TROPICAL ATMOSPHERE-OCEAN INTERACTION AND

THE EL NIÑO/SOUTHERN OSCILLATION PHENOMENON

by

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ABSTRACT

A coupled atmosphere–ocean model is developed and used to study tropical Pacific events associated with ENSO. The atmosphere and ocean components are constructed separately, using prescribed forcing from a composite of observations during recent ENSO events. It is demonstrated that much of the observed atmospheric variability in the equatorial zone can be understood as a thermally forced response to changes in equatorial SST during ENSO. The simple model used here is steady-state and linear except in its treatment of heating. The nonlinearity in heating is indicated to be very important, and can be identified with the effects of low-level transport of water vapor in the tropical atmosphere.

With the present ocean model the observed SST anomalies can be understood as a redistribution of heat in the equatorial upper ocean, forced directly by the sizeable equatorial wind stress anomalies during ENSO. The model dynamics are linear, and include a single baroclinic mode plus a frictional surface layer. The thermodynamics are more complete, and allow an explicit evaluation of the individual components of the surface layer heat budget. An important conclusion of the study is that the oceanic surface layer and all components of the temperature prediction equation are essential to give realistic SST anomaly fields.

In the coupled calculations, the two components are combined, and the system is allowed to evolve freely from a prescribed initial state. For the parameter choices indicated by the forced calculations, the coupled system exhibits warm events with many realistic features, including amplitude, spatial structure, inter-event spacing, and temporal development which is 'phase-locked' to the annual cycle. Through a number of additional calculations, several conclusions are reached. Among them are the following: (i) the mean temperature, wind, and current fields largely determine the observed spatial pattern of ENSO anomalies, (ii) the normal annual cycle in mean wind, temperature, and current fields amounts to a modulation in the positive feedback between atmospheric and oceanic anomalies, and thus determines the temporal development and duration of warm events, (iii) both local and remote interactions between the atmosphere and ocean are essential to give ENSO-like variability, and (iv) well-defined, deterministic physics can account for the observed aperiodicity of ENSO events. Finally, the results suggest that ENSO is explainable as essentially a tropical Pacific phenomenon, despite its apparent influence over a much larger domain.

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Chapter I

Introduction

From time to time, and to various degrees, the following events are observed to occur. Coastal regions of Ecuador and Peru are inundated with several times the normal rainfall, while regions further inland experience dry conditions. The normally cool, nutrient-rich waters off the coast become anomalously warm, and the local fishing industry suffers catastrophic losses. In the central Pacific, near-equatorial islands experience very heavy rainfall and increased incidence of tropical cyclone development. Further west, the Indonesian region receives somewhat less than normal rainfall, and eastern Australia experiences severe drought. Seasonal climate anomalies are often observed in regions very far away, for example over India during summer, and over North America during winter. After several seasons of anomalous conditions in the equatorial Pacific, conditions return to near normal for perhaps two years, or perhaps ten years, before another such major episode occurs. These events constitute what is perhaps the most significant known pattern of interannual climate variability on the timescale of decades or less (Kidson, 1975, Weare et al., 1976).

Although some features of the large-scale atmospheric oscillations were identified several decades ago (Walker, 1923), it is only much more recently (Wyrtki, 1975, Bjerknes, 1966, 1969, 1972) that concomitant large-scale oceanic fluctuations were identified and considered potentially important during these events. (Before this, El Niño was known only as a coastal phenomenon for the ocean.) A great deal of observational
and modeling work has now been done on the problem. The emerging view is that these events involve mutual interactions between the tropical atmosphere and ocean. Such is the subject of this study.

The most recent event occurred during the course of this work, in 1982-3. Not only is this the most pronounced event ever recorded, but it is also the most thoroughly documented. Perhaps most importantly, the event exhibited unprecedented characteristics in its development which have forced as reappraisal of pre-existing theories for ENSO (El Niño/Southern Oscillator). This new information has contributed significantly to the direction of this work.

In addition to the variations in precipitation and coastal SST mentioned above, there are a number of prominent common features to the recorded ENSO events. Among them are the following.

(a) — coupled oscillations. There is a truly remarkable temporal correlation between interannual variations in eastern tropical Pacific SST (sea surface temperature) and numerous meteorological variables — surface pressure, surface wind, 200 mb wind, mean tropospheric temperature, cloudiness and rainfall — in regions spanning the entire tropical Pacific Ocean (see, e.g. Julian and Chervin, 1978; Angell, 1981; Newell et al., 1982; Arkin, 1982; Kidson, 1975; Heddinghaus and Krueger, 1981; Rasmussen and Carpenter, 1982). These findings clearly demonstrate the effective communication between the tropical upper ocean and the atmosphere on interannual timescales.

(b) — spatial structures. The original index of the Southern Oscillation (Walker, 1923) was the difference in surface pressure anomalies between the eastern Pacific ocean region and the eastern Indian Ocean
region. Subsequent study has confirmed that such is the scale of atmospheric variability associated with ENSO events (Egger et al., 1981, Barnett, 1983). During the mature phases of an event, spatially coherent patterns of anomalous surface wind and SST extend along the equator from west of the dateline to the eastern coast, with the largest wind anomalies in the central Pacific, and the largest SST anomalies in the eastern Pacific. The characteristic anomaly patterns additionally extend poleward in each hemisphere for at least 1500 kilometers (Weare et al., 1976, Rasmussen and Carpenter, 1982). Nearly one quarter of the world is directly affected by ENSO.

(c) — atmospheric vertical structure. There is a markedly out-of-phase relationship between 200mb and near-surface level wind anomalies in the equatorial central Pacific, as seen, for example, by comparing the Arkin (1982) 200mb wind analysis with the Rasmussen and Carpenter (1982) surface wind analyses, or comparing the 200mb and 850mb wind fields in Arkin et al. (1983). Calculations with greater vertical resolution (Gutzler, 1984) indicate that this is the dominant scale of vertical variability in zonal winds near the equator in the western Pacific.

(d) — timescales. ENSO behaves more like an event than a regular oscillation. Anomalous conditions can develop over a time period much smaller than either the duration of anomalous conditions or the inter-event time period. Rasmussen and Carpenter (1982) show (see their Figure 2) that in the interval from 1953 to 1974, the dominant period in the spectra for several ENSO indices lies between three and four years. Their Figures 8 and 15 show that the duration of warm events is slightly greater than one year. It is significant that these time periods are
not intrinsic either to the atmosphere or ocean alone — thus the suggestion that ENSO is an essentially interactive phenomenon. In Chapter IV it is demonstrated that a coupled system can exhibit such behavior.

(e) — phase locking. There is a marked tendency for anomalous winds and SST in the central Pacific to achieve their largest amplitude near the end of the calendar year, and decay rapidly thereafter, during an event. In the far eastern Pacific there is also a tendency for SST anomalies to peak around May or June. In the composites of Rasmussen and Carpenter(1982), the peaks in SST anomaly occur first in May at the coast, and then in December in the eastern and central Pacific. However, in the 1982-83 event, the first major peak occurred in December in the eastern and central ocean and then a separate maximum occurred at the coast the following May-June. Thus, although the order of the peaks was different, the time of the year that they occurred was the same.

(f) — extratropical connections. There are numerous studies which relate meteorological events in the extratropics to ENSO (e.g., Walker,1923, Horel and Wallace,1981, van Loon and Madden,1981, Chiu et al.,1981, Trenberth,1976, Pan and Oort,1983, Wallace and Gutzler,1981). There are also studies relating ENSO to fluctuations in the Indian Monsoon circulation (Barnett,1983, Angelil,1981). Whether there is a precise relationship between ENSO and midlatitude climate anomalies or the Indian monsoon is still an open question, although such possibilities make the problem even more intriguing.

A number of fundamental questions are raised by these observations. What initiates events, and what causes their termination somewhat more
than a year later? Why is the development phase-locked to the annual cycle? What determines the characteristic spatial structure of ENSO anomalies? Which processes are most important in the interaction between atmosphere and ocean during ENSO? For example, is the essential interaction largely local or remote? Is ENSO primarily a tropical Pacific phenomenon, which in turn affects other regions, or must one consider, say, extratropical influences to explain it? Can the observed aperiodicity be understood in a deterministic sense, or are stochastic processes fundamentally important? Are ENSO events predictable, and if so, how far in advance? These are among the most interesting questions to be addressed by observationalists and modelers, and they constitute the focus of this work.

A coupled atmosphere-ocean model has been constructed and used to study ENSO. The goal has been to find the simplest model which can simulate the main features of the phenomenon as outlined above, while at the same time specifying as little as possible. As will be seen, the model falls between two categories represented by previous studies: those in which all or most of the physics is highly parameterized, and GCM-level models. The difficulty with the highly parameterized models is that the physics is not explicit, so that it is difficult to interpret the results physically. With GCM models it is difficult to separate out the individual contribution of a given process, and thus it is hard to identify what is essential and what is of secondary importance. (Also, calculations with GCM models are very expensive.) The hope is that with an intermediate level model it may be easier to capture and identify the essential physics.
The model developed here is a perturbation model — that is, in both the atmosphere and the ocean, no attempt is made to simulate the mean fields or the annual cycle. Rather, these are specified and only the departures from mean conditions are explicitly calculated. Simulating the mean and annual cycle in the tropical atmosphere or ocean is a very complex problem, and is not possible with a simple model. That problem amounts to understanding the time-dependent general circulation of the atmosphere and the ocean, perhaps not only in the tropics, but globally. The hypothesis of this study is that the essential physics governing ENSO anomalies is less complex, so that in an appropriate perturbation model the problem can be simplified considerably.

Only deterministic processes are examined. The intent is to determine whether the observed aperiodicity of ENSO could result from physics already known to be important in the phenomenon, and thus included in the model, rather than from some additional (and unidentified) effect.

Finally, only the tropical Pacific domain is considered. It is the goal of this study to understand what aspects of ENSO can result specifically from tropical interaction between the atmosphere and ocean in the Pacific sector, where the largest systematic climate anomalies occur.

The approach in building the final coupled model has been to develop and test the component models independently, and then to combine them. The atmosphere model is tested by prescribing the SST-field from observations and attempting to reproduce the observed wind anomaly field. Similarly, the ocean model is tested by prescribing the wind stress field, based on observed wind anomalies, and attempting to reproduce
the observed SST anomaly field. The observed fields are taken from
the composite ENSO analyses of Rasmussen and Carpenter (1982). After
gaining some knowledge of the component models in this manner, they
are coupled for the interactive experiments which were designed to examine
the fundamental questions posed above. In Chapter II, the atmospheric
model is presented and discussed in detail, together with model results
and corresponding observational results. Chapter III follows the same
course with the ocean model. Chapter IV deals with the coupled model
calculations, and compares the present results to those of other investi-
gations. Finally, Chapter V attempts to interpret the results from
a physical standpoint, and to speculate on the successes and failures
of the model. Some opinions are also offered as to fruitful directions
for future studies.
CHAPTER II

Atmospheric Modeling of ENSO

2.1 Introduction

How do tropical SST anomalies affect the atmosphere? Ultimately the response must be understood in terms of a heating anomaly which develops as a result of the changing thermal boundary conditions. The processes involved can be complex, however, and may involve feedback mechanisms which make the problem anything but straightforward. The ocean communicates with the atmosphere through the boundary flux of sensible heat and the flux of water vapor which evaporates from the surface. These fluxes depend directly on the sea surface temperature, but they also depend (non-linearly) on the wind speed, the near-surface air temperature, and the near-surface specific humidity. Furthermore, in the case of water vapor, the heating effect is not realized until condensation occurs. Thus evaporation can occur in a given region without any associated local heating — that is, the moisture can be advected by the low level wind field to another, possibly distant region where active convection is occurring, and only then release its latent heat. Beyond this, there is still another potential feedback process: an increase in convective activity means an increase in cloud cover, which may alter the local radiation budget and produce additional heating anomalies.

Obviously, the problem is not at all simple. What is remarkable is that very simple linearized atmosphere models which parameterize heating directly in terms of SST anomalies give results which in some respects are similar to observations and to GCM simulations. The model
used here is of this type. Some hypotheses concerning the limited success of such a simple model will be presented in Section 2.8.

This chapter and the next will be concerned with comparing model results for low-level wind and SST anomalies with observations from recent ENSO events. The observations will be displayed and discussed along with the model results to facilitate the comparisons. The present model has two versions: one with a convergence feedback parameterization, and a simpler version without it. They will be presented separately along with a scaling analysis in the following sections. Subsequently, the simulation experiments will be described.

2.2 Atmosphere Model I (without convergence feedback)

The starting point for developing this model was the work of Gill (1980), who in turn specialized the earlier work of Matsuno (1966). Gill specifically addressed the problem of understanding the low level tropical wind field in terms of zonally asymmetric heat sources. By assuming a distribution of heating which is designed to mimic the large heat source over the Indonesian region, plus another idealized axisymmetric line source of heating north of the equator which approximates an ITCZ, he obtains a wind field which in some respects resembles the mean wind field in the tropical Indian Ocean and Pacific Ocean regions. Early in this work the same model was applied to the ENSO problem by assuming that anomalous heat sources could be directly related to SST anomalies. The results are reported in Zebiak (1982). The model, with some modifications, has been used in the subsequent work to be reported here.

In the model, the problem is approached by assuming a vertical structure for the flow which has the same form as the prescribed thermal
forcing: $\sin (\pi Z/Z_T)$ for $0 < Z < Z_T$, where $Z_T$ is the tropopause height (the convenient vertical coordinate is physical height because a Boussinesq approximation is made). Evidence that the heating due to cumulus convection has such a distribution is given by Yanai et al. (1973) and others. If one assumes a vertical structure of $\cos (\pi Z/Z_T)$ for the wind and pressure fields, and $\sin (\pi Z/Z_T)$ for the vertical velocity, then the linear equations describing the flow become separable in $Z$, and one arrives at the so-called divergent barotropic equations (or shallow water equations). These have the following form on an equatorial beta-plane:

\[ (u_a)_t - \beta_0 y v_a = (-1/\rho_0) p_x + F \]  \hspace{1cm} [2.2.1a]

\[ (v_a)_t + \beta_0 y u_a = (-1/\rho_0) p_y + G \]  \hspace{1cm} [2.2.1b]

\[ (p/\rho_0)_t + N^2/m^2 ((u_a)_x + (v_a)_y) = -agQ/mc_p \]  \hspace{1cm} [2.2.1c]

where $N$ is the buoyancy frequency, $g$ is the acceleration of gravity, $\alpha = 1/T_0$ where $T_0$ is a mean tropospheric temperature, $c_p$ is the specific heat of air at constant pressure, $\rho_0$ is the reference density, $F$ and $G$ are external momentum forcing terms, and $m = \pi/Z_T$ where $Z_T \sim$ tropopause height.

Appendix A gives a more complete derivation of equations (1). If the equations are nondimensionalized according to

\[ u^* = U_a u_a \]

\[ v^* = U_a v_a \]

\[ p^* = \rho_0 U_a c_a p \]

\[ x^* = (c_a / 2\beta_0)^{1/2} x \]
\[ y^* = (c_a / 2 \beta_o)^{3/2} y \]
\[ t^* = (2 \beta_o c_a)^{-1/3} t \]
\[ (c_a = N/m) \]

(dimensional variables temporarily denoted with asterisks), then the nondimensional equations have the form

\[ (u_a)_t - yv_a / 2 = -p_x \] \[ (2.2.2a) \]
\[ (v_a)_t + yu_a / 2 = -p_y \] \[ (2.2.2b) \]
\[ p_t + (u_a)_x + (v_a)_y = -Q_1 \] \[ (2.2.2c) \]

where \( Q_1 = \frac{a g}{m c_p} \frac{\dot{Q}}{(2 \beta_o c_a)^{1/3} U_a} \), and \( \dot{Q} \) is the dimensional heating rate/unit mass. Gill's system is obtained by considering steady circulations in the presence of heating and dissipation. He parameterizes momentum dissipation with Rayleigh friction, and thermal dissipation with Newtonian cooling. He further assumes the decay time of both processes is the same, leading to:

\[ s u_a - yv_a / 2 = -p_x \] \[ (2.2.3a) \]
\[ s v_a + yu_a / 2 = -p_y \] \[ (2.2.3b) \]
\[ s p + (u_a)_x + (v_a)_y = -Q_1 \] \[ (2.2.3c) \]

\( (s = (2 \beta_o c_a)^{-1/3} s^*, \text{ where } s^* = (\text{dimensional decay time})^{-1}. ) \)

The relevance of the model has been demonstrated empirically by several authors. It is nonetheless important to examine the validity...
of the model from a theoretical standpoint. Perhaps the primary point of controversy is the assumed form of the vertical structure. According to linear inviscid theory, one expects low frequency thermally forced tropical motions to consist of vertically propagating waves with short vertical wavelengths, as described, for example, by Lindzen (1967). The structure assumed by Gill amounts to placing a rigid lid on the atmosphere at $\bar{Z} = Z_T$, which, of course, precludes the possibility of vertical energy propagation. Furthermore, the scale of the vertical structure does not correspond to that of the inviscid waves.

The relevance of the model rests on an assumption about the effect of small scale motions on the large scale flow. The primary heat source for large scale tropical motions arises from the net effect of cumulus convection (e.g., Nobre (1983)). If one assumes that the same small scale vigorous motions associated with this heating also effectively mix momentum, then the problem of interest is one of forced, strongly dissipative motions rather than forced inviscid motions. Chang (1977) has examined the linear dynamics in the presence of dissipation. He found that the dissipation had the effect of inhibiting vertical propagation, and eliminating waves with small vertical wavelengths. Geisler and Stevens (1982) examined the non-Boussinesq linear equations with dissipation by placing a lid at a very high altitude (60 km) and calculating a large number of vertical modes. They used a heating profile like that of Gill and summed up the modes to examine the validity of considering a single mode for which $w$ vanished at the same altitude as the heating. They found the response became negligibly small well below their lid, and that the final solution had a structure qualitatively like that of the forcing (for steady or low frequency forcing). Thus it would
seem that if one accepts the idea that strong dissipation accompanies strong heating for the large scale motions, the theory offers some support for assuming a vertical structure as Gill does. If the dissipation is very effective so that vertical propagation is highly suppressed and small vertical scales are eliminated, then one may expect the response to have a vertical scale like that of the forcing. The situation with respect to the horizontal distribution is different, as will be discussed later.

Observations of low-frequency, large-scale tropospheric motions in the tropics seem to agree much better with the 'viscous' theory than with the inviscid theory. This may be considered evidence that the assumptions have some validity, or at least that the resulting model has some relevance to the problem.

Another potential problem with the model is that it is linearized about a state of rest, so that no inertial effects are considered. Hoskins and Karoly (1981), using a 5-layer baroclinic steady state model, found that for a subtropical thermal forcing, the local tropical response was rather insensitive to the specified mean zonal flow. This was not the case for the remote response in midlatitudes. The same conclusions were reached by Simmons (1982). Still other investigators, using GCM models, (e.g. Keshavamurty (1983), Shukla and Wallace (1984)) have found that the local tropical response to heating anomalies resembles Gill's result despite the neglect of inertial effects in the latter. Nobre (1983) developed a nonlinear model to study the atmospheric response to specified heating. His results also show that even for a large tropical heating anomaly, the near-equatorial response is qualitatively the same as the linear results with zero basic state. Again, this
is not true away from the equator, where the nonlinear results depart significantly from the linear results.

In light of this, it is felt that the simple linear model is a reasonable starting point for the present simulation studies because the focus is on the tropical domain. As will be shown in Chapter III, from an oceanic standpoint the near equatorial events are what matters most. In order to successfully simulate the observed winds outside the equatorial zone, it almost certainly would be necessary to include inertial effects.

In the simplest version of the present model, heating is parameterized in terms of SST anomalies by considering changes in evaporation. Since the model assumes a steady heat source, equilibrium changes in evaporation are computed by linearizing the Clausius-Clapeyron relation about the mean state SST-field (which is specified in terms of monthly means). This gives

$$\Delta e = E_0 (\text{SSTA}) (d(e_s)/dT) \bigg|_{T=T_M} \sim E_0 (\text{SSTA}) \frac{b}{T_M^2} e^{-b/T_M}$$  \hspace{1cm} [2.2.4]

where $\Delta e$ is the evaporation, $T_M(x,y)$ is the mean SST, $E_0$ and $b$ are constants. Scaling the evaporation by its value for $T_M = T_{\text{ref}}$, a reference temperature, and assuming that local atmospheric heating is proportional to this evaporation, gives the following form for the heating:

$$Q_1 = \alpha (\text{SSTA}) \frac{T_{\text{ref}}^2}{T_M^2} \exp \left( b \left( \frac{1}{T_{\text{ref}}} - \frac{1}{T_M} \right) \right)$$  \hspace{1cm} [2.2.5]
An estimate for the appropriate value for the constant \( \alpha \) is derived in Appendix A. With this relation the effect of a given SST anomaly is reduced in regions where the total SST is cool. For the range of \( T_M \) in the tropical Pacific the difference in \( Q_1 \) for a given SST anomaly varies by nearly a factor of 2.

Equations [2.2.3], together with [2.2.5], comprise the simplest version of the present model.

### 2.3 Some Simple Scale Analysis

Before proceeding to simulation studies, it is instructive to consider the model behavior using some simple scaling arguments. This can be done for the nondimensional system by taking advantage of the fact that the parameter \( \varepsilon \) is small. First nondimensional scales are defined as follows:

\[
\begin{align*}
  u_a & \sim U_a \\
  v_a & \sim V_a \\
  p & \sim P \\
  Q_1 & \sim Q \\
  \partial/\partial y & \sim 1/Y_1 \\
  \partial/\partial x & \sim 1/X \\
  y & \sim Y
\end{align*}
\]

System [2.2.3] sets no intrinsic scale for \( x \), so the scale of the local response determined by the forcing. There are, however, two scales in \( y \): one set by the forcing (\( Y_1 \)), and another determined by the distance from the equator (\( Y \)). The procedure is to express \( X \), \( Y \) and \( Y_1 \) in terms
of \varepsilon in accordance with the scale and position of the forcing to be examined. Then, a dominant balance of terms in equations [2.2.3] gives the relative magnitude of the dependent variables in terms of the forcing (Q), and \varepsilon. Three examples will be discussed explicitly.

