oscillator may not be of much practical importance.

White et al. (1985) have found evidence for precursive Rossby wave activity in the western Pacific associated with ENSO: XBT data suggest a deepening of the thermocline prior to El Niño events, with some suggestion that higher latitude sections show earlier thermocline displacements. As just mentioned, however, the strength of the surface height perturbations at 5°–7° are not necessarily good predictors of eventual El Niño strength. Similarly, the arguments of Wyrtki (1985) that a precursive “build up” of surface height in the western Pacific is related to El Niño imply some predictive skill in this relationship. This build up is another expression of the Rossby wave kinematics, and the lack of a strong predictive relationship between tide gauge measurements and El Niño strength is also akin to the lack of a strong relationship between the height perturbations in Fig. 10a and the eventual ENSO strength (best thought of here as the wind anomaly in Fig. 10c).

d. Summary

We wish to emphasize the following points:

• ENSO-like, low-frequency internal variability was found in nonlinear coupled atmosphere–ocean models.

• The destabilizing mechanism was argued to be the “coupled instability” of Lau (1981) and Philander et al. (1984), albeit in a more complicated setting.

• The bifurcation resulting from the “coupled instability” of the stationary state in a closed ocean basin leads to a nonlinear oscillation whose period is the underlying time scale of ENSO.
• Equatorial trapped waves in the ocean appear to play a crucial role in determining the time scale.

In the discussion of the results we have begun to address the question of what mechanism is responsible for the long time scale of ENSO. We feel the kinematic “period-doubling” argument presented is central to understanding the model’s low frequency. In subsequent papers we will speculate on the dynamics of the nonlinear oscillations.

Acknowledgments. The authors gratefully acknowledge the helpful suggestions and comments made by Dr. D. E. Harrison and an anonymous reviewer. This work was jointly supported by the Oceanic Processes Branch and the Global Scale Atmospheric Processes Research Program of NASA Headquarters.

REFERENCES