First, consider the local response to a heating which has a large \(x\)-scale, and a \(y\)-scale comparable to the equatorial deformation radius (which has a value of unity in the nondimensional system), and is situated off the equator. Thus it is appropriate to choose \(X = O(1/\varepsilon)\), \(Y = O(1)\), \(Y_1 = O(1)\). The terms in equations [2.2.3] then have the following magnitudes:

\[
\varepsilon U_a, \quad V_a, \quad \varepsilon P \quad [2.3.1a] \\
\varepsilon V_a, \quad U_a, \quad P \quad [2.3.1b] \\
\varepsilon P, \quad \varepsilon U_a, \quad V_a, \quad Q. \quad [2.3.1c]
\]

Starting with [2.3.1a], there are four possibilities: (i) \(\varepsilon U_a \sim V_a\) and \(\varepsilon P\) is smaller, (ii) \(\varepsilon U_a \sim \varepsilon P\) and \(V_a\) is smaller, (iii) \(V_a \sim \varepsilon P\) and \(\varepsilon U_a\) is smaller, and (iv) \(\varepsilon U_a \sim V_a \sim \varepsilon P\). In case (i), since it is assumed that \(U_a \sim V_a/\varepsilon\), the term \(\varepsilon V_a\) cannot balance \(U_a\) in [2.3.1b], so it follows that \(U_a \sim P\). But this is inconsistent with the assumption in (i) that \(\varepsilon P\) is smaller than \(\varepsilon U_a\). In a similar fashion one examines each of the four possibilities. The result in this case is that (ii) and (iv) lead to no inconsistencies. In making conclusions one must allow each variable to have the largest magnitude permitted by the various possible balances. In this case, then, the choice is (iv), and the balance is \(\varepsilon P \sim \varepsilon U_a \sim V_a \sim Q\). Thus the local response will have meridional winds which are small compared with the largest zonal winds. Of course
this ratio will depend on the value of $\varepsilon$. It can also be anticipated
that the zonal winds will be sensitive to changes in $\varepsilon$, whereas the
meridional winds will not. Figure 2.1 shows the model wind field for
a heating such as that just considered, and for two different values
of $\varepsilon$. The local response is as expected from the scaling considerations.

Consider next a heating which has large scale in $x$ and $y$, and
is centered on the equator. Looking first at the region away from
the equator, it is appropriate to set $X = O(\varepsilon^{-1})$, $Y_1 = O(\varepsilon^{-1})$, $Y = O(1)$. The
same procedure as above yields the balance $U_a \sim V_a \sim Q \sim \varepsilon P$. The
expectation is for small meridional and zonal winds away from the equator,
with little sensitivity to changes in $\varepsilon$. On the equator the situation
is different, however. Here $Y$ and $1/Y_1$ vanish (for $Q$ symmetric about
the equator), and the magnitudes of the terms are then as follows:

$$
\varepsilon U_a, \varepsilon P \quad [2.3.2a]
$$

$$
\varepsilon V_a \quad [2.3.2b]
$$

$$
\varepsilon P, \varepsilon U_a, \varepsilon V_a, Q \quad [2.3.2c]
$$

Since there is nothing to balance $\varepsilon V_a$ in $[2.3.2b]$ the meridional wind
must vanish at the equator. The remaining balance is $\varepsilon U_a \sim \varepsilon P \sim Q$.
Thus, the zonal winds on the equator are sensitive to changes in $\varepsilon$,
and are considerably larger in magnitude than the winds off the equator.
Figure 2.2 shows the computed wind field for an equatorially centered
large-scale heat source. In accordance with the analysis, the winds
are large and zonal near the equator, and $U_a \sim V_a$ away from the equator.

Consider finally a heating which is smaller scale in $x$ but of
large scale in $y$, and centered on the equator. For this we set $X = O(1)$,
Fig. 2.1 Prescribed heating and resultant model winds, for (a) $f=0.1$, and (b) $f=0.2$.

Fig. 2.2 As in Fig. 2.1.
\( Y_1 = O(1/\varepsilon) \). Away from the equator \( Y \gtrsim O(1) \), and the consistent balance gives \( V_a \sim P \sim \varepsilon U_a \sim Q \). The winds become more meridional than zonal, and the meridional winds are insensitive to changes in \( \varepsilon \). At the equator, \( U_a \sim Q \sim P/\varepsilon \), and thus the zonal winds are insensitive to changes in \( \varepsilon \). Such a situation is depicted in Fig. 2.3.

Many such calculations have been done, and the results may be summarized as follows.

1. The ratio of zonal to meridional component of the local wind response varies as the ratio of zonal to meridional scale of the heating, for regions away from the equator.

2. In regions away from the equator, the sensitivity of the local wind response to \( \varepsilon \) varies markedly with scale, small scales are comparatively insensitive to \( \varepsilon \). The sensitivity in zonal component relative to meridional component is again proportional to the ratio of zonal to meridional scale of the heating.

3. For heating which extends across the equator, the balance near the equator is different than that away from the equator. If the heating is locally symmetric about the equator, then the equatorial wind response is zonal. If the heating is also large scale in \( x \) the sensitivity to changes in \( \varepsilon \) is high.

4. Convergence approximately balances heating when the horizontal scales are small, whereas for large scales the dissipation term \( \varepsilon p \) is also important.

5. For heating fields with zonal scale larger than the equatorial deformation radius, the effect of including a mean easterly
Fig. 2.3 As in Fig. 2.1
wind field $\bar{u}$ of magnitude 10m/sec or less is negligible. For smaller scales the effect is significant (this result only holds for the assumed vertical structure, and not for small equivalent depths).

All of the above applies only to the local wind response. The response outside the heating region is discussed by Gill (1980). The winds decay to the east and west of the source, with a scale set by the dissipation parameter, $\varepsilon$. The decay scale is larger to the east of the source than to the west. This is attributed to the difference in wave propagation speed between Kelvin waves (which propagate eastward) and long Rossby waves (which propagate westward). The meridional scale of the remote response is the same as that of the local response.

2.4 Atmosphere Model II (with convergence feedback)

Equation [2.2.5] assumes that there is a one-to-one correspondence between the anomalous flux of latent heat at the ocean surface and the vertically integrated heating in the atmosphere above. In fact, such a one-to-one correspondence does not hold in general. A large part of the latent heat released in the active convection regions is observed to arise from low level moisture convergence (see, e.g., Newell et al. (1974), Ramage (1977), Cornejo-Garrido and Stone (1977)). A similar result is found in GCM models when forced by prescribed SST anomalies (Keshavamurty (1984), Shukla and Wallace (1983)). Webster (1981) attempted to address this in the context of a simple model. His approach was to use an iterative procedure in which the diabatic heating for each iteration depends on the low-level convergence from the previous iteration, with the initial diabatic heating specified.
For the tropical cases the result is that the final heating is larger in magnitude and smaller in scale than the initial prescribed heating. The procedure is efficient and seems to approximate the moisture convergence effect which is known to be important. One problem, however, is that the mean low level wind field is ignored. A heating anomaly should be expected to grow only if the total wind field is convergent, since only then can there be an anomalous influx of moisture. Consider, for example, an initial heating anomaly which occurs in a region of mean divergence. The heating gives rise to an anomalous wind field which is convergent. However, if the mean divergence is strong enough, the total field remains divergent, and there is no source of moisture to allow additional heating. In contrast, consider an initial negative heating anomaly in a region of mean convergence. The induced anomalous flow is divergent, but if the mean convergence is strong enough, the total remains convergent. The reduction in convergence implies less influx of moisture and less total heating, or equivalently, a negative heating anomaly. Thus the initial anomaly is enhanced. If the mean wind field is ignored, the result is that positive heating anomalies are always enhanced, and negative heating anomalies are never enhanced. It is likely that the modulating effect of the mean convergence is a first-order effect on the feedback process, so it is included in the present scheme.

The feedback is incorporated into the model with an iterative procedure similar to Webster (1981), in which the heating at iteration $n$ is determined by the relation
\[
Q_{n+1} = \begin{cases} 
Q_0, & \text{if } (\delta_M + \delta_n') > 0, \delta_M > 0 \\
Q_0 - \beta(\delta_M + \delta_n'), & \text{if } (\delta_M + \delta_n') \leq 0, \delta_M > 0 \\
Q_0 + \beta(\delta_M), & \text{if } (\delta_M + \delta_n') > 0, \delta_M \leq 0 \\
Q_0 - \beta(\delta_n'), & \text{if } (\delta_M + \delta_n') \leq 0, \delta_M \leq 0.
\end{cases}
\] [2.4.1a]

where \(Q_0\) is the initial prescribed heating, \(\delta_M\) is the mean divergence, and \(\delta_n'\) is the anomalous divergence at iteration \(n\) (which is determined by solving equations [2.2.3] for the heating at iteration \(n\) and then computing the divergence).

According to [2.4.1], if the total wind field is divergent and the mean wind field is divergent (case a), there is no feedback, regardless of the sign or magnitude of the anomalous heating. This is because it is assumed that when the wind field is divergent there is no convection to release latent heat. Therefore if both the mean wind and the total wind are divergent, there is no latent heating in either case, and thus no anomaly in latent heating. If the mean wind field is divergent, but the anomalous convergence is large enough to make the total wind field convergent (case b), enhancement occurs in proportion to the amount of total convergence (which is less than the anomalous convergence in this case). If the mean wind field is convergent, and there is anomalous divergence which is large enough to make the total wind field divergent (case c), enhancement occurs only in proportion to the strength of the mean convergence (which is less than the strength of the anomalous divergence in this case). Finally, if the mean wind field is convergent and the total wind field is also convergent (case d), then feedback occurs in proportion to the full strength of the anomalous convergence/
occurs in proportion to the full strength of the anomalous convergence/divergence.

Usually the strength of the mean convergence/divergence is large compared with anomalies so that the cases of no feedback (a) and full strength feedback (d) apply. In regions where the mean convergence/divergence is small, however, the other two cases arise and must be treated consistently in order for the iteration procedure to converge.

The parameter $\beta$ may be considered an efficiency factor for the CISK-like feedback process. If $\beta = 1$, then the efficiency is 100%, and in fact the process is unstable, as will be shown shortly. For $\beta < 1$ the process is stable. In Appendix A some dimensional arguments are made which suggest independently that $\beta < 1$ is appropriate for the atmosphere.

The scheme is only a partial description of the real moisture feedback process. It implicitly assumes that low level convergence of moisture is proportional to the convergence of mass: i.e., that the specific humidity of the near-surface air is constant over the ocean. This assumption is not as bad as it may seem because the feedback process is primarily operative in the mean convergence regions, and those regions tend to coincide with nearly constant (and high) specific humidity and warmest SST (see Newell et al. (1974)). A more serious problem is the neglect of moisture advection. In the present scheme, feedback can occur only in response to a local forcing. The possibility does not exist that moisture could be advected into a region of convergence even if there were no initial heating anomaly there. The fact that the largest rainfall anomalies during ENSO occur in the central Pacific,
and the largest SST anomalies occur much further to the east suggests that moisture advection is important. To model this properly would require keeping a detailed moisture budget, and this has not been attempted with the present, otherwise simple model. One may anticipate, however, that this model will appear to overestimate the local wind response in some areas while underestimating it in others. This is the case, as will be shown subsequently. At the same time, including only the convergence feedback results in a considerable improvement over the strictly SST-forced parameterization.

It is possible to gain some insight into the model feedback by considering the case of an f-plane centered on the equator, i.e., the nonrotating case. For small-scale motions (smaller than the equatorial deformation radius) near the equator this would be a reasonable approximation to the full equations. It is possible to show either by computation or by the analysis of the previous section that the feedback is most efficient on the equator. Thus the results concerning amplification and convergence of the scheme for this case can be considered global bounds for the system. (Also, the following considers the case of maximal feedback, i.e., [2.4.1d].)

The system under study for \( f = 0 \) is

\[
\begin{align*}
\mathbf{u}_a^i &= -p_x^i & [2.4.2a] \\
\mathbf{v}_a^i &= -p_y^i & [2.4.2b] \\
\mathbf{e}^i + (u_a^i)_x + (v_a^i)_y &= -\alpha Q_o - \beta((u_a^{i-1})_x + (v_a^{i-1})_y), [2.4.2c]
\end{align*}
\]

where the superscript \( i \) refers to the \( i \)th iteration, and \( \alpha Q_o \) is the
prescribed initial heating. Equations [2.4.2] can be combined to form
the single equation

\[- \varphi^2 p^i + \varepsilon^2 p^i = -\alpha a Q_o - \beta \varepsilon^2 p^{i-1}.\]  \hspace{1cm} [2.4.3]

Now consider the Fourier modes which have the form

\[Q_o = Q e^{i(kx + ly)}, \quad p^i = p_o^i e^{i(kx + ly)}.\]

Substituting into [2.4.3] yields

\[
p_o^i = \frac{-\alpha a Q + \beta (k^2 + l^2) p_o^{i-1}}{k^2 + l^2 + \varepsilon^2} \hspace{1cm} [2.4.4]
\]

If the initial heating is not maintained (i.e., a set to zero), then
the feedback from then onward would give

\[
p_o^i = \frac{\beta (k^2 + l^2)}{k^2 + l^2 + \varepsilon^2} p_o^{i-1} = a p_o^{i-1} \hspace{1cm} [2.4.5]
\]

In [2.4.5], the coefficient 'a' varies between 0 and \(\beta\) over the whole
range of wavenumbers. A sequence such as [2.4.5] diverges if the ratio
of successive elements is larger than 1, and converges to zero if the
ratio is less than 1. It follows that if \(\beta > 1\) the sequence diverges
for at least some wavenumbers, and if \(\beta < 1\) the sequence approaches
0 for all wavenumbers. A finite nonzero solution therefore requires
\(a \neq 0\): in other words, an imposed forcing which is maintained for all
iterations.

The sequence [2.4.4] can be written in the following form:
\{p_0, p_0[1+a], p_0 + a(1+a), p_0[1+a+a^2], p_0[1+a+a^2+a^3], \ldots \} \quad [2.4.6]

where

\[ a = \beta(k^2 + 1^2)/(k^2 + 1^2 + \varepsilon^2) \quad [2.4.7] \]

and

\[ p_0 = -\varepsilon\alpha Q/(k^2 + 1^2 + \varepsilon^2). \quad [2.4.8] \]

For \( a < 1 \), the sequence converges to

\[ p_\infty = p_0/(1-a) = -\varepsilon\alpha Q/[(\varepsilon^2 + (1-\beta)(k^2 + 1^2)]. \quad [2.4.9] \]

For \( a > 1 \) the sequence clearly diverges. Thus, if \( \beta > 1 \), the procedure does not converge. For \( \beta < 1 \), it converges and gives an amplification of the pressure anomaly which is largest for small wavenumbers. The final value increases with increasing \( \beta \), and decreases with increasing \( \varepsilon \). Using [2.4.2a] and [2.4.2b] the following relations can be derived for the final amplitudes of the Fourier components of \( u_\phi \) and \( \delta \) (divergence):

\[ |U_\omega| = k|Q|/[(\varepsilon^2 + (1-\beta)(k^2 + 1^2)] \quad [2.4.10] \]

\[ \delta_\omega = (k^2 + 1^2)Q/[(\varepsilon^2 + (1-\beta)(k^2 + 1^2)]. \quad [2.4.11] \]

From [2.4.10] it is seen that the final zonal wind anomaly approaches zero for both \( k \to 0 \) and \( (k,1) \to \infty \). It is maximized for an intermediate value of \( k \). The divergence anomaly, on the other hand, amplifies most for small scales, with a maximum value \( \sim Q/(1-\beta) \). Therefore if \( \beta \) has a value close to unity, the final convergence anomaly can be very
large for a small-scale forcing. This result will be important in understanding the behavior of the coupled atmosphere-ocean models of Chapter IV.

Some of the above results are illustrated in Figures 2.4 and 2.5. Figures 2.4a and 2.4b show the initial wind and divergence fields, respectively, for a prescribed large-scale positive heating anomaly of the form

\[ Q_o = \begin{cases} 
\cos(\pi x/2L_x) \cos(\pi y/2L_y) & |x-x_o| \leq L_x, |y-y_o| \leq L_y \\
0, & \text{otherwise}
\end{cases} \]  

[2.4.12]

with \( L_x = 4 \), \( L_y = 4 \), and with zero mean divergence (recall the dimensional scale for \( x \) and \( y \) is about \( 10^0 \) latitude). The atmosphere parameters \( \beta \) and \( \epsilon \) have the values .75 and .3, respectively. Figures 2.4c and 2.4d show the same fields after the iteration procedure has converged.

The convergence amplifies by a factor of two, and decreases in meridional scale, especially to the west of the heating maximum. The wind field amplifies to a lesser extent, with the largest enhancement at the equator to either side of the heating maximum.

In figure 2.5a and 2.5b are shown the same sequence, except for a small-scale initial forcing: specifically for \( L_x = L_y = 1 \) in [2.4.12]. In this case there is a much closer balance between heating and convergence, with the initial convergence maximum 91% of the initial heating amplitude, as opposed to 57% in the previous case. The amplification in convergence is also larger than before, being more than a factor of three in this case. Whereas the convergence magnitudes are larger for the small scale forcing, the wind magnitudes are much smaller, as expected from [2.4.10] and [2.4.11]. However, the amplification of the wind field
Fig. 2.4 (a) initial wind field, and (b) initial divergence field, for heating of the form $Q = \cos(\pi x/8)\cos(\pi y/8)$. (c) wind field, and (d) divergence field, after convergence feedback. Mean divergence field is assumed to be zero.
Fig. 2.5 As in Fig. 2.4, except for a small-scale heating: $Q = \cos(\pi x/2)\cos(\pi y/2)$. 
is larger for the small scale forcing, owing to the larger increase in convergence (and thus, heating).

The calculations so far are analogous to those of Webster (1981) because the mean divergence was prescribed to be zero. In the next case, illustrated in Figures 2.6a and 2.6b, an idealized nonzero mean divergence is considered. The prescribed initial forcing is the same as in Figure 2.4, but the mean divergence field has the following form:

\[
\delta_M = \begin{cases} 
-6.0, & 0.5 < y < 1.5 \\
+2.0, & y \leq 0.5, \ y > 1.5. 
\end{cases} \tag{2.4.13}
\]

This divergence field corresponds to a mean ITCZ between 5°N and 15°N, with divergence to the north and south (a nondimensional value of 1.0 for \(\delta_M\) corresponds to a dimensional value of \(10^{-6}\) sec\(^{-1}\)). Again, \(\beta = 0.75\) and \(\varepsilon = 0.3\). The initial wind and divergence fields are the same as Figures 2.4a and 2.4b. The final wind and divergence fields are shown in Figures 2.6a and 2.6b. Comparing with 2.4c and 2.4d shows the important influence of the mean field. In the convergence zone, amplification occurs, and to a greater extent than before because the mean convergence imposes a smaller scale on the heating field. The divergence anomalies amplify much more than before in the mean convergence zone. This is because a negative heating anomaly is allowed to develop in these regions as long as the total flow remains convergent. Without the mean convergence, negative heating anomalies do not develop, and the divergence anomalies arise only as a result of the remote response to the positive heating anomaly.
Fig. 2.6 (a) wind field, and (b) divergence field, after convergence feedback, with prescribed mean convergence between $y=0.5$ and $y=1.5$, and divergence elsewhere (see text). Initial heating, wind, and divergence are the same as for Fig. 2.4.

Fig. 2.7 (a) wind field, and (b) divergence field, after convergence feedback, with prescribed mean convergence for $x<0$, and divergence for $x>0$ (see text). Initial heating, wind, and divergence are the same as for Fig. 2.4.
The largest wind anomalies also are found in the mean convergence zone, rather than on the equator as before, and the flow has a northward component across the equator in the heating region. The positioning of the mean convergence off the equator produces a noticeably asymmetric response to a symmetric initial forcing.

Figure 2.7 shows the result for the same forcing but for a mean divergence field which varies only in \( x \):

\[
\delta_M = \begin{cases} 
-2, & x < 0 \\
+2, & x > 0 
\end{cases} \quad [2.4.14]
\]

The divergence/convergence boundary is therefore at the central longitude of the forcing. The initial wind and divergence are the same as before, (Figs. 2.4a and 2.4b), and the final fields are shown in Figures 2.7a and 2.7b. Again, amplification is greater in the mean convergence zone as a result of scale reduction and the growth of negative heating anomalies. In the eastern half of the initial forcing region, which has mean divergence, the final heating/convergence is less than before, and thus the associated remotely induced divergence further east is reduced.

Despite these differences, the final wind fields are qualitatively similar. This is because the symmetry about the equator is preserved in this case, and for the most part, the effect of the mean convergence is only to modulate the relative amplitude of the final wind anomalies.
2.5 Computational Procedures

The system [2.2.3] is solved by using spectral decomposition in x, and finite differencing in y. Writing \((u_a, v_a, p, Q) = e^{ikx}(U_a, V_a, P, \tilde{Q})\), and substituting into [2.2.3] gives

\[
\begin{align*}
\varepsilon U_a - yV_a/2 &= -ikP \\
\varepsilon V_a + yU_a/2 &= -P_y \\
\varepsilon P + ikU_a + (V_a)_y &= -\tilde{Q}.
\end{align*}
\]

These equations may be combined to give

\[
(V_a)_{yy} + (-y^2/4 + ik/2\varepsilon - \varepsilon^2 - k^2)V_a = -\tilde{Q}_y + (ik/2\varepsilon) \tilde{Q} \quad [2.5.2]
\]

The boundary conditions for the problem are that \(V = 0\) at the poles. In practice the solution decays rapidly outside the forcing region (which is tropical) so the solution is insensitive to the exact location of the meridional walls. Equation [2.5.2] with boundary conditions is solved for each \(k\) by using a second order finite-difference representation, and solving the resulting system of linear equations. That is, writing

\[
(V_a)_{yy} = \frac{V_{a}^{n+1} + V_{a}^{n-1} - 2V_{a}^{n}}{\Delta y^2}, \quad Q_y = \frac{Q^{n+1} - Q^{n-1}}{2\Delta y},
\]

and substituting into [2.5.2] gives a system of equations for the \(\{V_{a}^{n}\}\) of the form
\[ V_{a}^{n+1} = \left( \frac{i n \Delta y^{2}}{2} y_{n}^{2} \right) V_{a}^{n} + V_{a}^{n+1} = \frac{-\Delta y}{2} \tilde{Q}_{a}^{n+1} + \frac{i n \Delta y^{2}}{2} \tilde{Q}_{a}^{n+1} \]
\[ n = 2, N-1 \]

\[ \left( \frac{i n \Delta y^{2}}{2} y_{1}^{2} \right) V_{a}^{1} + V_{a}^{2} = \frac{i n \Delta y^{2}}{2} y_{1} \tilde{Q}_{a}^{1} + \frac{\Delta y}{2} \tilde{Q}_{a}^{2} \]

\[ \left( \frac{i n \Delta y^{2}}{2} y_{N}^{2} \right) V_{a}^{N} + V_{a}^{N+1} = \frac{\Delta y}{2} \tilde{Q}_{a}^{N-1} + \frac{i n \Delta y^{2}}{2} \tilde{Q}_{a}^{N} \]

In [2.5.3], \( y_{1} = y_{\text{south}} + \Delta y \), \( y_{N} = y_{\text{north}} - \Delta y \), where \( y_{\text{south}} \) and \( y_{\text{north}} \) are the boundaries of the domain, \( \Delta y \) is the grid spacing, and \( N \) is the total number of interior grid points.

In these calculations \( y_{\text{south}} = -8.0 \, (80^\circ \text{S}) \), \( y_{\text{north}} = +8.0 \, (80^\circ \text{N}) \), and \( \Delta y = 0.2 \, (\sim 200 \text{ km}) \). The spectral decomposition is done using a 64-point FFT, and the domain is the full 360\(^\circ\) longitude, giving a spatial resolution (in \( x \)) of about 600 km.

Once the \( \{V_{a}^{n}\} \) are found, the \( \{U_{a}^{n}\} \) are obtained using

\[ (k^2 + s^2) U_{a}^{n} = \frac{\Delta y}{2} V_{a}^{n+1} - \frac{\Delta y}{2} V_{a}^{n-1} + i k \tilde{Q}_{a}^{n} \quad n = 2, N-1 \]

\[ (k^2 + s^2) U_{1} = \frac{\Delta y}{2} V_{1}^{1} - \frac{\Delta y}{2} V_{1}^{1} + i k \tilde{Q}_{1}^{1} \]

\[ (k^2 + s^2) U_{N} = \frac{\Delta y}{2} V_{N}^{N} - \frac{\Delta y}{2} V_{N}^{N} + i k \tilde{Q}_{N}^{N} \]

which may be derived from combining [2.5.1a] and [2.5.1c], taking finite differences and applying the boundary conditions.
The systems \(2.4.3\) and \(2.4.4\) must be solved for each \(k\). The result is the solution in terms of the spectral coefficients of \(u_a\) and \(v_a\), from which the actual wind field may be obtained by inverse transforming.

In the calculations with convergence feedback, a system of the form of \(2.5.1\) must be solved for each iteration of the scheme, where the heating at each iteration is obtained using \(2.4.1\). The iteration proceeds until the difference between successive divergence fields is everywhere less than a specified value (typically 0.15 was used). The mean divergence fields must be specified in \(2.4.1\): they are obtained using the long-term monthly mean wind fields from the Rasmussen and Carpenter (1982) data set. The SST anomalies which are used to compute the initial heating anomaly according to equation \(2.2.5\) are obtained from the same data set. Linear interpolation is used to compute the values at the model gridpoints.

2.6 Simulation Experiments

The typical ENSO warm event as described by the composite data analysis of Rasmussen and Carpenter (1982) (hereafter called RC) has a duration of somewhat more than a year. Figure 2.8, reproduced from their paper, shows a composite index for sea surface temperature anomalies near the South American coast during a three year period encompassing the warm event. Indicated by arrows beneath the figure are six times for which model simulations will be presented. They capture the primary features of the full life cycle. Adopting the notation of RC, in which year 0 is defined to be that of the coastal warming, the six reference
Fig. 2.8 El Niño composite of eastern Pacific SST anomalies (after Rasmussen and Carpenter, 1982). Arrows indicate the times for which model results will be presented.
times are August (-1), December (-1), May (0), December (0), April (1), and August (1).

In all of the simulations to be shown here, the model parameter values are as follows. For model I (no feedback): \( \alpha = 1.6, \varepsilon = 0.3 \) (see equations [2.2.3] and [2.2.5]). For Model II (with feedback), \( \alpha = 1.6, \varepsilon = 0.3, \beta = 0.75 \) (see equations [2.2.3] and [2.4.1]). These choices are somewhat arbitrary. From the physical arguments in Appendix A, \( \alpha \) and \( \beta \) are expected to be 0(1); beyond that a subjective choice was made based on the results of a large number of cases. The model results depend almost linearly on \( \alpha \), so the sensitivity is not high in this case. The sensitivity of the divergence field to \( \beta \) is very high, but the sensitivity of the wind field itself is much less, as seen from equations [2.4.10] and [2.4.11] (especially for large-scale initial forcing). In practice, increasing \( \beta \) by, say, 10% has a greater effect on computation time than on the final wind field; decreasing \( \beta \) significantly (a factor of 2 or more) gives a result not appreciably different from the model without feedback.

The results are very sensitive to the value of \( \varepsilon \), the friction parameter. This is expected from the analysis of section 2.3. As pointed out there, for equatorial forcing of large zonal scale, the local response is sensitive to \( \varepsilon \). However, the response some distance (in \( x \)) away from the forcing is even more sensitive to \( \varepsilon \), since the solution decays as \( \exp (-n \varepsilon |x - x_o|) \) away from a source at \( x_o \), where \( n = 1 \) for the region to the east of \( x_o \), and \( n \geq 3 \) for the region to the west of \( x_o \) (see Gill, 1980). Mostly because of the magnitude of the model response to the east of the forcing, better agreement with observations is obtained using a large value of \( \varepsilon \). A value \( \varepsilon = 0.3 \)
corresponds to a dimensional decay time of about 1 day. Although this seems to be exceedingly frictional, it is not inconsistent with what Holton and Colton (1972) deduced from their diagnostic study of the tropical atmosphere.

a. August (-1): Pre-event Phase (Figs. 2.9)

This period is characterized by anomalously cool SST in the central and eastern equatorial regions, and anomalously warm SST in the far western equatorial region. Positive SST anomalies also are found on the southern flank of the normal position of the South Pacific Convergence Zone (SPCZ). (For reference, seasonal maps of SST, surface wind, and surface convergence generated from this data set are presented and briefly discussed in Appendix B.) Figure 2.9a shows the composite SST anomaly field for August (-1). Figures 2.9b and 2.9c show the composite wind anomaly field and divergence field for the same time. There are easterly wind anomalies in the western equatorial zone, with poleward flow across the mean positions of the SPCZ in the central Pacific. Correspondingly, there is anomalous divergence between about 10°N and 10°S-15°S in the central Pacific, and convergence further poleward. Figs. 2.9d and 2.9e show the model wind and convergence fields using Model I (no feedback), and Figures 2.9f and 2.9g show the corresponding fields using Model II (with convergence feedback). In all cases the calculated model heating is very similar to the convergence fields which are shown. The wind fields using Model I and Model II are qualitatively similar, with larger amplitude in the regions of mean convergence in the latter case. Both have easterlies in the same region as does the composite, northerly anomalies across the SPCZ region.
Fig. 2.9 (a) composite SST anomalies, (b) composite wind anomalies, and (c) composite divergence anomalies, for AUG(-1).
Fig. 2.9 Cont'd. (d) predicted wind anomalies, and (e) predicted divergence anomalies, for the same time, using Model I. (f) predicted wind anomalies, and (g) predicted divergence anomalies, for the same time, using Model II (see text). Scale for wind vectors and contours for divergence are the same as for the composites.
and little response in the eastern regions. Looking at the convergence fields, the situation is very different, with much closer agreement with the observations in the case of model II. This is the case in general, and suggests that the low-level convergence field is indeed important in determining the atmospheric heating. A detailed comparison between model II convergence and composite convergence reveals a number of differences. One must be cautious, however, in making such a comparison. First, the model convergence field is highly sensitive to the parameter values, as was shown earlier. Also, the details of the computed divergence field for the composited data should be viewed as highly uncertain due to the quality of the data and the averaging and compositing procedures (see RC for a discussion of this). This is also true for the wind fields, but to a lesser degree. For both the wind fields and the divergence fields, it is only sizeable, spatially coherent features that should be considered meaningful for purposes of model comparisons. In this light, the major discrepancy between the model II divergence field and the composite field is the enhanced central Pacific ITCZ which is evident in the composite, and absent in the model results.

b. December (-1), Onset Phase (Figs. 2.10)

At this time the composites reveal an interesting development near the dateline (Figs. 2.10a and b): a small region of warm SST anomaly near the equator and to the south, together with westerly wind anomalies in the same region and to the west. The features appear to develop simultaneously to within the time resolution of the composites. This period has been termed onset because from this time until the
end of the ENSO event equatorial westerly wind anomalies and warm SST anomalies are continuously present. At this time there also appears to be a region of warm SST anomaly near 25°S and 90°W, in the region of the South Pacific High (SPH). The SPH itself appears to be weaker than normal, and the associated tradewinds in the eastern ocean between 10°S – 20°S are anomalously weak. Whether this development has any connection to the equatorial events at the same time or later is at best uncertain. There is also less reliability of the data in this region (see RC). Accordingly, the discussion here will concentrate only on the equatorial anomalies.

As shown by Figs. [2.10d–g], the model II results are in better agreement with the composites than the model I results, but even there the differences are considerable. The major feature of the model wind field is the region of easterly anomalies just to the east of the warm SST anomaly, a feature not found in the composite. The model westerly anomalies are too weak and displaced too far westward. Although in this case one may not judge the discrepancies to be serious, they point to a problem with the model which is much more general. Observations indicate that anomalously wet episodes in the central and eastern Pacific during ENSO are characterized by equatorial westerly wind anomalies in the region of precipitation anomaly, and little or no apparent easterly anomalies further east. For an isolated heat source spanning the equator, the model always produces a more symmetric wind response, with westerlies to the west and easterlies to the east. Part of the problem in this case is evident in the convergence fields. The composite shows that the convergence maxima do not lie over the SST anomaly maximum, but are displaced off the equator. Over part of the warm anomaly, the
Fig. 2.10 As in Fig. 2.9, except for DEC(-1).
flow is actually divergent. The model heating parameterization cannot produce such features since the initial heating field (similar in form to the Model I convergence) depends directly on the SST anomaly distribution. The feedback can only enhance the heating anomaly in the regions of mean convergence.

The results suggest that the model could be improved substantially by including moisture advection — i.e., to have a complete moisture budget. For example, looking at the total wind field for this time (not shown), it can be seen that the convergence maxima lie downstream of the SST anomaly maximum, which itself lies in a region of divergence. Only with a full moisture budget could this likely be reproduced. If it were reproduced, however, then the model heating would look different, with the maxima off the equator. In that case the resulting wind field would more closely resemble the composite as well (the equatorial easterly response to the east diminishes rapidly as the heat source is moved off the equator, whereas this is not true for the westerly response in the source region and to the west).

c. **May (0) Peak Phase** (Figs. 2.11)

Perhaps the most striking oceanic event during ENSO is the rapid and large rise in SST in the eastern Pacific, especially near the South American coast. In the typical event the rise starts in early spring, and SST peaks in May or June. The rise follows by a few months the appearance of sustained westerly wind anomalies in the western and central equatorial Pacific. Figs. 2.11a–c show the composite fields for May (0), Figs. 2.11d–e the Model I simulations, and Figs. 2.11f–g the model II simulations. There is still a local maximum of SST anomaly
near the dateline which now extends somewhat further eastward than before, as well as the larger anomalies extending westward along the equator from the coast. Thus the whole equatorial ocean east of 160E is warmer than normal by this time. Westerly wind anomalies are still present near the dateline and to the west. There is almost no zonal wind anomaly in the eastern regions where the SST anomaly is largest. On the other hand, meridional wind anomalies are large in the vicinity of the ITCZ at this time, and are consistent with the ITCZ being displaced southward of its normal position. Poleward of 10°N and 1°S, the eastern Pacific wind anomalies are small. There is anomalous convergence near the equator in the entire longitude range of warm SST anomalies, with divergence to the north and south. The region of convergence is noticeably smaller in meridional extent than that of the SST anomalies, suggesting that the anomalous heating is so as well.

The character of the model I and model II results is clearly the same. What makes the model II results more comparable to the composites is the increased relative amplitude of the response near the dateline and in the ITCZ region. The maximum westerly anomalies are again displaced to the west of those in the composite. The model wind field near the equator in the east compares favorably with the observations. Away from the equator they differ qualitatively, the model giving substantial easterly anomalies as opposed to the observed weak westerly anomalies. As will be seen, this problem persists and even worsens during the mature phases of the event.

The model convergence is not as equatorially confined as that of the composite, and the divergence anomalies are weaker. Also the longitudes of maximum convergence differ. However, the two fields
basically agree in that they show near-equatorial convergence eastward of 160°E. The differences in the meridional scale of the convergence again can be traced to the model heating parameterization which, for example, gives considerable heating at (5°S, 100°W) where apparently there should be none.

d. December (O) Mature Phase (Figs. 2.12)

Following the coastal peak in SST in May–June, the warm anomalies spread westward so that the central Pacific SST anomaly grows monotonically until December. After June the coastal SST anomaly decreases until September, at which time the maximum SST anomaly is located off the coast at 100°W. Thereafter the anomalies grow everywhere, but the maximum remains off the coast. This growth follows a significant collapse of the equatorial tradewinds in the central Pacific which occurs in late (N. Hem.) summer, and is sustained until December. This feature is seen in the composite wind anomalies of Fig. 2.12b. The westerlies are bounded to the north and south by sizeable equatorward winds, reflecting an equatorward displacement of the ITCZ and SPCZ. East of 130°W the anomalous winds are small and directed northward across the equator. Also at this time strong easterly anomalies appear in the western Pacific and are sustained thereafter for several months. The anomalous convergence is large and spans the equatorial central Pacific. It is more equatorially confined than the SST anomalies, and centered considerably further to the west.

The region of largest wind anomalies is very well reproduced by model II and less well so by model I, the difference between them being primarily in magnitude. In the far west there are discernible easterly
anomalies only with model II, and still they are very weak compared with the composite easterlies. In the east, model II gives weak southerlies across the equator as seen in the composite, and model I does not. With either model I or model II, the zonal winds off the equator in the east bear no resemblance to the observations. This discrepancy cannot be attributed solely to the heating parameterization, although this is again part of the problem. Much of the easterly flow in the model arises from the remote response to the large equatorial heating further to the west. The model response to the east of an isolated heat source, attributable to eastward propagating Kelvin waves (Gill, 1980), is characterized by easterly flow of relatively large meridional extent. The Kelvin wave component is most effectively excited by equatorially centered heating of relatively large meridional extent such as that found during the mature phase of ENSO. The composites show no sign of a significant Kelvin wave signal at any time, including the mature phase. It is thus during the mature phase that the disagreement is greatest. Very near the equator the problem is much less serious because the local heating there tends to induce westerly anomalies which partially offset the remotely induced easterly anomalies.

The large difference between the model I and model II convergence shows the degree to which the feedback process can alter the initial heating, both in magnitude and distribution. The effect of the mean convergence zones, especially the SPCZ and ITCZ, are evident. As usual, the model II convergence compares more favorably with the composite. However, while the model II convergence anomalies delineate clearly the normal positions of the mean convergence zones, the composite is more indicative of a movement of the zones. The result is a somewhat
Fig. 2.12 As in Fig. 2.9, except for DEC(0).
Fig. 2.12 Cont'd. As in Fig. 2.9, except for DEC(0).
different spatial distribution of the convergence anomaly maxima. Specifically, the composite has more equatorial convergence in the central Pacific, and less convergence at all latitudes in the far eastern region. In the eastern Pacific ITCZ region the composite convergence is centered a little closer to the equator, with divergence to the north that is stronger and also equatorward of that in the model. Similar differences are evident at other times as well, but are accentuated in the mature phase when the anomalous fields are largest.

There is a clear indication that some features of the anomalous fields involve sizable adjustments of the mean fields, and that these adjustments cannot be well described by the model. In other words, the anomalies associated with ENSO can be as large as the mean fields, so that only if the model can well describe the mean fields is it likely to reproduce accurately the details of the observed anomalies.

To summarize for the mature phase, it appears that the model underestimates the equatorial heating in central Pacific, and overestimates the eastern Pacific heating. The winds in the central Pacific, which comprise the largest equatorial anomalies, are well simulated by the model. In the eastern Pacific, the model winds exhibit a large remote response to the central Pacific heating which is absent in the observations. Near the equator in the east, the remote response in the model is offset by the effect of the local heating, and the agreement with observations is better.

e. April (1) Termination Phase (Figs. 2.13)

Shortly after December (0) both the eastern Pacific SST anomalies and the equatorial westerly wind anomalies begin to dissipate rapidly.
The equatorial easterly winds in the western Pacific persist and are still present in April (1), as seen in Figure 2.13b. In the central Pacific, starting in January (1), the equatorial maximum in wind anomalies moves south of the equator, and the meridional flow is southward across the equator. The central Pacific convergence anomalies likewise move south of the equator. By April (1) in the eastern Pacific cold SST anomalies have developed near the coast and to the west along the equator. Warm SST anomalies exist only in the central ocean now. The wind anomalies in the east are small, and there is anomalous divergence near the equator, and convergence to the north and south.

At this time it is difficult to say that model II performs better than model I. Neither compare well with the composites. The problems with the model that have already been described are particularly evident at this time. In the east, the easterlies are strong even on the equator since the local heating anomaly is negative now, and this enhances the remotely forced easterlies driven by the heating to the west. In model II the effect is more pronounced than in model I. In the central Pacific the wind anomalies are largest near the equator, rather than to the south as in the composite. However, there is southward flow across the equator, in agreement with the composite. In the western Pacific, neither model I nor model II give the large easterly anomalies seen in the composite. The model correctly gives convergence in the central ocean and divergence in the east, but both are centered too near the equator.
Fig. 2.13 As in Fig. 2.9, except for APR(1).
f. **August (+1) Post-Event Phase** (Figs. 2.14)

By this time cold SST anomalies have spread further westward than in April, extending now to 160°W as shown in Fig. 2.14a. A small region of warm SST anomaly remains near the dateline on the equator, and to the southeast. Another region of warm SST anomaly has developed to the north and east of Australia. The wind anomaly field shows that easterlies still prevail in the western Pacific. In the central and eastern regions, the only sizeable zonal wind anomalies are around 10°N to the east of 120°W: the meridional wind anomalies are poleward on either side of the equator. The flow is divergent between the equator and 10°N in the east, with convergence to the north and south. Another region of convergence is located near the equator at the dateline and to the west.

Overall, the model II results more resemble the composites. The model II wind field compares favorably with the composite in the SPCZ region, and in the far eastern regions. However, to the east of the warm SST anomalies the model again gives equatorial easterlies where none are present in the composite, and fails to reproduce the easterlies in the western Pacific. The model convergence field resembles the composite in that it has convergence near the dateline and divergence in the eastern half of the ocean north of the equator. The two fields differ in that the composite shows a zone of convergence just south of the equator in the east, whereas the model field has divergence extending to 10°S.
Fig. 2.14 Cont'd. As in Fig. 2.9, except for AUG(1).
2.7 Models v. Observations: Index Comparisons

An additional means for evaluating the model simulations is to compare time series of computed indices. The indices, rather than focusing on the detailed spatial structure of the fields, allow a comparison of the temporal aspects of the model results and the observations. Four indices are presented here — they are defined to coincide with indices currently used by observationalists. The first two, TW1 and TW2, measure anomalies in the equatorial tradewinds in the western Pacific and eastern central Pacific, respectively. The domain for TW1 is the region between $5^\circ$N and $5^\circ$S, and $135^\circ$E and $180^\circ$. TW2 covers the region between $5^\circ$N and $5^\circ$S and $180^\circ$ and $140^\circ$W. The third and fourth indices measure the pressure anomalies in the vicinity of Darwin ($12.4^\circ$S, $130.9^\circ$E), and Tahiti ($17.5^\circ$S, $149.6^\circ$W), the components of the conventional South Oscillation Index (SOI).

Figure 2.15a shows the index TW1 over the three-year period, calculated using (a) the composite data, (b) the model I wind fields, and (c) the model II wind fields. It is evident that the two model indices are essentially the same except that the model II index is larger in magnitude, making it more comparable to the composite index during the growth phase. Even with model II, however, the maxima of TW1 in January (0) and September (0) are underestimated considerably. The most conspicuous difference between either of the model indices and the composite index occurs in the period between November (0) and June (1). After peaking sharply in September (0), the composite index declines rapidly to near zero by the end of the year, and remains small until June (1). The model indices peak in October, but remain large well into the following year. This difference is due primarily to the failure
Fig. 2.15a TWI indices (see text), computed from composite data, Model I output, and Model II output. Time is indicated in years, from the beginning of year(-1) to the end of year(+1).

Fig. 2.15b As in Fig.2.15a, except for TW2 index (see text).
of the model to simulate the large easterly anomalies in the far western Pacific which are observed to develop in this period (see Figs. 2.13b and 2.14b). The index confirms the earlier conclusion that, particularly in the case of model II, the simulations and the composites compare favorably in the growth phase, but compare poorly in the termination phase.

Figure 2.15b shows that, further to the east, the overall comparisons are better. The relationship between the model I index and the model II index is not as simple as before, although the difference in magnitude, for the most part, is small. In this region the models overestimate the maximum anomaly, rather than underestimating it as they do in the west. The problems noted earlier with model easterly anomalies during the onset phase (December (-1)) and post-event phase (June (1) and following) are evident here as well, but apart from these times, the agreement is good.

The pressure indices are shown in Figs. 2.16a-c. Figure 2.16c is reproduced from RC (the data tape obtained does not include the pressure data), and shows the composite pressure for Darwin and Easter Islands, normalized by the variance. In Figure 2.16a are shown the nondimensional model pressure anomalies at the location corresponding to Darwin. The model II value is consistently larger than that of model I, reflecting the enhancement due to the feedback which is always operative in the model in this region. In both the model results and the composites, the tendency is for positive pressure departures to prevail during the ENSO year. Compared with the composite, however, the model pressure anomaly becomes positive 1-2 months early, and dissipates much too soon. It becomes negative in January (1), a full five months
Fig. 2.16a Pressure anomalies at the location corresponding to Darwin, Australia, computed from the Model I output and the Model II output.

Fig. 2.16b Model pressure anomalies for the location corresponding to Tahiti.

Fig. 2.16c Composite pressure anomalies (after Rasmussen and Carpenter, 1982).
before the composite does so. This possibly is related to the discrepancies in the near-equatorial winds during the same period.

Figure 2.16b shows the model pressure anomaly at the point corresponding to Tahiti. Unfortunately, RC do not show the composite pressure anomaly for Tahiti. According to their discussion, however, the tendency is for falling pressure late in year (-1), and continuing into year (0), peaking in the latter part of year (0), and then returning to near normal early in year (1). The model results show a similar trend, except that the negative maximum and the return to normal conditions occur later by about 2-3 months. In this case, the model I and model II results are almost the same because the feedback is not effective here, the region being one of mean divergence.

2.8 Discussion

The simulations have shown that the model can reproduce many of the salient features of the atmospheric observations for ENSO events — that is, sustained westerly wind anomalies and convergence anomalies in the equatorial central Pacific, equatorial flow across the usual positions of the mean convergence zones, higher than normal pressure in the western Pacific and lower than normal pressure in the eastern Pacific during the warm event year, and the opposite trend in each of these in the adjacent years. Other features are not well modeled — the appearance of strong easterly anomalies in the western Pacific during the mature phase, and the weakening of the Southeast Trades during most of the warm event. Still other features found in the model results are not found in the observations — most notably, easterly anomalies in the eastern Pacific during the mature phase and following.
The present model, even with the convergence feedback, is exceedingly simple in view of what is being simulated — it is not surprising that it does not exactly reproduce everything that is observed. In fact, one could ask why the model has even the limited success that has been demonstrated. The suggestion is made here that this limited success is somewhat fortuitous, resulting from the tendency for the effects of several model simplifications to offset each other, particularly in the equatorial zone during the growth phases of ENSO. At times and places where this is not the case, the discrepancies are significant.

One problem, suggested in many of the results, is the lack of a detailed moisture budget which includes evaporation, advection, and convergence of moisture. The problem is partially addressed by considering just the convergence effect (as in model II), but the remaining disagreement has often pointed to the remaining processes. The parameterization that has been used evidently overestimates the local heating that results from a SST anomaly in a region of mean divergence. This was suggested by the larger scale of the model response particularly in the eastern Pacific. It also could explain some of the discrepancies during the onset phase which were described earlier. During the peak and mature phases of the event, the composites suggested an equatorward movement of the mean convergence zones, sometimes resulting in a divergence anomaly over a region within the domain of the warm SST anomaly. Such a situation cannot be reproduced with the current heating parameterization, but could possibly be reproduced with a parameterization that does not dictate a one-to-one correspondence between local heating and SST anomalies.
Another problem of the model pertains to its predicted response to the east of an equatorial heating anomaly. The observed equatorial wind anomalies in the central and eastern Pacific are asymmetric in the sense that large westerly anomalies occur at times, but at no time do large easterly anomalies occur, regardless of the SST anomaly pattern. The model has no such asymmetry, and predicts easterly winds to the east of a heat source, decaying with a scale proportional to the friction parameter. (This is largely why the model results are more comparable to the observations for a large value of $\varepsilon$). An explanation of the peculiar asymmetry of the anomalies is suggested by looking at the total fields. Using measurements of outgoing longwave radiation (OLR) as an indicator of convective heating in the tropics, and comparing them with observed low level wind fields yields a striking relationship: in all regions of intense convective heating the winds are small in magnitude, and in all regions of very small heating, the winds are large (equatorward of 20°). In the dry zone spanning the equator in the central and eastern Pacific, for example, the equatorial tradewinds are generally strong. During an ENSO year, however, the active convection zone in the western Pacific expands eastward into the central Pacific (or even into the eastern central Pacific as in 1983). The total convective heating field then looks as usual, except shifted and/or expanded eastward. However, the above relationship holds even in this situation — large easterly winds exist in the dry zone which is now contracted eastward, and the winds are weak in the active convection zone which now includes the central Pacific. In terms of the wind anomaly fields, this means that, to the east of heating anomaly, in the region which is normally dry and is still dry now, the wind anomalies are small: but in the
region which is normally dry and now wet, the usual large easterly winds are replaced with weak winds, resulting in a large westerly anomaly (all of this can be seen easily in the data analysis of Arkin et al., 1983). Thus it seems that the problem of understanding the observed anomalies can be reduced to understanding this aspect of the total wind field — that large, nearly uniform low-level easterly winds prevail in the Pacific dry zones, and small, cyclonic, convergent low-level winds prevail in the regions of heavy precipitation.

Using the present model and a realistic distribution of total heating, this feature of the total wind field cannot be reproduced. However, the above discussion suggests a solution to the problem: the model friction should not be spatially uniform, but rather it should depend on the intensity of convective activity. The very large friction assumed by the current model should exist only in active convection zones, and elsewhere the flow should be relatively inviscid. Physically, such a prescription makes much sense. It has been argued that only the convective motions can conceivably provide the strong damping that is assumed in the model. The most consistent model formulation would therefore relate the effective dissipation to the intensity of convection. In the divergent, dry zones there is no mechanism to provide such dissipation, and the frictional decay time should approach what is more typically assumed for the atmosphere — perhaps 6-7 days. It is proposed here that with such a specification for dissipation the model could more closely approximate the observed mean low-level wind field, and therefore better approximate the observed ENSO anomalies.

If these speculations are correct, it means that it is not possible to reproduce the observed ENSO anomalies with a perturbation model,
because the frictional specification applies directly to the total fields, and not the anomalous fields. It also means that it is very important for the model to approximate closely the actual heating distribution, which, in turn, requires including a moisture budget as discussed earlier.

In the present model, the deficiencies in the specification of heating and dissipation tend to offset each other in the equatorial regions during the growth phases of ENSO. The reason for this is that the heating is overestimated in the eastern equatorial Pacific, tending to induce equatorial westerly anomalies, and the remote response to the east of the large central Pacific heating is overestimated, tending to give easterly anomalies. The result is weak equatorial wind anomalies in the eastern Pacific, as observed.

Thus, while having indicated specific ways to construct a better and more realistic model, the present study takes advantage of the ability of this very simple (and computationally efficient) model to approximate many of the important equatorial developments during ENSO. For reasons discussed in the following chapter, it is specifically these equatorial developments in the wind field that are most important in producing observed patterns of SST anomalies during ENSO.
CHAPTER III

Ocean Modeling of ENSO

3.1 Introduction

The results of Chapter II as well as those of numerous other investigations indicate that much of the observed atmospheric variability during ENSO can be related, directly or indirectly, to changes in sea surface temperature. What, then, causes the observed changes in SST? Little work has been done on this specific problem until very recently. Much more work has been done on understanding the dynamical response of the tropical ocean to ENSO-like anomalies in wind stress. To varying degrees the dynamical changes can be related to observed anomalies in surface currents, sea level, thermocline displacement, and sea surface temperature. The results of the dynamical studies suggest that many of the observed oceanic changes during ENSO owe their existence to equatorial wind stress anomalies.

Wyrtki (1975) was the first to suggest that fluctuations in the strength of the Southeast Trades spanning the equator in the Central Pacific were responsible for oceanic El Niño events. (The general association of El Niño with changing wind patterns was postulated earlier by Bjerknes, 1966.) Since the most dramatic changes in sea surface temperature occur in the far-eastern Pacific, it had earlier been assumed that El Niño resulted from a cessation of local coastal upwelling, which requires a weakening of the near-coastal winds. With a number of observations Wyrtki demonstrated that no such systematic changes occur in the near-coastal winds. He presented evidence that, prior to large events, the central Pacific equatorial winds were stronger

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than normal for perhaps a year or more, resulting in an accumulation of mass in the western Pacific. He also showed that with the occurrence of El Niño was a relaxation of the strong tradewinds, and he proposed that a large eastward flow of mass in the form of an equatorial Kelvin wave was responsible for the large thermocline displacements and sea surface temperature anomalies in the eastern ocean. The idea is that the sea surface temperature warms, not as a result of a reduction in upwelling, but because of the large inflow of warm water from the west which displaces the normally cool subsurface waters. Thus the theory related the observed eastern Pacific phenomena to atmospheric changes in the central Pacific, the first explanation of El Niño by remote forcing.

McCready (1976) and Hurlburt et al. (1976) were the first to demonstrate Wyrtki's theory with a simple ocean model. Using idealized wind stress forcing and a linear ocean model, they showed that large coastal signals in thermocline depth anomaly reminiscent of those observed during ENSO could be generated effectively by equatorial (zonal!) wind stress anomalies. In the models this response could be identified with eastward propagating, equatorially trapped Kelvin waves, and westward propagating long Rossby waves (plus coastal trapped Kelvin waves). These studies and several others that have followed (especially Busalacchi and O'Brien (1981) and Cane (1984)) show that simple dynamical ocean models can reproduce some features of the observed ENSO variability, based solely on prescribed ENSO-like variations in wind stress. They also demonstrate that stress anomalies poleward of about 5° contribute very little to the signal in the waveguide region along the equator and the South American coast, suggesting that only the winds in the equatorial band are important
for generating El Niño. Because of the success of these simple models, the relevance of linear equatorial waves to El Niño is now undisputed among oceanographers.

However, SST is not so well predicted by the dynamical models. The simplest relationship that is often assumed is that SST anomalies are linearly related to upper layer depth anomalies. This may be argued heuristically as follows. The tendency of upwelling is to cool the oceanic surface layer by transporting cooler waters from below. If the upper layer depth increases (a deepening of the thermocline), then the mean temperature gradient in the upper layer must be reduced, resulting in a decrease in temperature advection by upwelling, and, presumably, an increase in surface temperature. The temperature rise is due to the warming processes which are normally in balance with the cooling associated with upwelling. This reasoning ignores the fact that upwelling is confined to narrow regions near the equator and near the coasts, so that the assumed relationship cannot hold over much of the domain. Outside the upwelling zones, change in SST must arise from other mechanisms which are not closely related to changes in upper layer depth. These include horizontal temperature advection and surface fluxes of heat.

Even in the upwelling zones there is considerable evidence that horizontal advection is important during ENSO. This has been suggested by Schopf and Harrison (1983) and Gill (1983), among others. The role of surface heating is more questionable; it is more difficult to estimate from available data. One hypothesis is that there is an approximate inverse relationship between ENSO SST anomalies and oceanic surface heating anomalies. Although this probably does not hold equally well
in all locations or at all times (Leetmaa, 1983), the results to date do not contradict the hypothesis, as is apparent from Weare (1983), Ramage and Hori (1981), and others.

If one assumes that the anomalous surface heating depends only on the local SST anomalies, then the influence of the atmosphere in producing SST anomalies is still reducible to the anomalous surface stress, provided now that the mean temperature and current fields are specified. The stress anomalies excite waves which transport mass and advect existing total temperature gradients, producing local temperature anomalies both at depth and at the surface. Once temperature anomalies exist, they can be advected by the mean current fields, further altering the local temperatures (this includes the mean upwelling effects previously discussed).

Clearly, once temperature advection is considered, the SST tendency is a rather complicated spatially and temporally varying function of the imposed surface stress. Because of the likely importance of advection, however, a successful modeling effort presumably must acknowledge this. Such is the conclusion of this study. As will be shown subsequently, the regional differences in temperature variability during ENSO, the spatial coherence of the temperature anomalies, and the movement of the anomaly maxima are all found to depend on advective effects in the model.

3.2 The Model

Observations indicate that the ocean dynamics relevant to ENSO are largely captured by simple linear theory, but that the thermodynamics
is more complicated. The present model was designed from this perspective. Dynamical quantities are calculated using a linear model and then applied to an advective temperature evolution equation. The calculation is straightforward, however, because the thermodynamics is not allowed to alter the dynamics. The assumption behind this is that the dynamical forcing due to changes in thermal structure is small compared to the direct forcing by the large anomalies in surface stress during ENSO. This is suggested by the success of the models which ignore thermodynamics. With the present model, the computational savings which results from decoupling the dynamics and thermodynamics in this way is significant. An otherwise costly nonlinear calculation becomes a straightforward linear calculation.

The dynamical model used is that of Cane and Patton (1984), modified by the addition of a homogeneous, frictional surface layer. The Cane and Patton (1984) model is a linear, shallow-water, equatorial beta-plane model, with an additional approximation which excludes high-frequency, short-wavelength motions. It is computationally very fast and efficient, and is especially suited for studying large-scale, low-frequency variability such as that characteristic of ENSO. Busalacchi and O'Brien (1980, 1981) demonstrated the relevance of the linear dynamics to the observed seasonal and interannual variability of the tropical Pacific. In the earlier work they forced the model with prescribed mean wind stress, plus annual and semi-annual harmonics as derived by Meyers (1979), and obtained upper layer depth variations which are similar to the observed annual thermocline variability. In the later study, they force the model with observed winds for a 10-year period (1961–1970)
encompassing two ENSO events, and are able to reproduce the salient features of the eastern Pacific upper layer depth variability.

The shallow-water equations arise immediately from considering a two-layer ocean in which only the upper layer is dynamically active (i.e., the pressure gradient is assumed to be zero in the lower layer). Such models are often referred to as reduced gravity models. The approximation is reasonable for the tropical ocean because of the existence of a sharp thermocline which inhibits the downward propagation of waves generated in the upper ocean. However, the shallow-water system is also relevant to the continuously stratified problem, where the equivalent depths arise as eigenvalues of the vertical structure equation (see, e.g., Moore and Philander, 1977). The complete solution is obtained by summing the contributions of all the vertical modes in this case.

This approach was used by Cane (1984) in order to simulate observed sea-level anomalies during ENSO. Using a prescribed wind stress generated from the Rasmussen and Carpenter (1982) composite data, and calculating the equivalent depths with a realistic density profile for the central Pacific, he was able to reproduce many of the salient features of the observations, especially in the eastern ocean where the variability is large and spatially coherent. He found that the largest single contribution to the total response came from the gravest barocline mode, for which he calculated an equivalent depth of 86cm.

In light of this, only the gravest mode is considered in the present study, and the same value for equivalent depth is adopted. For an equivalent depth $H_o = 86cm$, the equatorial Kelvin wave speed is $c = 91 \text{ m/sec}$ (it is defined by $c^2 = gH_o$). The results of Schopf and Harrison
(1984), who use a nonlinear model with a carefully constructed active mixed layer, suggest that the higher baroclinic modes may be ineffective in producing SST anomalies in the presence of a realistic mean circulation. Accordingly, using only a single mode in a much simpler linear model may be more justifiable than the linear theory alone would predict.

The dimensional governing equations for the single mode model are:

\[ U_t - \beta_o y V = -gH_o h_x + \tau^x/\rho - rU \]  
\[ \beta_o y U = -gH_o h_y + \tau^y/\rho - rV \]  
\[ h_t + (U_x + V_y) = -rh. \]

where \( H_o \) is the equivalent depth, \( g \) is the acceleration of gravity, \( U \) and \( V \) are the zonal and meridional transport in the upper layer, \((\tau^x, \tau^y)\) are the components of surface stress, \( r \) is a linear dissipation coefficient, and \( h \) is the anomaly in upper layer depth about the mean depth. The term '\( V_t \)' that would normally appear in [3.2.1b] is eliminated by the long-wave low-frequency approximation (Cane and Patton, 1984).

(In the reduced gravity formalism, the expression \( gH_o \) would be replaced with \( g' \), and \( h \) with \( H \), where \( g' = g \Delta \rho/\rho \), \( \Delta \rho \) is the density difference between the two layers, and \( H \) is the upper layer thickness.) Defining the mean upper layer thickness as \( H_1 \) gives the relation \((U,V) = H_1 (u,v)\), where \( u \) and \( v \) are the components of velocity.

Equations [3.2.1] should not be expected to give a good estimate for the three-dimensional current field near the ocean surface since the entire upper layer is assumed homogeneous, whereas in reality a
turbulent well-mixed layer exists near the surface. In attempting to predict changes in sea surface temperature it is important to better resolve the near-surface region. The simplest conceivable approach has been adopted: a linear, homogeneous, frictional surface layer is added (see Cane, 1979; Schopf and Cane, 1983). The approach is motivated by the following considerations. The effective friction of the surface layer is assumed to be very much greater than that below. Accordingly, nearly all of the frictionally induced (or Ekman) transport in the upper ocean should occur in this region alone. If the layer has a nearly constant depth, as assumed here, then no independent pressure gradients can develop in this layer. Therefore, the horizontal pressure gradient will still be nearly uniform throughout the entire upper layer (including the surface layer), and the component of the flow which is related to the pressure gradients will be likewise.

By this reasoning it follows that the non-frictional component of transport predicted by equations [3.2.1] need not be altered, but that the frictional (Ekman) transport should be concentrated in the surface layer. Equations for a homogeneous surface layer, consistent with the above discussion, are (dimensionally)

\[-\beta_0 y V_s = \tau(x)/\rho - r_s U_s\]  \[3.2.2a\]

\[\beta_0 y U_s = \tau(y)/\rho - r_s V_s.\]  \[3.2.2b\]

Solving for the transports gives
\[ U_s = (r_s \tau(x) + \beta_0 y \tau(y) / \rho(\beta_0^2 y^2 + r_s^2)) \]  \[ V_s = (r_s \tau(y) - \beta_0 y \tau(x) / \rho(\beta_0^2 y^2 + r_s^2)) \]  

or

\[ U_s = \frac{1}{\rho(f^2 + r_s^2)} [r_s \tau + f(\tau X \hat{k})], \]

where \( f = \beta_0 y \) and \( \hat{k} \) is the unit vector in the vertical direction.

Away from the equator the transport approaches \((\tau X \hat{k})/\rho f\), which is the usual result from Ekman theory. The transport is independent of both the form and magnitude of assumed friction. Very near the equator, where \( f \rightarrow 0 \) the balance must be different from the usual Ekman theory. In this model, the equatorial transport is \( \tau / \rho r_s \), and is thus directed as the local stress, with magnitude inversely proportional to the strength of the dissipation.

The total transport obtained from solving [3.2.1] includes the frictional transport, although by assumption it is distributed uniformly throughout the entire upper layer. This transport must therefore be subtracted out, and then added in just the surface layer. By doing this, the total transport is unchanged, but the near-surface currents are modified. Using the fact that the layer transports are related to the layer currents by the layer depths, the final surface currents are

\[ U^e_{sfc} = u^e - U_s^e / H_1 + U_s^e / H_s \]  \[ [3.2.5] \]
where \( u^* \) is the dimensional velocity from solving [3.2.1], \( H_1 \) is the upper layer depth (which includes the surface layer), \( H_s \) is the surface layer depth, and \( u^*_{sfc} \) is the frictional transport defined by [3.2.4].

A vertical velocity at the base of the surface layer can be determined by conservation of mass in the surface layer. Taking the vertical velocity to be zero at the surface, it follows that

\[
w^* = H_s \left( \nabla_h \cdot u^*_{sfc} \right)
\]  

([3.2.6])

(\( \nabla_H \) is the horizontal divergence operator)

The dominant contribution to \( w^* \) comes from just the frictional component of the surface flow (the remaining component is nearly geostrophic), and this portion, using [3.2.3], has the form

\[
w_f^* = \frac{1}{\rho(f^2 + r_s^2)} [r_s (\nabla_h \cdot \tau) - f(\nabla \times \tau) \cdot \hat{k}]
\]  

\[
+ \frac{\beta_o}{\rho(f^2 + r_s^2)} [(f^2 - r_s^2) \tau(x) - 2fr_s \tau(y)]
\]  

([3.2.7])

Away from the equator, upwelling/downwelling is dominated by the curl of the wind stress, as usual. On the equator upwelling is produced by either an easterly wind stress or a divergent wind stress, with the former usually dominating. Finally, coastal upwelling can be calculated by assuming that the frictional flow goes to zero at the boundary (or equivalently, that the wind stress goes to zero at the boundary).
For an eastern boundary, it is seen that upwelling/downwelling occurs if the wind stress is directed equatorward/poleward in regions off the equator.

Using [3.2.1], [3.2.4], [3.2.5] and [3.2.6], the near-surface current structure is determined. These currents are used to compute the advective components in the temperature evolution equation. The surface layer heat loss is taken to be proportional to the surface temperature anomaly as discussed earlier. It should be emphasized that such a simple specification could not be used to predict the mean SST. In that case the surface heating plays a dominant role and cannot be parameterized simply in terms of the temperature. For example, the heating due to insolation (together with atmospheric feedback processes such as cloud cover) has not entered explicitly into the temperature anomaly equation but is certainly a major factor in the total energy budget for the mixed layer. Schopf and Cane (1983) have examined the problem of modeling the SST field with a more sophisticated model. Their results suggest that many of the assumptions of the present model make it inadequate for simulating the mean SST. For these reasons the present study is restricted to the temperature anomalies, and the mean temperature fields are specified.

For the temperature anomaly evolution, the role of upwelling requires special attention, because SST can be affected only if the total vertical velocity is positive. There are four distinct cases to consider. If the mean field is downwelling and the total field is also downwelling, then clearly there is no effect on the surface temperature anomaly. If the mean field is downwelling but the total
field is upwellng, then the surface temperature anomaly is affected
in proportion to the strength of the total upwellng (which is less
than the anomalous upwellng in this case). If the mean field is upwellng
but the total field is downwellng, then the surface temperature anomaly
is affected, but only in proportion to the strength of the mean upwellng
(which is less than the strength of the anomalous downwellng in this
case). In other words, the cooling effect that is normally operative
can be eliminated, but no further influence can be exerted. Finally,
if the mean field is upwellng and the total field is upwellng, then
the surface temperature anomaly is affected according to the sign and
magnitude of the anomalous vertical velocity. Anomalous upwellng
produces anomalous cooling and anomalous downwellng produces anomalous
warming.

With components as just described, the temperature anomaly equation
has the following form (dimensionally):

$$\frac{\partial T_s}{\partial t} = -u_{sfc} \cdot \nabla (\bar{T}_s + T_s) - u_{sfc} \cdot \nabla T_s - \gamma_1 F_1(w)T_z - \gamma_2 F_2(\bar{w}, w)(\bar{T}_z + T_z) - \alpha T_s$$

[3.2.8]

where bars denote mean fields. The functions $F_1$ and $F_2$ are defined by

$$F_1(\bar{w}) = \begin{cases} -\bar{w}, & \bar{w} > 0, \\ 0, & \bar{w} < 0 \end{cases}, \quad F_2(\bar{w}, w) = \begin{cases} 0, & \bar{w} \leq 0 \text{ and } \bar{w} + w' \leq 0 \\ \bar{w} + w, & \bar{w} \leq 0 \text{ and } \bar{w} + w' > 0 \\ -\bar{w}, & \bar{w} > 0 \text{ and } \bar{w} + w' \leq 0 \\ w, & \bar{w} > 0 \text{ and } \bar{w} + w' > 0 \end{cases}$$

They describe mathematically the asymmetries in upwellng effect which
were just described. The vertical advection terms are evaluated at
the base of the mixed layer, but estimates for temperature changes at the surface are sought; thus, the factors $\gamma_1$ and $\gamma_2$ are introduced. Physically it seems reasonable to choose $\gamma_1$ and $\gamma_2$ less than one. In other words, because the layer has a finite depth, the mean temperature changes for the layer are expected to be less than those estimated at a point corresponding to the base of the layer.

The mean surface temperature is prescribed using the long term monthly mean fields computed by Rasmussen and Carpenter (1982). There is comparatively little data for the mean vertical temperature structure. Fortunately it is only very near the equator (and near the coastline) that matters, because elsewhere there is either mean downwelling or vertical advection is unimportant. A number of observations have been made in the equatorial plane. For these calculations values of $\overline{T_z}(x)$ at the equator were obtained using Figure 3.1, (and other similar profiles) which is reproduced from Colin et al. (1971). There is some ambiguity in how to prescribe $\overline{T_z}$ for the present model, since the surface layer depth is assumed to be constant in the model, whereas Figure 3.1 indicates that it is actually variable, being much deeper in the western Pacific. However, if a value of $\overline{T_z}$ appropriate to the depth of the model surface layer is chosen everywhere, then the variability is at least qualitatively accounted for. In regions where the mixed layer is very deep and entrainment of cool water from below probably has negligible effect on SST, the near surface temperature gradient is very small. $\overline{T_z}$ at the depth of the model surface layer is therefore very small. By using this value of $\overline{T_z}$ in [3.2.8] the model sensitivity to anomalous upwelling is small, as expected for the real ocean in this case. In regions
Fig. 3.1: Isotherms along the equator in the Pacific Ocean (after Colin et al., 1971).
where the mixed layer is very shallow and entrainment of cool water is important for SST, the estimated $T_z$ at this reference level is very large and model upwelling effects are likewise important. By estimating $T_z$ in this fashion, then, the model has a sensitivity to local upwelling, which is analogous to what might be expected in the actual ocean. The values for $T_z$ used in the simulations are listed in section 3.5. Clearly, these values are not well determined and the choices are somewhat subjective. The goal is only to mimic qualitatively the effect of the observed variability in thermal structure.

The anomalous temperature gradient $T_z$ also must be parameterized. Both the temperature anomalies at the surface and at depth contribute to the anomalous gradient. The temperature anomaly at the surface, of course, is predicted explicitly. The temperature anomaly at depth is assumed to arise from depressing or raising the thermocline according to the anomalous inflow or outflow of mass in the upper layer. This corresponds roughly to considering vertical displacements of the mean subsurface temperature profile in accordance with the anomaly in upper layer depth. To quantify this simply, the mean vertical temperature profile is approximated by the analytic form of a hyperbolic tangent function. Thus, by considering vertical displacements of this profile in proportion to the upper layer depth anomaly, the resulting subsurface temperature anomaly at a fixed reference level has the form

$$T_{\text{sub}} = A(\tanh [B(\bar{h} + h)] - \tanh [B\bar{h}]),$$  \[3.2.9\]

where $\bar{h}$ is the mean thermocline depth, and $A$ and $B$ are constants related to the steepness and total amplitude of the temperature variation in
the mean profile. According to [3.2.9], \( T_{\text{sub}} \) always has the sign of the upper layer depth anomaly, but varies in magnitude depending on the mean depth of the thermocline. For a given layer depth anomaly, the estimated subsurface temperature anomaly is largest where the thermocline is shallowest. Values for \( \bar{h}(x) \), \( A \), and \( B \) were derived from Figure 3.1 and are given in section 3.5.

The anomalous temperature gradient, \( T_z \), is determined from the surface and subsurface temperature anomalies according to

\[
T_z = \frac{(T_s - T_{\text{sub}})}{H_s}
\]

[3.2.10]

where \( H_s \) is the surface layer depth.

It remains only to specify the mean current fields in [3.2.8]. This is done by running the dynamical model ([3.2.1], together with [3.2.4], [3.2.5] and [3.2.6]) with the long-term monthly mean stress fields calculated from the Rasmussen and Carpenter (1982) data set. Temperature prediction is not done; only the dynamical quantities are calculated. Based on the results of Busalacchi and O'Brien (1981), there is reason to expect that the model can simulate much of the observed annual cycle in currents and upper layer depth. In the present case, an explicit prediction of surface currents can be made and compared with available surface current data. The results of this calculation are presented in section 3.4.

For reference, the full model equations are rewritten in Appendix C in dimensional and nondimensional form.
3.3 Computational Procedures

The numerical procedure for solving equations [3.2.1] is given in detail in Cane and Patton (1984). The solution is obtained by treating the Kelvin wave portion of the solution analytically and calculating the remainder of the solution by finite difference methods on a staggered grid, taking advantage of the fact that this part of the solution has no eastward propagation. The procedure allows large timesteps and is limited only by accuracy considerations. The scheme introduces no spurious sources or sinks of mass or energy.

For all the present calculations the timestep was 10 days, and the grid spacing was $2^\circ$ latitude (220 km) in $x$ and $0.5^\circ$ latitude ($\sim$56 km) in $y$.

The boundary conditions for the ocean are those of no flow perpendicular to the basin boundaries. Although the model is designed to allow irregular geometry, a rectangular basin is chosen here, with boundaries at $29.75^\circ$N, $29.75^\circ$S, $124^\circ$E and $80^\circ$W.

The temperature anomaly equation is evaluated on a spatially uniform grid using forward time differencing. Since the temperature does not influence the dynamical calculation, the accuracy or stability of that calculation is not affected by the temperature prediction scheme. Horizontal advection terms are evaluated using the modified upwind difference scheme. This has the form:

$$
(uT_x)_{i,j} = \begin{cases} 
(u_{i,j-1} + u_{i,j})(T_{i,j} - T_{i,j-1})/2\Delta x, & u_{i,j} > 0 \\
(u_{i,j} + u_{i,j+1})(T_{i,j+1} - T_{i,j})/2\Delta x, & u_{i,j} < 0 
\end{cases} 
$$

[3.3.1a]
\[ (vT_y)_{i,j} = \begin{cases} \frac{(v_{i+1,j} + v_{i,j})(T_{i+1,j} - T_{i,j})}{2\Delta y}, & v_{i,j} \geq 0 \\ \frac{(v_{i-1,j} + v_{i,j})(T_{i,j} - T_{i-1,j})}{2\Delta y}, & v_{i,j} < 0 \end{cases} \]

where \( i \) and \( j \) are the grid indices in the \( y \) and \( x \) directions, respectively.

The scheme is second-order accurate, and is transportive and conservative — it advects only in the direction of the flow, and does not produce artificial sources. These features of the scheme are important because in part of the domain (that is, away from the upwelling zones) there are no local sources for temperature anomalies: they can be produced only by advective processes, and therefore are sensitive to deficiencies in the numerical scheme.

Equations [3.3.1] are numerically stable only if \( \Delta t \leq \text{MIN}(\Delta x/u_{\text{max}}, \Delta y/v_{\text{max}}) \) where \( u_{\text{max}} \) is the largest zonal component of velocity, and \( v_{\text{max}} \) is the largest meridional component of velocity. This sets an upper bound on \( \Delta t \) which diminishes with increasing velocities. The temperature and current fields are evaluated on a grid with \( \Delta y = 2^\circ \) latitude (~220km) and \( \Delta x = 5.625^\circ \) latitude (~620km). This seemingly peculiar choice anticipates the coupled model calculations, in which the atmosphere model uses such a grid. Within the domain of the ocean, then, there are 27 grid points in \( x \) (the first at 129.375\(^{\circ}\)E, the last at 84.375\(^{\circ}\)W) and 30 points in \( y \), though the calculation is done only between 19\(^{\circ}\)N and 19\(^{\circ}\)S (20 points). Since the grid for the dynamical variables \((u,v,h)\) is finer, domain averages of these variables are used in evaluating the terms in the temperature equation. A domain is made up of 15 points on the finer grid (3\(^{\circ}\) in \( x \) by 5\(^{\circ}\) in \( y \)), making the domains adjacent and nearly nonoverlapping.
With the above choices for \( \Delta x \) and \( \Delta y \), it is necessary to reduce
At from the value of 10 days used by the dynamical model. A value
of 2.5 days was used, which guarantees stability for zonal currents
less than 2.8 m/sec and meridional currents less than 1 m/sec. Thus
the temperature equation is iterated 4 times for each timestep in the
dynamical model, and the variables \((u,v,h)\) are held fixed. \( T_s \) and
\( T_z \) are updated at each iteration.

Evaluation of the mean upwelling field requires special attention
at the eastern boundary because of the idealized geometry of the model.
For the interior grid points, the upwelling is computed using a centered
difference analog to equation [C.27]. At the eastern boundary, however,
the calculation is done with a one-sided difference equation in \( x \), i.e.,

\[
\bar{w}_{i,j} = \frac{(v_{sfc})_{i+1,J} - (v_{sfc})_{i-1,J}}{2\Delta y} + \frac{(u_{sfc})_{i,J+1} - (u_{sfc})_{i,J}}{\Delta x} \tag{3.3.3}
\]

with \((u_{sfc})_{i,J+1}\) set to zero (being over land). This is done so that
\( \bar{w}_{i,J} \) better approximates the near-coastal mean upwelling, which is
a rather small-scale feature not otherwise well resolved on the coarse
grid used for \( w \). Away from the equator, the major contribution to
\( \bar{w}_{i,j} \) comes from the Ekman flux component which depends only the local
value of \( \tau(y) \), the longshore wind stress (see [C.25]). Because of
this, the geometry idealization of the model presents a problem. In
regions where the model coastline differs from the actual coastline,
an incorrect estimate for the coastal upwelling is made if the stress
field (which comes from observations) is evaluated at the model boundary
instead of the actual boundary. In an attempt to remedy this problem,
the input wind fields were modified so that the winds adjacent to the actual coast also occur at the model coast. In regions where the model coast is to the east of the actual coast, the winds at the actual coast were extended uniformly eastward to the model coast, and in regions where the model coast is to the west of the actual coast, the winds adjacent to the model coast were replaced with those adjacent to the actual coast. In this way the Ekman-induced upwelling at the model coast approximates that at the actual coast while still preserving the simple geometry.

In the temperature anomaly calculations the output is smoothed by application of two passes of a (1-2-1) filter in y and one such pass in x. This is done only for the output, and does not affect the evolution of the temperature anomalies.

3.4 Mean Circulation Calculation

Equations [3.2.1], [3.2.4], [3.2.5] and [3.2.6] together comprise the mean circulation model. The forcing is prescribed from the long-term monthly mean wind fields computed by Rasmussen and Carpenter (1982). Stresses are defined using a standard bulk formula

\[ \tau = \rho C_D u_a |u_a| \]  

[3.4.1]

where \( \rho C_D \) is taken to be \( 3.2 \times 10^{-3} \text{ kg/m}^3 \). The model timestep is 10 days, and the spatial grids are as described in the previous section. Spatial and temporal interpolation are used to compute the instantaneous forcing at each model gridpoint. The ocean dissipation coefficient \( \tau \) is taken to be \( (2.5 \text{ years})^{-1} \), and the surface layer friction coefficient
(\(r_s\)) is taken to be \((2 \text{ days})^{-1}\). This seemingly large value for \(r_s\) gives model upwelling velocities of the order of \(10^{-3} \text{ cm/sec}\), comparable to what can be inferred from observations and what is found in nonlinear models (see Cane, 1979). For decreasing \(r_s\) the model equatorial upwelling velocity increases rapidly (like \(r_s^{-2}\), cf. [3.2.7]).

The mean upper layer depth \(H_1\) is taken to be 150m: the surface layer depth is 50m. The values are chosen to be representative of the average depth of the equatorial thermocline (measured by the \(20^\circ\) isotherm), and the average depth of the near-constant temperature layer at the equator (see Figure 3.1).

To describe the mean cycle the model is run until all transients have decayed, thus giving an annually periodic solution. By year 6, the solution is indistinguishable from that in year 5. The results for March, June, September and December of year 6 are presented here.

In Figures 3.2(a)–(d) are shown the model predictions for upper layer depth anomaly for the four months. The total model upper layer depth is obtained by adding the mean upper layer depth \(H_1\) (150m) to the anomalies. In all seasons the upper layer shallows from about \(170^\circ\text{E}\) to \(100^\circ\text{W}\) near the equator. The annual mean depth at the easter near the east coast is about 50m: the annual mean depth in the vicinity of \(170^\circ\text{E}\) is about 180m. The overall east-west slope is smallest in June and largest in December. These features are in agreement with the observational results of Meyers (1979), who describes the annual variability of the \(14^\circ\) isotherm in the equatorial plane. Along the equator the results are very similar to those of Busalacchi and O'Brien (1980), although away from the equator the structure is different.
Fig. 3.2 Model upper layer depth departures from the mean depth (150m), for (a) March, (b) June, (c) September, and (d) December. $x$ is measured in degrees of longitude east of the Greenwich meridian, and $y$ is measured in degrees of latitude from the equator. Contour intervals are 20m.
Their model gives an upper layer thickness which is nearly independent of latitude west of 120°W and poleward of 8°, whereas this model gives a rapid deepening of the upper layer poleward of about 10°N and 5°S with maxima near 20° in either hemisphere. The disagreement can be traced to differences in boundary conditions. Their model has boundaries at 12°S and 18°N, and allows propagation of coastal Kelvin waves and Rossby waves through the boundaries. The present model covers a larger domain and uses no-normal-flow boundary conditions as described earlier. The meridional structure obtained here is not inconsistent with observations in the central Pacific between 19°S to 19°N (see e.g., Wyrtki, 1974).

The model surface layer currents for the four months are shown in Figures 3.3(a)–(d). All of the major equatorial surface currents are evident — the westward flowing South Equatorial Current between ~3°N and 17°S, the (primarily) eastward flowing North Equatorial Countercurrent between ~5°N and ~10°N, and the westward flowing North Equatorial Current between ~10°N and ~17°N. In agreement with the observational results of Meehl (1982), the North Equatorial Current is strong during northern fall and winter, and weak during summer: the South Equatorial Current is strong during late summer/early fall and weak during winter: the North Equatorial Countercurrent is strong during summer and fall, and reverses to a westward flowing current during spring. In the far-western Pacific the model predicts sizeable eastward currents on the equator between May and August, at the time when the upper layer is shallowing in the west. The shallowing of the thermocline is evident in Meyers (1979) results, though strong eastward flow at the surface apparently is not observed. Possibly the predicted eastward mass flow is accomplished
in the subsurface Equatorial Undercurrent which exists within the deep mixed layer of the western Pacific. There is evidence that the Undercurrent is strong precisely during this period in the western Pacific (Philander, 1973). Such vertically varying features are, of course, precluded by the assumptions of the present model. However, if the undercurrent is indeed stronger at this time, then the model surface currents would compare more favorably with the vertically averaged mixed layer currents than with the actual surface currents.

The observational results of Meehl (1982) are based on shipdrift-derived currents from pilot charts. Wyrtki (1974 a,b) has estimated the zonal currents from sea-level fluctuations, thereby neglecting any frictionally induced component to the flow. His results differ somewhat from those of Meehl and from those presented here, especially in terms of the relative phase of the different current fluctuations. This suggests that the frictionally induced component of the surface layer flow is not negligible. Busalacchi and O'Brien (1980) find with their model that (i) the North Equatorial Current is strongest during summer and weakest during winter, (ii) the South Equatorial Current is strongest during late fall/early winter and weakest during spring, and (iii) both of these currents are very weak (< 10 cm/sec). The results differ from the observations (Meehl, 1980) in timing and amplitude (the observed currents are of the order of 50 cm/sec). Since the models are otherwise very similar (they give essentially the same result for upper layer pressure gradients), the better agreement obtained here presumably derives from the modifying effect of the surface layer.
The ratio of meridional to zonal component of the currents in Figures 3.3 appears to be larger than is observed. Because the model surface layer is taken to be homogeneous, the Ekman component of the current is directed the same as the Ekman transport — essentially poleward for easterly winds. In actuality, one expects the flow at the surface to be more nearly in the direction of the wind, and thus to be more zonally directed than the overall transport (classical Ekman theory gives a transport directed 90° to the right of the surface stress, but a surface flow only 45° to the right/left of the surface stress in the Northern/Southern Hemisphere). The discrepancy in direction therefore is to be expected with a homogeneous surface layer model.

The computed upwelling fields are shown in Figures 3.4(a)–(d). Being proportional to the divergence of the surface layer currents, the upwelling fields contain no new information, but they highlight the primary source regions for surface temperature anomalies. Sizeable upwelling is generally restricted to the Southern Hemisphere portion of the east coast and a 3°–4° band near the equator. Because of the coarse resolution of the model, the coastal upwelling is probably underestimated in magnitude, and overestimated in spatial extent. Upwelling is weakest in the central and eastern Pacific in March when the ITCZ spans the equator, and is strongest in summer and fall when the ITCZ is far north of the equator and the Southeast Trades are most intense. Coastal upwelling is generally greatest a few degrees south of the equator, and is most pronounced in the period from May to August. In the western Pacific, upwelling is weaker and more variable in location. During (Northern) winter and spring, upwelling is found to the north
Fig. 3.4 Model upwelling velocities for (a) March, (b) June, (c) September, and (d) December. Contour intervals are $10^{-3}$ cm/sec (heavy line is the zero contour).
of the equator, and downwelling to the south. During summer and fall the situation is just the opposite, with downwelling to the north of the equator, and upwelling to the south.

Monthly values for the mean upwelling and surface currents obtained in this calculation were used in the temperature anomaly simulations to be described next.

3.5 Simulations of SST Anomalies During ENSO

The full set of equations summarized in Appendix C comprise the SST anomaly model. The forcing for this simulation is taken from the three year ENSO composite anomalies of Rasmussen and Carpenter (1982), with stress anomalies calculated as before for the mean winds (see [3.4.1]).

The following parameter values were used: \( r_s = (2 \text{ days})^{-1} \), \( r = (2.5 \text{ yrs})^{-1} \), \( H_0 = 86 \text{ cm} \), \( H_1 = 150 \text{ m} \), \( H_s = 50 \text{ m} \), \( A = +28.0^\circ \) if \( h > 0 \) and \(-40.0^\circ \) if \( h < 0 \), \( B = .0125 \text{ m}^{-1} \) if \( h > 0 \) and \(+.03 \text{ m}^{-1} \) if \( h < 0 \), \( \gamma_1 = \gamma_2 = .75 \), \( a_s = (125 \text{ days})^{-1} \). Values used for \( \overline{h}(x) \) and \( \overline{T_z}(x) \) at the 27 model gridpoints are listed in Table 3.1. Estimating \( \overline{T_z} \) directly from Figure 3.1 would give smaller values in the west and larger values in the east by perhaps a factor of two. The 'flatter' profile given here was chosen deliberately to reflect the fact that the observations were made in non-ENSO periods when the east-west slope of the thermocline is largest. During ENSO the relaxation of the trade winds causes the thermocline to rise in the west and deepen in the east. Thus, the long term mean temperature profile, which includes ENSO and non-ENSO periods, and which is the appropriate profile for a temperature anomaly
<table>
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**Table 3.1.** Values for $\bar{T}_z$ and $\bar{h}$, in units of °C per 100 meters, and meters, respectively.
model, should be flatter than that of Figure 3.1. The model was run using both the profile given here and that estimated directly from Figure 3.1, with the former giving more realistic results in terms of amplitude and spatial scale of predicted temperature anomalies. Using the non-ENSO profile tends to give temperature anomalies that are too large in the east and too small further to the west, as would be anticipated from the above discussion.

The parameters A and B were given different values for positive and negative $h$ to account for the fact that once the upper thermocline surfaces in the eastern ocean, further temperature anomalies are not observed to occur. The explanation for this must involve the details of mixed layer physics which are ignored by this model. Accordingly, the observed behavior is imposed on the model by choosing A and B for negative $h$ which prevents large subsurface temperature anomalies from occurring. An alternative (and simplistic) way to view it is that once the mean thermocline rises above the model surface layer base, the local temperature gradient flattens again, and temperature anomalies at this depth become insensitive to $h$, much as the subsurface temperature in regions above the main thermocline in the western ocean are insensitive to $h$. In any case, for the chosen values, $h = +30m$ gives a positive subsurface temperature anomaly of $\sim 6^\circ C$ in the east and $\sim 0.5^\circ C$ in the west. For an equivalent negative departure, the corresponding values are $-3^\circ C$ and $-0.001^\circ C$. For larger $h$, the positive subsurface temperature anomalies continue to increase, whereas the negative anomalies remain nearly constant. These magnitudes appear reasonable, judging from the profiles of Figure 3.1. The sensitivity of the model to the values
for A and B is significant quantitatively, but not qualitatively, because the source regions are determined by the mean upwelling field which is entirely independent of the temperature parameterizations. A and B (together with $\gamma_1$ and $\gamma_2$) modulate only the magnitude of the sources.

The values for $r_s$, $r$, $H_0$, $H_1$ and $H_5$ are the same as was used in the mean circulation run. The appropriate value for $\alpha_s$ is highly uncertain from any observational results. Here it was chosen on empirical grounds. The temperature structure in the upwelling zones is comparatively insensitive to $\alpha_s$. The predicted temperature anomalies do increase somewhat as $\alpha_s$ is decreased, but the associated changes in temperature advection tend to lessen the sensitivity.

The model was initiated with all anomaly fields set to zero, starting in January (-1), and integrated forward in time using the prescribed stress anomalies. In order to demonstrate the effect of transients in the model, the calculation was continued beyond December (+1) (the end of the data set) by returning to January (-1) and proceeding through December (-1 - 1), then recycling again, etc., for five cycles. The total integration time is 16 years, starting from January (-1) to December (15). This procedure is reasonable because, as pointed out by Rasmussen and Carpenter (1982), the predominant spectral component in their data set corresponds very nearly to three years, which is exactly the period covered by the data set. To put it another way, the anomalous wind and SST fields for January (-1) are quite similar to those of December (+1), making it reasonable to consider integrating the model further by recycling the data input. An important question concerning the model can be addressed through this procedure: how important are transient
effects in the model — that is, how long does it take for the model to approach a periodic solution?

Some different measures of the near-equatorial wind stress forcing for the run are presented in Figures 3.5(a)-(c). The indices TAU1 and TAU2 represent area average values of the zonal wind stress in the domains (135E - 170W, 5S - 5N) and (170W - 135W, 5S - 5N), respectively. The function \( b(x,t) \) measures the equatorial wind stress in a different way: it is defined by

\[
b(x,t) = 2^{-1/2} \int_{-\infty}^{\infty} \psi_0(y) \tau(x,y,t) \, dy
\]

[3.5.1]

where \( \psi_0(y) \) is the lowest order Hermite function (equal to \( \pi^{-1/4} e^{-y^2/2} \)), and \( \tau(x) \) is the nondimensional zonal wind stress (\( y \) has been nondimensionalized here by the equatorial deformation radius for the ocean). Thus \( b(x,t) \) measures the projection of the forcing onto the Kelvin wave component of the flow, as described by Cane (1984). He also shows that the model layer depth anomalies at the east coast may easily be obtained from \( b(x,t) \) by integrating along characteristic lines which have a slope \( x/t = c \), where \( c \) is the Kelvin wave speed.

The major features of the equatorial wind anomalies which have been described earlier are evident also in the stress anomaly indices. Year (-1) is characterized by predominantly easterly anomalies, except for a brief period of westerly anomaly around May in the eastern central region (evident in TAU2 and \( b \) near 145W). Near the end of year (-1), westerly anomalies develop in the western region, relax somewhat in the spring, and then strengthen sharply in the fall of year (0). Meanwhile, during summer, the westerly anomalies spread eastward into the central
Fig. 3.5 (a) TAU1 index, (b) TAU2 index, and (c) $b(x,t)$, computed from the composited data (see text).
Pacific region and are largest there near the end of the year, by which time the westerly anomalies in the western region have declined precipitously and easterly anomalies have appeared. Year (1) is characterized by steadily decreasing westerly anomalies in the central Pacific, becoming easterly by mid-year, and continued easterly anomalies in the far-western Pacific.

As shown by TAU1 and TAU2, recycling the data at the end of year (+1) does not amount to a large discontinuity in the near equatorial wind stress, although there is an induced westerly impulse in the central Pacific which may not have a counterpart in any actual event.

Figure 3.6 displays three SST anomaly indices as calculated from the model results, compared with corresponding indices calculated with the composite data. The indices were chosen to coincide with those now routinely computed by observationalists. NINO1 covers the region \(5^\circ S - 10^\circ S, 80^\circ W - 90^\circ W\) which is characterized by the largest observed temperature anomalies. NINO3 spans a large region of the eastern equatorial Pacific, from \(90^\circ W\) to \(150^\circ W\) and \(5^\circ S\) to \(5^\circ N\). NINO4 includes the region between \(160^\circ E\) and \(150^\circ W\), and between \(5^\circ N\) and \(5^\circ S\); it is indicative of the variability near the dateline, which often appears to be distinct from that further to the east.

It is clear from Figure 3.6 that the effect of transients is minimal. The solution has the character of a forced, dissipative response to the prescribed forcing, with the differences between even the first and second cycles small.

Looking more carefully at the NINO1 indices, the pattern of variability is seen to be quite similar, with the exception of the large second
Fig. 3.6 (a) NINO1 indices (°C), from model output (light line) and from composited data (heavy line). (b) same for NINO3 indices. (c) same for NINO4 indices.
peak at the end of year (0) in the model which is not found in the observations. It should be pointed out that the stations along the South American coast do typically observe a large second peak in temperature at this time (see RC, Figure 8), comparable in magnitude to the one earlier in the year. All that is evident in NIN01, however, is a slight hesitation in the otherwise rapid decline following the earlier peak. In all other instances there is a one-to-one correspondence between relative maxima and minima in the model index and those in the observed index (during the first cycle). In year (-1) the model results would much more closely resemble the observation if the model were initiated with the observed SST anomaly distribution for January (-1). In that case, the small positive anomaly which occurs in August-September would more closely parallel the observed minimum in negative anomaly. This discrepancy is even more pronounced in the later cycles. The reason for this can be traced to the spurious westerly impulse which results from recycling the data set. This impulse forces a Kelvin wave which depresses the thermocline to the east, thereby producing a positive SST tendency through the mean upwelling effect. The result is that the temperature is rising during the first part of year (2), when both the model results for the first cycle and the observations show that it should be falling.

It is likely that the difference in the relative amplitude of temperature anomaly maxima arises because an annual cycle in subsurface temperature structure is not included in the model. That is, \( \bar{h}(x) \) and \( \bar{T}_z(x) \) were taken to be time-independent. In fact, there is a significant annual cycle. According to the results of the mean circulation
run described earlier, the thermocline in near-coastal region normally rises in early spring, and deepens in the fall. Furthermore the annual variability in layer depth is of the same order of magnitude as the interannual variability. This means that the sensitivity of the model to h would be appreciably greater in early spring than in fall, if an annual cycle were included. The result would be a larger initial temperature peak, and a smaller secondary one, in accordance with the observations. Furthermore, with increased spatial resolution near the coast, this annual modulation would be even more pronounced.

Turning next to the NIN03 indices, the agreement is seen to be better. Again there is a very good temporal correspondence between maxima and minima. The amplitude of the variability on the timescale of 2–3 months is larger for the model, but one must remember that, in addition to event to event compositing, the data represent 3-month running means. The amplitude of the maximum positive anomalies are the same, although there is a tendency for the model to underestimate negative anomalies in year (+1) for this region as a whole. It appears that this also might be resolved by including a seasonal variability in subsurface thermal structure.

The amplitude of temperature anomalies in the NIN04 region is much smaller than further to the east, both in the model and in the observations. Compared to the observations, however, the model appears to underestimate the magnitude of anomalies somewhat, especially during the period of largest anomalies late in year (0). In this case, the agreement is slightly better for the later cycles than for the first
cycle, but the general pattern of variability during all cycles is very similar to the observations.

Figure 3.7 shows a record of the model $h$-field along the equator during the final three-year cycle. Clearly evident are the coastal maxima in February to March and October to November of year (0) which precede the temperature maxima in this region. Another noteworthy feature is the indicated eastward progression of negative $h$ that occurs in the central Pacific in the fall of year (0). This movement is in conjunction with the eastward movement of westerly stress anomalies, and clearly indicates a forced response rather than a free wave signal.

In Figure 3.8-3.13 maps of observed and predicted SST anomalies for the times Aug (-1), December (-1), May (0), December (0), April (1), and August (1) are shown. The model results are for the first ENSO cycle in the simulation; the corresponding fields in later cycles are very similar.

a. August (-1): Pre-Event Phase (Figure 3.8)

The observations show small warm anomalies in the far-western Pacific and cool anomalies across the rest of the equatorial ocean, but largest in the central Pacific. In agreement with this, the model also has the largest negative anomalies in the central Pacific, and warm anomalies in the far-western region. However, the model predicts a tongue of warm anomaly in the eastern Pacific at this time. At the time this warm anomaly appears in the model (July (-1)), the observations show a decrease in the eastern Pacific negative anomalies, and at the time the warm anomaly disappears in the model (~ October (-1)), the
Fig. 3.7 The equatorial h-field across the basin, between months 132 and 168 (year(11)-year(14)) of the model run. Contours are 5m.
observations show an increase in negative anomalies. The temporal behavior is thus very similar. The spatial structure would be so as well if the model anomalies were superimposed on a field which was negative in the eastern and central Pacific. It is noteworthy that the composite temperature anomaly field for January (-1) has exactly this structure. Because the model was initiated with zero temperature anomalies at this time, the discrepancy can possibly be accounted for by the difference in initial states.

b. **December (-1): Onset Phase** (Fig 3.9)

This is the time when equatorial westerly wind anomalies appear in the vicinity of the dateline (see Figure 2.10b). To within the resolution of the time averaged data, the development of local warm SST anomalies is simultaneous. Aside from this small region of warm anomalies, the composite shows persisting cool anomalies further to the east, and a region of warm anomaly near 20°S in the eastern ocean. Similar features are found in the model SST anomaly field, though in all cases they are underestimated. This is a general problem for the western central Pacific region, and suggests that the model sensitivity may be too low in this region. Another possibility, though, is that the assumed relationship between SST anomalies and surface heat fluxes does not hold in this region.

c. **May (0): Peak Phase** (Figure 3.10)

This is the time period of the first, and usually largest, temperature anomaly peak near the east coast. The composite shows warm anomalies
Fig. 3.8 (a) composite SST anomalies for AUG(-1) and (b) model SST anomalies for the same time. Contours are .25°C (heavy line is the zero contour).

Fig. 3.9 As in Fig. 3.8, except for DEC(-1).
extending across the whole tropical ocean east of 150°E, with one local maxima in the central Pacific, and another, much larger one at the east coast south of the equator (3°S). The meridional extent of the warm anomalies is about 10°N and 20°S. In the far western Pacific small negative anomalies are beginning to develop at this time. Very similar features are found in the predicted SST anomaly field. The amplitude of the coastal peak is slightly underestimated, as is the anomaly maximum in the central Pacific. However, the spatial extent and location of the predicted warm anomaly region agrees well with the composite. The model also predicts small negative anomalies in the far-western region. It is clearly the effect of the mean upwelling field which causes the axis of maximum temperature anomalies to move south of the equator near the east coast. Model runs which ignored the mean upwelling gave anomaly fields which were symmetric about the equator.

d. **December (0): Mature Phase** (Figure 3.11)

By this time large positive anomalies fill the central and eastern ocean (160°E to the east coast), and negative anomalies exist to the west of 150°E near the equator. The maximum SST anomaly now is found near 120°W rather than at the coast as in May and June. Also evident is the poleward spreading of the warm anomalies which accompanies the growth in magnitude in the central and eastern Pacific during the fall. In most respects, the model results are similar. The major difference is found at the east coast, where the model predicts a second, sharp temperature peak, and the composite reveals no such pronounced feature.
As discussed earlier, at least part of the problem is the failure to include a seasonal cycle in the subsurface thermal structure of the ocean. The separate warm anomaly maximum which occurs at 120°W in the model is in good agreement with the composite. In both the composites and the model the central Pacific warms considerably between May and December, although again the magnitude is underestimated by the model in this region. Poleward expansion of the warm anomalies occurs in the model as in the observations except near the east coast to the north of the equator. The problem here is likely due to the geometry modifications that were made in this region, which locally affects both the mean and anomalous current fields. Presumably, this could be remedied by building a more realistic geometry into the model. Finally, the model predicts negative anomalies in the western Pacific, though smaller than those observed. The near-equatorial boundary between positive and negative anomalies occurs near 160°E, as in the composite.

e. **April (1): Termination Phase** (Figure 3.12)

Between January and April a rapid decay of the warm anomalies in the eastern Pacific occurs. The April composite already shows signs of the cold anomalies at the east coast which intensify and spread westward in the ensuing months. Warm anomalies persist in the central Pacific, though greatly reduced in magnitude from the December-January maximum. In the west, small cold anomalies still persist at this time. Certainly the most significant development to be modeled and understood during this time is the precipitous decline of the large warm anomalies in the entire eastern half of the tropical Pacific. This development
is well captured by the model. At the east coast, negative anomalies are just appearing at this time, and further to the west small anomalies have replaced the large anomalies of the previous December. The details of the temperature anomaly patterns differ significantly. However the characterization of being relatively nonuniform in space and rapidly changing in time applies equally well to both the model results and the composites at this time.

f. August (1): Post-Event Phase (Figure 3.13)

Between April and August the composites show a westward spreading of the cold anomalies to about 160°W and a poleward spreading to 10°N and 10°S. In August, warm anomalies continue to persist only in a small region in the central Pacific near the dateline. The model significantly underestimates the expansion of cold anomalies in the eastern Pacific. This may be due to an inadequate parameterization for $T_{\text{sub}}$ when $h$ is negative (recall that different parameter values were used in the idealized parameterization for $T_{\text{sub}}$ when $h$ was positive versus negative). Alternatively, it may result from the lack of a seasonal cycle in $\bar{T}_Z$ and $\bar{h}$ which was discussed above. In agreement with the composite, the model SST anomaly field shows warm anomalies persisting in the central Pacific (though too far east), and cold anomalies near the east coast. Both in the model and in the composites the magnitude of the anomalies during the negative (cold) phases of the ENSO cycle is considerably smaller than during the positive (warm) phase.
3.6 Mechanisms Controlling the SST Anomalies in the Model Simulation

It is a simple matter to determine the individual contributions of the various terms in the temperature equation. The level of agreement suggests that those processes important in the model also are so in the real ocean. A separate analysis was carried out for the three separate regions represented by NINO1, NINO3, and NINO4 which were defined earlier. In the following, contributions due to meridional advection are not discussed. Although advection of anomalous temperature gradients by the mean meridional currents is fundamentally important away from the equator, it is negligible in the upwelling regions where the largest temperature anomalies develop. Advection due to anomalous meridional currents is small (except at the east coast where it varies exactly in phase with the other terms). Consequently, the net effect of meridional advection is to spread meridionally the anomalies that develop by other mechanisms. This means that meridional advection cannot control the temporal variability of the area-averaged SST anomalies being considered here.

Figure 3.14 shows the relative contributions of mean upwelling, anomalous upwelling, zonal advection, and surface heat flux to the NINO1 variability. The heat flux, by assumption, is exactly proportional to the local SST anomaly. Thus it can be used (and is convenient) for making visual comparisons between the total temperature anomaly variations and the various forcing terms. In the case of NINO1 it is clear that the mean upwelling effect is the dominant controlling mechanism. Both the anomalous upwelling and zonal advection effects seem to vary in phase with the mean upwelling component, although they
Fig. 3.14 Contributions to model NINO1 variability. In (a), Q refers to the anomalous heat flux, $\bar{W}$ refers to vertical advection associated with the mean upwelling, and $w'$ refers to vertical advection associated with the anomalous upwelling. In (b), $u$ refers to the combined effect of all zonal advection terms.
are smaller in magnitude. The anomalous temperature gradients which are advected by the mean upwelling field arise from subsurface temperature anomalies, which, in turn, arise from upper layer depth anomalies. Thus the anomalous SST variability in this region apparently can be related very simply to the local variations in thermocline depth. Looking back to Figure 3.7, it can be seen that the upper layer depth anomalies and local SST anomalies are clearly correlated (with a slight lag). For example, the precipitous decline in SST anomalies in the spring of year (+1) is associated with a corresponding shallowing of the eastern ocean thermocline. The shallowing results directly from developments in the wind field: the advent of easterly anomalies in the western Pacific, and the weakening of westerly anomalies in the central Pacific (the axis of maximum westerly anomalies moves south of the equator in January of year (+1)).

The situation for NINO3 is more complicated, as shown by Figure 3.15. For this region, the contributions due to local upwelling and zonal advection can be as important as the mean upwelling. During the warm event period all the terms tend to be additive, contributing to the large signal. At other times, the upwelling-related terms tend to cancel, and zonal advection largely determines the surface temperature changes.

In the NINO4 region, the balance of terms is different from either of the above cases (Figure 3.16). Here the mean upwelling effect almost always acts to oppose the local SST anomalies rather than to create them. The anomalous upwelling term is large, and well correlated with surface temperature anomalies, but it is generally canceled by the
Fig. 3.15 As in Fig. 3.14, except for NINO3.
Fig. 3.16 As in Fig. 3.14, except for NINO4.
mean upwelling term, making the total influence of vertical advection rather small. Thus, the dominant mechanism controlling SST anomalies here is zonal advection. Still, vertical advection is not altogether unimportant; e.g. the continued rise in SST anomaly in the fall of year (0), when zonal advection temporarily becomes small, is apparently due to upwelling effects.

These results indicate that a realistic model simulation of the observed pattern of SST anomalies over the whole domain of interest cannot be achieved without including all components of anomalous temperature advection. Using a simple dynamical model and some observations from the 1972 ENSO event, Gill (1983) has proposed that the primary mechanism for producing SST anomalies in the equatorial central Pacific is zonal advection. The present results suggest a similar conclusion, except that a larger role for vertical advection is indicated in the eastern Pacific. In addition, this model indicates that meridional and vertical advection are very important near the east coast, and meridional advection by the mean meridional currents is dominant in interior regions away from the equator.

3.7 Simulation of the 1982-3 ENSO Event

A data set of tropical Pacific surface winds in the period July (1980) to March (1983) was obtained from Dr. J. O'Brien at Florida State University. Stress anomalies were computed using a monthly climatology developed from the same data set over the longer period, January (1961) to March (1983). The model was run with these stress anomalies, starting in October (1981). Defining 1982 as year (0), the integration
proceeds for 18 months between October \((-1)\) and March \((+1)\). Unfortunately, the data does not extend to the end of the ENSO event — large wind anomalies and SST anomalies persisted in the eastern Pacific for several more months beyond March, 1983. However, data from this period was not yet available when the set was acquired.

Figure 3.17 shows a time-longitude plot of \(b(x,t)\) which measures the equatorial zonal stress anomalies as described earlier. Comparing with Figure 3.5(c), it is clear that the 1982–3 event was not only greatly larger in magnitude than the composite event, but it also proceeded quite differently. The eastward expansion of large anomalies which occurs only briefly in the composite event continued onward until the anomalies reach the far-eastern Pacific in this event. Also, accompanying the large westerly anomalies during 1982 are apparently sizeable easterly anomalies in the eastern Pacific. The first large westerly anomalies in the composite event occur near the dateline in January of year \((0)\), whereas in the 1982–3 event they occur at about the same time, but further west near \(145^\circ\)E. In both cases, the westerly anomalies decrease during the spring, and then return in June and remain through the end of the event.

All of the above must be qualified by emphasizing that the anomalies were derived from different data sources, using different processing techniques, and with different climatologies. Because of this, biases can be introduced which make quantitative comparisons unreliable. For example, the wind field climatology being used here covers a twenty-year period which only partially overlaps with that of Rasmussen and Carpenter’s climatology. With such short records, the large interannual
variability of the winds can lead to different results for mean conditions, and thus different estimates for the anomalies.

In view of the above, the most serious limitation of this calculation is the lack of a SST data set which is compatible with the wind data set used. The model uses the Rasumssen and Carpenter climatology for mean SST, rather than a climatology for the same period as the surface wind climatology. Furthermore, the model results will be compared with the Climate Analysis Center SST anomalies, which are based on yet another climatology. Clearly, quantitative evaluation is not possible. The model results can be judged only with respect to the most salient features of the temporal and spatial variability.

Figure 3.18 shows the model predictions for the three SST anomaly indices NINO1, NINO3, and NINO4 over the course of the integration. Figure 3.19, reproduced from Arkin et al. (1983), shows the corresponding anomaly indices computed using observations and the Reynolds (1982) climatology. The model indices NINO1 and NINO3 decrease in spring of year (0), then rise in May (0), decline again briefly during early summer, and then rise sharply through the end of the year. After that, NINO3 decreases, but NINO1 continues to increase, and is increasing sharply in March (+1), at the end of the integration. It is only at this time that NINO1 surpasses NINO3 in magnitude. The NINO4 index has a much more uniform behavior, rising slowly throughout the whole of year (0), and then decreasing by March (+1). Roughly similar features are found in the observations. The quantitative agreement would be much better if ~1°C were added to the model results. This discrepancy could easily arise from the climatology problem. Aside from this,
Fig. 3.17 $b(x,t)$, computed from the observed wind anomalies between August, 1981 and March, 1983.

Fig. 3.18 Model predictions for NINO1, NINO3, and NINO4 (in °C). Time is in years where year(0) corresponds with 1982,
Fig. 3.19 NINO1, NINO3, NINO4, and other indices computed from observations between 1979 and 1983 (after Arkin et al., 1983).
the initial rise in NINO1 and NINO3 is apparently underestimated by the model. Recall that this is the same problem encountered in the composite simulation. The explanation is presumably the same — namely, the lack of a seasonal cycle in subsurface thermal structure in the model. The model is not expected to reproduce the high frequency variability of the observations because the input wind stress was averaged with a 3-month running mean. In both the observations and the model, the near coastal warming slightly lags the eastern Pacific warming measured by NINO3. This is generally regarded as one of the most significant oceanographic differences between the 1982-3 ENSO event and the recent preceding ones.

In Figure 3.20 are shown the predicted SST anomaly field for May (0), and the March-May computed anomaly field from Arkin et al. (1983). This is the time characterized by the initial, large coastal warming in the composite event. There is no evidence of such large warming in either the observations or the model predictions for this event. Neither exhibit large anomalies at any location at this time.

Figure 3.21 shows the corresponding maps for December (0), the time of maximum temperature anomaly for the composite event (the observations are for the period December to February). The patterns are quite similar, with large positive anomalies in the central and eastern ocean, maximized near 130°W. As mentioned before, there is a nearly uniform 1°C difference between the two maps. The discrepancy north of the equator near the east coast was encountered in the earlier simulation as well, and was attributed to the model simplification of the basin geometry there.
Fig. 3.20 (a) model SST anomalies corresponding to May, 1982; and (b) observed SST anomalies for March-May, 1982 (after Arkin et al., 1983).

Fig. 3.21 (a) model SST anomalies corresponding to December, 1982; and (b) observations for December, 1982-February, 1983 (after Arkin et al., 1983).
Fig. 3.22 (a) model SST anomalies corresponding to March,1983; and (b) observations for March-May,1983 (after Arkin et al.,1983).
A comparison for March (+1) is shown in Figure 3.22 (the observations are for the period March to May). By this time during a normal event, temperatures are returning to near normal in the eastern Pacific. In neither the observations nor the model results do they do so in this case. At this time both maps show a decrease in anomaly amplitude in most of the central and eastern Pacific, but an increase in amplitude at the coast. Both also show a more asymmetric distribution of temperature anomaly about the equator, with more of the warm anomaly region lying to the south in the eastern Pacific. Judging from the trend in NINO1 at the end of the integration, the coastal anomaly in April and May would likely be even greater than in March, in agreement with the observations.

An evaluation of the mechanisms controlling SST anomalies was done as in the earlier simulation. Perhaps not surprisingly, the same local balances were found to hold. Again, the conclusion is that the full temperature anomaly equation is required to produce a realistic simulation.

3.8 Discussion

The simulations have shown that the model can reproduce the important features of observed SST anomalies during ENSO events, given only the sequence of observed wind anomalies. This is true not only for the composite canonical event, but also for the event of 1982–3, which evolved quite differently. In view of the many simplifying assumptions of the model, these results are noteworthy. Perhaps the most significant conclusion that can be drawn here is that, at least to lowest order,
the observed ENSO SST variability can be explained solely in terms of the changing patterns of surface wind. Secondly, the results suggest that the relevant dynamics is very simple, involving essentially linear motions with a simple vertical structure. Finally, the indication is that the mechanisms controlling SST anomalies are more complicated, and require an explicit representation for horizontal and vertical advection in the oceanic near-surface layer.

In some of the details of the simulation, the model results differed from the observations. The two most significant instances were the relative amplitude of the initial coastal warming in May to June, compared with that at the end of the year, and degree of eastern Pacific cooling in the year following the warm event. These discrepancies were attributed to inadequacies in the parameterization of vertical temperature gradients, especially through the lack of a seasonal cycle in the subsurface temperature structure. This could be remedied by developing a parameterization for the total temperature gradient which depends on the total thermocline depth (or upper layer depth in the model), and specifying the mean monthly layer depth from the mean circulation calculation that was forced with long-term monthly mean winds. In the anomaly calculation, the total upper layer depth could be calculated at each time and place, together with the total upwelling velocity. From this the total vertical advection could be calculated using the temperature gradient parameterization. The total temperature anomaly tendency (due to vertical advection) would then be obtained by subtracting the corresponding vertical advection calculated from just the mean upwelling and mean upper layer depth for the particular time and place. Not only does this procedure guarantee
a consistent treatment of anomalous temperature gradients and mean temperature gradients, it also offers a means for including the influence of a seasonally varying subsurface temperature structure. In addition, an increase in spatial resolution near the east coast would undoubtedly improve the model.

Even in its present form, however, the model is quite suitable for the preliminary atmosphere-ocean interaction studies undertaken here. There are no shortages of possible mechanisms of interaction, even with this relatively simple model. There are the local influences of the wind field which induce anomalous horizontal and vertical surface layer currents, which, in turn, can create local temperature anomalies by advecting the existing nonuniform temperature field. In addition there are the local and/or remote influences of the wind field through the excitation of large-scale upper ocean waves with associated current anomalies and thermocline displacements. The role of these various mechanisms in the presence of an atmosphere that is sensitive to particular patterns of SST anomalies is the ultimate focus of this work, and the subject of the following chapter.
Chapter IV
Coupled Atmosphere - Ocean Modeling of ENSO

4.1 Introduction

The preceding chapters have presented separate simulations of observed surface wind anomalies and SST anomalies during ENSO, in each case taking the other as fixed. Such calculations are important and necessary in the development of component models, but they do not permit one to address some of the most fundamental questions concerning ENSO. Among them are the following. What starts an event, and what causes its termination somewhat more than a year later? Why do the events tend to be phase-locked to the annual cycle? What determines the time span between events, observed to be considerably variable but with a strong preference for three to four years? Is external forcing necessary to trigger events, or can they be explained solely in terms of tropical Pacific atmosphere-ocean interaction? Is it appropriate to consider ENSO a type of interactive instability, and, if so, what is the nature of the instability and what prevents it from occurring more frequently or more regularly? A related question — is ENSO predictable? These questions cannot be answered with forced calculations because the characteristics of part of the system are imposed, thus determining the response of the remainder of the system. Whether a coupled system would produce the same behavior cannot be determined in this way.

There have been very few interactive modeling studies to date. For the most part, until recently, they have fallen into two categories: those in which one or both of the components of the model are highly
parameterized, and those in which both components are complex GCM models. Studies of the first type include McWilliams and Gent (1978), Lau (1981), McCreary (1983), McCreary and Anderson (1984) and Nicholls (1984). Studies of the second type include Bryan et al. (1975), Wells (1979), and Washington et al. (1980). Very recently a third category of models has begun to appear, reflecting a philosophy similar to that adopted here. These are intermediate-level models in which the dynamics of the atmosphere and ocean are treated in a fairly simple, but consistent fashion. The studies of this type include Philander et al. (1983), Anderson and McCreary (1984), and Gill (1984). Several of the above will be discussed subsequently in the context of the present results.

This model differs from the other recent intermediate-level models primarily in its treatment of the thermodynamics in the atmosphere and ocean, especially through the inclusion of a moisture feedback process in the atmosphere and a thermodynamically active surface layer in the ocean. In both cases the formulations have been kept simple. They are a first attempt to represent explicitly the thermodynamic processes that are thought to be as important as the dynamics in this problem. The results of the forced calculations have supported this view.

4.2 Computational Procedures

The coupled model is completely determined by the component models as described in Chapters II and III. Using the atmosphere component, the wind stress is calculated at each timestep, and used to integrate the ocean model forward to the next timestep. The atmosphere affects
the ocean only through the action of the surface wind stress, and the ocean affects the atmosphere only through the effect of the SST anomaly field.

The grid spacing is the same as before: $\Delta x = 5.625^\circ$ latitude, $\Delta y = 2^\circ$ latitude for the atmosphere and for the SST anomaly field, and $\Delta x = 2^\circ$ latitude, $\Delta y = 0.5^\circ$ latitude for the remainder of the ocean variables. As a result of the different grid lengths a numerical problem can arise in the calculation. If the ocean develops small scale features, these features can grow locally in a fashion inconsistent with the governing equations, which predict that such features should decay. As an example, consider a warm SST anomaly away from the equator. The atmospheric response to a heating of this form is basically a cyclonic flow about the source (see Chapter II). The oceanic surface layer flow induced by this wind field is divergent (see Chapter III), implying upwelling and a reduction of the initial temperature anomaly, rather than growth. The problem is that the atmosphere numerical model can give locally incorrect results when the forcing has grid-scale oscillations. To insure accurate results, the forcing should not have variations on a scale smaller than two to three grid lengths (about $5^\circ$ in this case). The problem is avoided here by applying a 7-point binomial filter to the calculated heating field before calculating the wind field response. This effectively eliminates those scales for which numerical aliasing occurs, while leaving the large-scale features unaffected.

The timestep for the calculations is as before: 10 days for the dynamical fields, and 2.5 days for the SST anomaly field. Because
the SST anomaly field can change on a timescale of days in these calculations, it is necessary to re-examine the assumptions of the atmosphere model. It was assumed that the atmospheric response, including the convergence feedback adjustment, occurs on a timescale which is short compared with that of the SST variation. While reasonable for the earlier forced calculations using seasonal mean SST anomalies, this cannot be expected to hold here. In the real atmosphere, the final response to an imposed surface temperature anomaly would be expected only after a time equal to the advection time between the SST anomaly region and the downstream region of mean convergence (where the heating is actually realized). In the case of an eastern Pacific SST anomaly this advection time can easily approach one month.

With a ten day timestep, then, the assumed separation of timescales between atmosphere and ocean breaks down. If this is ignored, the result for the coupled model is easy to anticipate. The change in atmospheric response at each timestep will be overestimated by the assumption that the response is continuously in steady-state balance. As a result, more rapid changes in SST will occur because of local wind effects. This induces yet larger changes in the atmosphere, and the combined interaction favors an artificially rapid development of anomalies, particularly at small scales where the atmospheric convergence feedback is most efficient.

There are several conceivable ways to deal with this problem. Of course one approach would be to make the atmosphere model time dependent with a moisture budget. Short of that there are a couple of alternatives. One is to recalculate the wind anomaly field less frequently (e.g.,
once per month). The other is to modify the feedback parameterization to reflect a finite atmospheric adjustment time. The former procedure is undesirable because it prohibits atmospheric changes on short timescales even if the forcing is independently changing rapidly. The latter alternative is less restrictive, and therefore was chosen here. An upper limit of three iterations is imposed on the feedback adjustment, which is carried out at each timestep. If the procedure converges to less than the specified tolerance in less than three iterations, it terminates as before. The total wind anomaly field is recalculated once per month. At the interim timesteps the wind anomaly field is incremented according to the change in the heating field for that timestep, using the same algorithm (see Appendix D). However, because the same tolerance is used for the feedback calculation in these cases, only large changes in the heating field produce significant feedback on a timescale shorter than one month. By restricting the feedback in this way, the total wind anomaly field tends to approach its steady-state configuration on timescales of one month or greater, while also being sensitive to large changes in forcing on shorter timescales.

In the actual calculations, the feedback iteration generally converges at or before three iterations, so that the imposed upper limit does not affect the results. In cases where it does, however, the result is to limit the rapid growth of small-scale features in the response. These are the features most enhanced by the convergence feedback, as described in Chapter II.

Other methods could be used to produce a similar result. For example, a spatial smoother could be applied after each iteration.
This also would limit the artificial growth of small scale features which results from assuming a continuously steady-state atmospheric response. The present method was chosen because it requires less computation. The savings approaches 50% in many cases. This difference is of course substantial for extended model runs.

4.3 Coupled Experiments: Simulation Run

In this calculation, the parameter values were all taken directly from the forced calculations. Thus the atmospheric parameters are as for model II of Chapter II, and the oceanic parameters are as in Chapter III. The imposed initial condition for the calculation has the form

\[ u_a = 2.0 \text{ m/sec} \left( \exp \left( -\frac{y^2}{40^0} \right) \right), \quad \text{for} \quad 145^0 \text{E} < x < 170^0 \text{W}, \]  \[ \text{(4.3.1)} \]

where \( y \) is measured in degrees of latitude from the equator. This imposed anomaly in zonal wind is applied for a period of four months (between months 0 and 4 in the calculation) and then removed. (Months 0, 12, 24, ... refer to December of year (-1), (0), (1), ..., respectively.) There is no other imposed forcing.

Figure 4.1 shows a record of the indices NINO1, NINO3 and NINO4 for 90 years of simulated time. Certainly the most striking result is the recurrence of warm events, deriving solely from self-interactions of the coupled system. After the first, rather weak warm event in year (0) which results from the imposed initial condition, the system exhibits recurring oscillations which tend to favor a 3 to 4 year period. The oscillations appear at times to be quite regular in amplitude and
Fig. 4.1 Coupled model predictions for NINO1 (heavy solid line), NINO3 (light solid line), and NINO4 (dotted line), during 90 years of simulated time.
structure, while at other times they become highly non-uniform, with widely varying amplitude and inter-spacing. The duration of the sizeable warm events, however, does not vary significantly, tending always to be between 14 and 18 months. It is clear that the warm events are very closely tied to the annual cycle, always peaking either in June or around the end of the year. The amplitudes of the larger events exceed 2°C in the eastern Pacific (actually they exceed 3°C near the coast in these cases), with more modest events having amplitudes of 1.0°C-1.5°C. The largest temperature anomalies occur furthest east, i.e., NINO1 displays the largest oscillations, although NINO3 typically is only slightly smaller. In the larger events, the SST anomalies peak first in the east (around November) and later toward the west (around January).

These features are reminiscent of the observations and the previous forced calculations. Consider first the larger warm events of the coupled model. All of these have a very similar structure, with rapidly increasing eastern ocean temperature anomalies starting in late spring and continuing until nearly the end of the year, followed by a moderate decrease until the following June and then a precipitous decline during the summer. Similar results were found in the composite-forced calculations (see Figure 3.6), except that the rapid temperature decrease tended to occur earlier there, resulting in a slightly shorter warm event duration (by two to four months). In the growth phase, the coupled model does not give a large coastal SST anomaly maximum in May–June of the ENSO year as is often observed, but this is not surprising since
the composite-forced calculation also failed to do so. Suggestions were made earlier as to the possible cause of this discrepancy.

The inter-event periods of the coupled model bear close resemblance to those of Figure 3.6, which were constructed from the composite data. The amplitudes of the negative SST anomalies appear to be somewhat larger in the coupled model. However, it is noteworthy that the long-term mean of all the coupled model indices is very nearly zero. Although one expects this to be the case for a relevant perturbation model, there is nothing in the model formulation which requires it.

Some of the smaller coupled model warm events have a different structure from the larger events just described. They tend to develop slowly for a whole year, starting in June and peaking in the following June, with a slow decline in the following six to eight months. These events do not appear to correspond to either the canonical case scenario or that of the 1982–3 event. However, a number of small amplitude anomalies which do not conform to the usual pattern of the larger events are suggested by the data. Figure 4.2, reproduced from RC, shows a time series of SST anomalies along two shipping lanes in the equatorial eastern and central Pacific. A number of small events do not correspond well with the larger ones. It is intriguing that the observations show extended periods without any major events (the late 1920's and the 1930's) and other periods with fairly regular and large events (the 1950's and 1970's). The model clearly displays such regimes as well.

Figure 4.3 shows the equatorial tradewind indices TW1 and TW2 for the model run. The temporal characteristics of these indices are,
Fig. 4.2 Indices of eastern and central Pacific SST anomalies, computed from observations (after Rasmussen and Carpenter, 1982).
Fig. 4.3 Coupled model predictions for TW1 (heavy line), and TW2 (light line), during the 90 year simulation run.
of course, closely linked to those of the temperature indices. The largest events give TW2 (eastern Pacific) anomalies which exceed the TW1 (central Pacific) anomalies. This is in contrast to the observations, but in agreement with the forced calculations of Chapter II. As discussed there, the atmosphere model tends to overestimate the eastern Pacific wind anomalies, and underestimate the central and western Pacific anomalies. For most events the two indices tend to peak at about the same time, as was the case in the forced calculation. In neither calculation is the observed development of western Pacific easterlies (and associated initial decrease in TW1) captured. There is little doubt that this is responsible for the delayed termination of the warm events in the coupled model, as compared to the observations. Finally, notice that during all warm events the two wind indices are nearly in phase, and during all extended periods without events, the indices are out of phase.

A more complete picture of the evolution of model SST and wind anomalies during an oscillation of the system is displayed in Figure 4.4, which traces the development in three-month intervals between the end of year (30) and the end of year (35). This period is characterized by a large warm event in year (31) and a small event in years (34) and year (35).

The sequence begins in December of year (30), at which time there are no appreciable anomalies either in SST or wind. By March (31), a region of warm SST anomaly has developed in the equatorial zone east of 170\(^\circ\) W, with a maximum near 130\(^\circ\)W. Associated with this are small westerly wind anomalies in the region 130\(^\circ\)W to 160\(^\circ\)W. The warm event
is well underway by June (31), with SST anomalies exceeding 1°C in the eastern equatorial Pacific and sizeable (~ 1 m/sec) westerly wind anomalies in the central Pacific. The development so far is more like that of the 1982-3 ENSO event than the canonical event because the eastern ocean tends to warm uniformly rather than first at the coast and then to the west.

The observed tendency for the westward spreading and amplification of both SST and wind anomalies in the fall of an ENSO year occurs also in the model. By September (31) warm anomalies extend as far westward as 160°E, and eastern Pacific anomalies exceed 2°C. Large westerly wind anomalies cover the whole equatorial central Pacific, with equatorward flow across the normal position of the ITCZ. The large easterly anomalies in the eastern Pacific, although not realistic, are the same as were obtained in the forced calculations. The region of small negative SST anomaly and easterly wind anomaly which has developed in the western Pacific at this time is also found in the RC composites.

The peak temperature anomalies occur in December (31), with a maximum at the coast and another one near 140°W. By December the SST anomalies also have expanded meridionally, compared with the preceding patterns. These features are realistic (see figure 3.11), except that the coastal maximum is exaggerated, as it was in the forced calculation. Westerly wind anomalies approach 2 m/sec in the central Pacific at this time. The observations show the development of easterly anomalies in the western Pacific at this time (see Figure 2.12). Such a feature does not develop in the model either at this time or in the immediately ensuing months. As mentioned above, the delayed termination of the
ensuing months. As mentioned above, the delayed termination of the
model warm event can be traced to this.

By March (32), temperature anomalies have begun to decrease, especially
at the east coast. A single maximum now exists in the eastern central
Pacific. The pattern is quite similar to the composite event for this
time, although the amplitude of the warm anomaly is about a factor
of two larger than the composite. Very large westerly wind anomalies
persist in the central Pacific, with increasing easterly anomalies
further to the east.

In June (32) the eastern ocean is still warm, though it is decreasing
rapidly. The westerly wind anomalies have decreased and receded westward,
and stronger easterlies are evident in the east. The patterns more
resemble those for 1983 than the composites for June following the
ENSO year. By this time, the composites show cold SST anomalies and
poleward wind anomalies in the eastern ocean.

During the summer of year (32) a dramatic change occurs in both
winds and SST, amounting to a rapid termination of the warm event.
By September (32) the equatorial eastern and central ocean is cold,
and the winds are primarily meridional and directed poleward. The
temperature pattern is not unlike that of the composite for this time,
which also shows an equatorial tongue of cold anomaly extending across
much of the basin.

The next year is characterized by cold SST anomalies and equatorial
easterly wind anomalies of varying magnitude. The equatorial easterlies
tend to be strongest in the western and central Pacific, except during
March and April, when they also occur in the eastern ocean. It is
during March and April that the region of mean convergence moves onto the equator in the eastern Pacific, allowing the convergence feedback to enhance the effect of cold SST anomalies. At other times there is no feedback, so that even in the presence of large negative temperature anomalies such as in September (33), the equatorial wind response is very small.

In December (33), a small region of positive SST anomaly appears in the central Pacific, the first sign of the rather peculiar warm episode that is to follow. The equatorial wind anomalies are weak easterly everywhere at this time, with minimum magnitude in the vicinity of the warm anomaly. The warm anomaly grows by March (34), and is then centered at the dateline, with weak westerly wind anomalies to the west. This configuration of temperature and wind anomaly is very reminiscent of that which precedes the warm event in the RC composites (see Figure 2.10), although considerably weaker.

By June (34), the wind and temperature anomalies have dissipated in the central Pacific, but have started to develop in the eastern Pacific. They continue to grow and expand westward until December (34), by which time the familiar mature phase pattern of anomalies is reached, although the amplitudes are still small.

After starting to relax as usual in March (35), the anomalies amplify again and reach their maximum magnitude in June (35), after which they dissipate rapidly during the summer and fall. By December (33) cold anomalies have developed in the eastern equatorial ocean, westerly wind anomalies have disappeared, and the coupled system is in its negative phase again.
Fig. 4.4 Sequence of model SST anomalies (left panels) and corresponding wind anomalies (right panels), during the 90 year simulation, starting with December of year (30) and proceeding to December of year (35).
Fig. 4.4 Cont'd.
While resembling observed patterns of SST and wind anomalies in its different phases, the evolution of the second, weaker warm event does not follow that of the larger model events or the larger observed events. As pointed out earlier, it is difficult to say that similar sequences do not occur in reality. First of all, small anomalies may not be detected reliably in the data. Secondly, the time records that are available indicate low frequency variability that differs substantially from the canonical ENSO scenario. It may be that only moderately large events follow the sequence depicted in RC, and that smaller (and larger) events can behave somewhat differently.

Another important element of the coupled system oscillations is the oceanic mass exchange along the equator. Figure 4.5 displays the thermocline depth anomaly $h(x, t)$ along the equator, together with $b(x, t)$ (a measure of the equatorial wind stress anomalies as described in [3.5.1]) for years (30) to (45) in the model run. During warm episodes $h$ is positive in the eastern ocean and negative to the west for a period of about a year. Immediately following a warm event, $h$ is generally negative across the whole basin, but becomes positive shortly thereafter in the west. The ensuing period is usually characterized by a succession of such positive upper layer depth (ULD) anomalies which seem to originate near the west coast and propagate progressively further eastward, until $h$ becomes positive across the whole basin (and particularly in the east). At this point another warm event generally occurs. This progression can be seen in the periods between month (360) and month (380), between month (400) and month (420), and between month (480) and month (500).
Fig. 4.5 Equatorial h-anomalies (left panels), and b(x,t) (right panels) for the period between month 360 (year 30) and month 540 (year 45) of the simulation run. Contours for h are 5m in the lower two panels, and 10m in the upper panel. Contours for b are 5 non-dimensional units in the upper two panels, and 10 units in the lower panel.
Somewhat similar features in the equatorial h-field were obtained in the forced calculation using the RC composite winds (Figure 3.7).

The plots of b show strong equatorial stress anomalies in the central Pacific during warm events, with weaker easterly anomalies to the east. Between warm events, a very different type of behavior is seen: easterly stress anomalies appear to develop in the eastern Pacific and propagate westward, reaching the western end of the basin in a period of approximately 10 months. The arrival of these features in the western Pacific coincides exactly with the appearance of the positive ULD anomalies described above. At times westerly anomalies appear in a fashion similar to that prior to a warm event, but the development quickly terminates (as during year (38) and year (44)). The difference between these cases and those that have warm events seems to be in the presence of easterly anomalies in the eastern Pacific. In the aborted event cases, easterly anomalies exist in the east at the time the westerly anomalies appear further west. As the westerlies start to grow, the easterlies do so as well, and shortly thereafter the development ceases. Preceding warm events, on the other hand, there are no significant easterlies in the east, either before the appearance of westerly anomalies or during their growth. The development of easterly anomalies can in turn be traced to the time of year that the would-be warm event is getting established. In each case of a terminated event, the event starts to develop in the early part of the year (January - March). On the other hand, the substantial warm events develop later (April - June). This suggests that the annual cycle exerts considerable influence over the development or non-development
cycle exerts considerable influence over the development or non-development of warm events.

The relative roles of zonal and vertical advection in producing the fluctuations in NINO1, NINO3, and NINO4 is examined in Figure 4.6. It is of interest whether the balance of terms in the coupled model is similar to that for the forced calculations. As seen in Figure 4.6, the balances are essentially the same as before. In the NINO1 region, the mean upwelling term is largest, but anomalous upwelling and zonal advection vary in phase with the mean upwelling. In the NINO3 region, zonal advection is dominant, with anomalous upwelling also contributing during significant warm event periods. Mean upwelling generally opposes the temperature anomalies for this region as a whole. For the NINO4 region, the net effect of vertical advection is negligible, and zonal advection alone is responsible for the temperature variability there.

In many respects, the results of this coupled calculation compare very favorably with observations and with the forced calculations. It is therefore worthwhile to try to understand in greater detail the mechanisms determining the model behavior. To the extent that this can be done, hypotheses can be formed concerning many of the important questions surrounding ENSO events. However, because of the complexity of the model (despite its many simplifications), it is not straightforward to analyze the behavior. The following section presents what has been found to be the most fruitful approach: a series of comparison calculations chosen specifically to examine the importance of particular physical processes.
Fig. 4.6a Contributions to the NINO1 variability between years 31 and 60 of the 90 year simulation run. Codes are the same as for Fig. 3.14.
Fig. 4.6b As in Fig. 4.6a, except for NINO3
Fig. 4.6c  As in Fig. 4.6a, except for NINO4.
4.4 Diagnostic Coupled Runs

A large number of coupled model calculations have been run. The results of several of them will be summarized here. The experiments were designed to investigate parameter sensitivities and to isolate, if possible, mechanisms controlling the timescales and spatial structure of the oscillations. Each run proceeds for a period of 30 years and is identical to the simulation run with the exception of what is listed below:

Run A: the effect of vertical advection in producing SST anomalies is reduced by decreasing $\gamma_1$ and $\gamma_2$ from .75 to .4 in [C.7].

Run B: the surface heat flux is decreased by a factor of two (the decay time is increased from ~4 months to ~8 months).

Run C: the oceanic free wave speed is decreased by 25%.

Run C2: the oceanic free wave speed is increased by 12%.

Run D: the annual cycle is removed by assuming continuous May conditions.

Run D2: as in Run D but additionally $\gamma_1$, is reduced to .5 in [C.7].

Run E: the effect of subsurface temperature anomalies is ignored, i.e., $T_{\text{sub}} = 0$ in [C.6].

Run F: the oceanic surface layer is removed.

Run G: the effect of all anomalous currents is removed.

Run H: the effect of the mean upwelling component of vertical advection is reduced by setting $\gamma_1 = .5$ in [C.7].

Run I: the initial condition for the run is changed by imposing the westerly wind anomaly between May and September rather than between December and April.

Run J: the atmospheric convergence feedback procedure is modified by doing exactly two iterations at every timestep.
Following is a brief presentation of each run. A more comprehensive interpretation of all the results will be offered in section 4.5.

4.4.1 Run A

This run examines the importance of upwelling in the coupled model. With reduced $\gamma_1$ and $\gamma_2$, SST is less sensitive both to subsurface temperature anomalies and to local wind stress. The former holds because $\gamma_1$ modulates the effective vertical temperature advection due to the mean upwelling, the mechanism by which subsurface temperature anomalies are transferred to the surface. The latter holds because $\gamma_2$ modulates the effective vertical temperature advection due to the anomalous upwelling, which primarily derives from the local wind stress as discussed in Chapter III.

Figure 4.7 shows the three temperature anomaly indices and the two equatorial wind anomaly indices for the run. The reduced influence of the wind field on SST has a dramatic effect. Within two months after the imposed wind anomaly is removed, temperature anomalies begin to decrease rapidly, becoming negative early in year (1). Thereafter the system exhibits very weak and rapidly diminishing oscillations. It is important to notice that even the initial, forced warm event does not resemble observed events in duration or amplitude. The ensuing oscillations eventually develop a three year period. This is clear in Figure 4.8, which shows $b(x,t)$ and $h(x,t)$ along the equator during the last ten years of the run. Although steadily decreasing in amplitude, the oscillations have a very uniform structure which repeats every three years. Periods of westerly anomalies in the central ocean are
Fig. 4.8 $b(x,t)$ (left panels) and the equatorial h-field (right panels), for the period between year 20 (month 240) and year 30 (month 360) of RUN A. Contours for $h$: .02m(lower), and .01m (upper). Contours for $b$: .02 units(lower), and .01 units(upper).

Fig. 4.7 Upper panel: NINO1 (heavy solid line), NINO3 (light solid line), and NINO4 (dotted line) for year 0 to year 30 of RUN A. Lower panel: TW1 (heavy line), and TW2 (light line) for the same period of RUN A.
shorter than in the simulation run, whereas the inter-event periods are longer but look similar, with westward propagating easterly anomalies. The propagation speed of these features is somewhat smaller than in the simulation run. Associated with these propagating features are positive ULD anomalies at the west coast as was found before.

Although it was found that zonal advection was dominant in much of the ocean for the simulation calculation, these results indicate that the effect of vertical advection is not unimportant. Without the sources of surface temperature anomalies associated with vertical advection, zonal advection alone can produce only very small anomalies. The establishment of a three year oscillation suggests an influence of the annual cycle, and indicates that the tendency for interannual variability may not depend critically on the strength of atmosphere-ocean coupling.

4.4.2 Run B

In this case the effective coupling is increased by decreasing the thermal dissipation of SST anomalies. With less thermal dissipation, a given wind forcing produces a larger temperature anomaly signal, which then forces a larger wind response, etc. As seen in Figure 4.9, the effect is enormous. Eastern Pacific temperature anomalies approach 8°C, and eastern Pacific westerly wind anomalies exceed 5 m/sec. Aside from being unrealistically large in amplitude, the warm events tend to have a duration of two years or more, rather than one year as observed. The period between events is quite variable, ranging from two years to six years, with an apparent preference for the longer periods.
Figure 4.10 shows b(x,t) and the equatorial h-field for the last ten years of the run. The long duration of the stationary westerly wind pattern during warm events is clear. The termination of this warm event pattern is exceedingly rapid, and occurs from east to west. Following is typically a sequence of westward propagating, rather small-scale wind features as has been observed in the other runs. The equatorial ULD anomalies are of course much larger than before, but follow a similar pattern preceding a warm event: i.e., a sequence of positive anomalies at the west coast which appear to propagate eastward. During warm events there is a nearly stationary pattern of positive ULD anomalies in the east and negative anomalies in the west, as seen in the other cases.

As opposed to the simulation run, the wind and temperature anomalies here clearly have non-zero mean, i.e., in the eastern ocean there are mean westerly wind anomalies and mean positive SST anomalies. This suggests that the smaller thermal dissipation makes the model less relevant to the real system. The amplitude, duration and spatial structure of the oscillations indicate likewise.

4.4.3 Run C

This calculation was designed to examine the importance of free waves in producing the interannual variability. With the Kelvin wavespeed used in the simulation run (2.2 m/sec), it takes almost exactly nine months for a Kelvin wave to travel eastward across the basin and the fastest Rossby mode to travel back. It is possible that the combination of this period and the annual period may give rise to interannual
Fig. 4.10 As in Fig. 4.7, except for RUN B.
Contours for h and b: 20m and 2G units, respectively.

Fig. 4.9 As in Fig. 4.7, except for RUN B.
variability. If the interaction were nearly linear, then maximum amplification would occur at intervals equal to the least common multiple of the two periods — in this case, three years. This was found to be the dominant period in the simulation run.

In the present calculation, this theory is tested by adjusting the wave speed so that the free wave period is one year rather than nine months. If the linear theory applies, there should be essentially no interannual variability in this case.

Figure 4.11 shows that the results do not conform to this theory. Not only are there interannual oscillations, but their timescales are larger than before. Warm events tend to last for nearly three years, and repeat in five to six year intervals. The amplitude of the oscillations is large, with SST anomalies generally exceeding $4^\circ$C at the coast and westerly wind anomalies of nearly 4 m/sec in the eastern equatorial region. In many respects the events resemble the larger ones of run B, although the duration is even longer in this case. Notice also that there are still sub-annual oscillations in the indices, showing that there is another source for such frequencies besides free wave propagation.

Figure 4.12 shows $b(x,t)$ and the equatorial $b$-field for the final 10 years. As in the previous runs, the stationary westerly wind pattern during warm events and propagating easterly wind features between events are clear. The westward propagation speed is about the same or slightly greater than that of the simulation run. Clearly these features are not associated with free wave propagation in the ocean. The mechanism responsible for these features is apparently the source of the sub-annual oscillations in the model indices.
Fig. 4.12 As in Fig. 4.8, except for RUN C. Contours for h and b: 20m and 20 units, respectively.

Fig. 4.11 As in Fig. 4.7, except for RUN C.
4.4.4 Run C2

This calculation complements the previous one. In this case the wave speed was increased, such that the free wave oscillation period is reduced from nine months to eight months. The linear theory for interaction with the annual period would suggest a two year oscillation for this case. Figure 4.13 shows the indices for the run. The oscillations are weaker, shorter in duration, and more frequent than in the simulation run, preferring a two to three year period. The TW1 anomalies typically exceed the TW2 anomalies, as was the case for the small events in the simulation run. In contrast, the very large events of runs B and C always had TW2 >> TW1 (which disagrees with observations of any real event).

The usual plots of b and the equatorial h field are shown in Figure 4.14. The propagating features generally move more slowly than in run C, and in some instances they become almost stationary (as in year (21) and year (26)). In such cases the h-field shows a smooth transition between negative and positive values in the west, rather than the pulse-like behavior found in other cases.

4.4.4 Run D

This run examines more closely the importance of the annual cycle in the model. Continuous May conditions are imposed throughout the integration (the choice of May as opposed to another month is arbitrary). The results are shown in Figures 4.15 and 4.16. Clearly, interannual oscillations still occur. In this case they are large in amplitude and almost perfectly periodic, with period 38 months.
Fig. 4.14 As in Fig. 4.8, except for RUN C2. Contours for h: 
1m(lower), 2m(upper). Contours for b: 2 units(lower), 1 unit (upper).

Fig. 4.13 As in Fig. 4.7, except for RUN C2.
As usual with large events, the large wind anomalies are strongly focused in the east. In some respects, the oscillations appear similar to those in runs where the annual cycle is included. However, an examination of $b(x,t)$ reveals some important differences. The development of westerly anomalies during warm events is very smooth and monotonic. As the westerlies grow, so do the easterlies further to the east. Because the mean winds are easterly, the easterly stress anomalies grow faster than the westerly stress anomalies, and eventually $h$ starts to decrease near the east coast. At this point the coastal temperature anomalies start to decrease (NINO3 exceeds NINO1 at this time), causing a westward displacement of both the temperature anomaly maximum and the easterly anomalies that occur to its east. The process continues rapidly, bringing about the termination of the warm event. This is not the way events terminate in the simulation run, where easterly stress anomalies are always much smaller than the westerly stress anomalies during the major warm events.

The 'negative phase' of the oscillations in this case look rather similar to many of the other runs. There are westward propagating easterly anomalies with associated ULD anomalies at the west coast which appear to propagate eastward. These features do not fundamentally depend on the annual cycle.

4.4.6 Run D2

This case is the same as run D, except the mean upwelling effect is reduced by changing $\gamma_1$ from .75 to .5. The amplitudes in run D were large because in May the eastern Pacific mean upwelling was at
Fig. 4.16 As in Fig. 4.8, except for RUN D. Contours for h and b: 10m and 10 units, respectively.

Fig. 4.15 As in Fig. 4.7, except for RUN D.
its maximum, so that the sensitivity of SST to $h$ is continuously at its maximum value. A reduction in $\gamma_1$ offsets this, while at the same time examining the model sensitivity to $\gamma_1$. The results of the calculation are shown in Figures 4.17 and 4.18. As expected, amplitudes are less. In addition, warm events are more frequent (between two and three years apart) and generally shorter in duration. There is a nearly perfect periodicity with period 85 months. The events tend to terminate as in run D, although the easterly stress anomalies are somewhat smaller here. Again, the behavior during the inter-event periods is similar to the earlier runs.

4.4.7 Run E

In this calculation subsurface temperature anomalies are not allowed to affect SST. In other words, the coupling between the atmosphere and the oceanic $h$-field is eliminated (see [C.5]). Thus SST anomalies are created only through the advection of mean temperature gradients by anomalous currents in the surface layer. Once surface temperature anomalies exist, however, they can be advected by mean (horizontal) surface currents. The attempt here is to determine the nature of interaction between the atmosphere and just the surface layer of the ocean, and to see whether this behavior is identifiable in the full model. Note that the surface currents include a component due to baroclinic motions, so that SST anomalies can be generated remotely as well as through strictly local wind forcing.

The indices are shown in Figure 4.19, and the plots of $b$ and equatorial $h$ in Figure 4.20. The results are strikingly different from any of
**Fig. 4.18** As in Fig. 4.8, except for RUN D2. Contours for h and b: 10m and 10 units, respectively.

**Fig. 4.17** As in Fig. 4.7, except for RUN D2.
the preceding runs. The dominant period in the time series is no longer interannual, but sub-annual. It is slightly greater than nine months. There is, however, an interannual component to the variability, with an exact periodicity at seven year intervals. The events are very short and weak, with exceedingly rapid termination. From Figure 4.19 it is clear that variations in NINO3 lead those in NINO1 and NINO4 by a few months. Correspondingly, changes in TW2 (eastern region) precede those in TW1 (central region).

The plots of b show that westerly anomalies develop to some degree every nine to ten months, and then are wiped out by westward propagating easterly anomalies similar to those found in other runs. Depending on the phase of the oscillations with respect to the annual cycle, anomalies achieve variable amplitude. This run shows the kind of behavior sought earlier, i.e., interannual variability due to the presence of two timescales. In this case they are the annual period and the nine to ten month period. Maximum amplification occurs when the two cycles have a particular phase relative to each other.

The equatorial h-field is oscillatory, with fluctuations in the eastern and western ends of the basin nearly out of phase. Large positive ULD anomalies appear at the west coast with the arrival of each westward propagating region of easterly wind stress, and then appear to propagate eastward without interruption, causing positive ULD anomalies in the east two to three months later. The behavior is very similar to that found during the 'negative phase' of oscillation in the full model, except there the positive ULD anomalies typically dissipated as they propagated eastward, producing a smaller effect in the east. On the
Fig. 4.20 As in Fig. 4.8, except for RUN E. Contours for h and b: 10m and 10 units, respectively.

Fig. 4.19 As in Fig. 4.7, except for RUN E.
other hand, the behavior here does not at all approximate that found in the warm event phase of the full model. The suggestion is that the warm event phase depends on the h-coupling which is ignored in this run.

Figure 4.21 shows the wind and SST anomaly fields for December (20). These maps are typical for the run in terms of the scale of the anomalies, which is \( \sim 40^\circ \) latitude in \( x \) and \( \sim 10^\circ \) latitude in \( y \). With such scales the anomalies are rather focused near the equator and generally of opposite sign in the eastern and central regions. Such spatial structure, of course, is unlike that characteristic of the full model.

4.4.8 Run F

This case examines the importance of the surface layer by omitting it. Thus all anomalous currents are produced solely by the baroclinic motions of the upper ocean, but \( h \) affects SST as in the full model.

Figure 4.22 shows the indices for the run, which have several unique features. First of all, oscillations grow and decay much more slowly than before. Secondly, there is essentially no sub-annual variability in the SST indices, in contrast to all previous cases. Also, the oscillations are regular (but not perfectly periodic) in five year intervals, which has not been found before. All of the indices vary nearly in phase, with negative departures looking more similar to positive ones than in other runs. In the western and central region amplitudes
Fig. 4.21 (a) SST anomalies, and (b) wind anomalies, for December of year 20 in RUN E.
are very small, whereas in the east they are more comparable to the simulation run.

The plots of \( b \) and the equatorial \( h \)-field are shown in Figure 4.23. For the first time, there are no pronounced westward propagating wind features. During the negative phase the easterly \( h \)-field anomaly pattern looks rather similar to the westerly pattern during warm events, except smaller in amplitude. The equatorial \( h \)-field shows a rather smooth, slow transition back and forth between warm \( h \)-field configuration and 'cold event' configuration.

It is apparent that without the surface layer the quasi-stationary warm event pattern of anomalies is still reproducible, although only with a slowly varying temporal structure. On the other hand, the behavior observed during the negative phase of the full model must owe its existence to surface layer-induced effects. Finally, the absence of sub-annual variability in this run demonstrates again that free mode oscillations in the ocean are not the ultimate source for such variability in the full model, but that the source derives rather from interactions between the atmosphere and the oceanic surface layer.

4.4.9 Run G

In this calculation current anomalies are ignored altogether, so that SST anomalies can be produced only by subsurface thermal anomalies associated with variations in \( h \). Horizontal advection by the mean surface currents can still modify the SST anomalies once they are created. This run is in a sense a more extreme version of the previous one. There the near-surface current anomalies were significantly reduced
Fig. 4.23 As in Fig. 4.8, except for RUN F. Contours for h and b: 2m, and 2 units, respectively.

Fig. 4.22 As in Fig. 4.7, except for RUN F.
by eliminating the surface layer. Here they are ignored altogether. The calculation isolates the interaction between atmosphere and ocean which involves thermal anomalies of the entire upper ocean. It is the complement of Run E, which included all the source mechanisms for SST anomalies except those due to variations in $h$.

Figure 4.24 shows the indices for the run, and Figure 4.25 the usual $b$ and equatorial $h$-plots. The timescale of the oscillations has increased greatly, to nearly a decade. The nine year quasi-periodicity is characterized by very slow growth and decay of anomalies, and almost no amplitude in the central or western regions. There is no sub-annual variability in the SST indices, but there is a clear, regular, annual component. The plots of $b$ show that the 'cold event' structure in the equatorial wind field is now very much like the warm event structure, with no evidence whatever of propagating features. The equatorial $h$-field undergoes very smooth and symmetric transition between its warm event and cold event configurations.

Figure 4.26 shows the wind and SST anomaly fields for June (13) and June (18), representing opposite phases of the cycle. The trapping of the anomalies in the eastern ocean and the symmetry between positive and negative states is clear.

4.4.10 Run H

Runs E, F and G indicate that the coupling between the atmosphere and the oceanic $h$-field tends to favor very low frequency, stationary oscillations and that the coupling between the atmosphere and the oceanic surface layer favors higher frequency, propagating disturbances. This
Fig. 4.25 As in Fig. 4.8, except for RUN G. Contours for h: 1m(lower), 2m(upper). Contours for b: 2 units.

Fig. 4.24 As in Fig. 4.7, except for RUN G.
Fig. 4.26 (a) SST anomalies, and (b) wind anomalies, for June of year 13 in RUN G. (c) SST anomalies, and (d) wind anomalies, for June of year 18 in RUN G.
notion is tested in a different fashion in Run H. Here all mechanisms are included as in the simulation run, but $\gamma_1$ is reduced from .75 to .5. Since $\gamma_1$ modulates the effect of $h$ on SST, the coupling between the atmosphere and $h$ is reduced in this run. According to the above theory, the results should look more like Run E where $\alpha$ was ignored, with more high frequency variability and less evidence of stationary anomaly patterns. As seen in the indices of Figure 4.27 and the plots of $b$ and $h$ in Figure 4.28, this is indeed the case. Compared with the simulation run, the time series are more rapidly varying (it is especially clear in the wind anomaly indices). The behavior is more like that of Run E in terms of temporal structure as well as the relative amplitudes of central and eastern Pacific anomalies. Central Pacific anomalies are substantial during warm events in the simulation run. They are much smaller in this run, and yet smaller in Run E.

The $b$ and equatorial $h$-fields for the run are also intermediate between the simulation run and Run E. As in Run E there are westward propagating easterly anomalies which originate in the east at fairly regular intervals (less than one year), and westerly anomalies which develop and dissipate in the eastern central region during the interim times. Sustained periods of westerly anomalies such as found during the larger events of the simulation run are not found here, nor are the longer warm event patterns in the equatorial $h$-field. On the other hand, anomalies are usually longer-lived than those of Run E.
Fig. 4.28 As in Fig. 4.8, except for RUN H. Contours for $h$ and $b$: 10m, and 10 units, respectively.

Fig. 4.27 As in Fig. 4.7, except for RUN H.
4.4.11 Run I

Although the dissipation mechanisms in the model virtually guarantee that the initial conditions do not substantially affect the characteristics of the results at later times, this run addresses the issue by changing the initial condition. Rather than imposing the westerly wind anomaly between December and April, in this case it is imposed between May and September of year (0). The indices are shown in Figure 4.29. The behavior in subsequent years shows all the same characteristics as was found in the simulation run. In fact, the comparison between year (1) to year (20) of this run and year (61) to year (79) of the simulation run (Figure 4.1) is striking. Other runs have also confirmed that the results are insensitive to the initial conditions.

4.4.12 Run J

In this last case the effect of the nonlinear atmospheric feedback is examined. For this run, the number of iterations of the feedback scheme is set uniformly to two. Thus, in situations of either large amplitude or rapidly changing SST anomalies, there is less feedback response in this calculation than in the simulation run. The SST anomaly indices are presented in Figure 4.30. Overall, the oscillations are more regular and slightly longer in duration, the timescale between events now being four years. The warm events (after the initial forced ones) are similar to the weaker ones in the simulation run, which were found to be the least similar to observations. It appears that the restricted feedback has had two major effects: (i) a somewhat reduced rate of growth and decay of anomalies which is associated with a longer
Fig. 4.29 As in Fig. 4.7, except for RUN I.

Fig. 4.30 NINO1, NINO3, and NINO4 indices for RUN J.
preferred period, and (ii) less irregularity in the system, with oscillations of nearly constant amplitude at constant intervals. Both of these effects make the results less like observations. Perhaps the most significant indication is that the nonlinearity of the atmospheric response is an important factor in the nonuniformity of events in the model. Additional calculations which are not presented here further support this conclusion.

4.5 Interpretation of Coupled Model Results

The various coupled model runs present a rich variety of behavior. In terms of the spatial and temporal characteristics of the oscillations, the most realistic results were obtained in the simulation run, which used formulations and parameter values directly suggested by the earlier forced experiments. This demonstrates the value of the individual component-building approach, especially in light of the rather high model sensitivities that have been found. Given the limited success of the simulation model, the goal now is to understand the processes governing its behavior in order to develop hypotheses regarding the important physics of real ENSO events. Taken together, the results from the other coupled runs contribute much toward this understanding, while still leaving some unanswered questions.

The different calculations have pinpointed two distinct types of interaction between the model atmosphere and ocean. The first type involves coupling between the atmosphere and the oceanic surface currents, and is exemplified best in Run E (where $h$ is ignored). This interaction by itself produces rather small scale, mobile disturbances which propagate
westward, leading to temporal variability on relatively short timescales (less than one year). The second type of interaction involves coupling between the atmosphere and thermal anomalies in the entire upper ocean, and arises through the action of the mean oceanic upwelling circulations. This interaction produces large-scale, stationary anomalies with a fixed spatial pattern somewhat resembling that of ENSO events. By itself it leads to temporal variability on very long timescales (of the order of a decade), as exemplified in Run G (where anomalous currents are ignored). The different behavior in the model runs derives largely from the relative importance of these two types of interaction for each choice of model specifications.

The propagating coupled anomalies can quite easily be understood by considering the governing equations for the oceanic surface layer and the atmosphere in the proximity of the equator. The dominant contribution to anomalous upwelling at the equator arises from the $\beta$-term in [3.2.10] involving the zonal wind stress. Thus, approximately,

$$w_s = -\beta \tau(x) / \rho r_s^2.$$  \[4.5.1\]

From [3.2.5], the zonal current anomaly at the equator is

$$u_s = \tau(x) / \rho r_s H_s.$$  \[4.5.2\]

The change in temperature due to these current anomalies is (see [3.2.11])

$$\frac{\partial T_s}{\partial t} = \tau(x) \left[ \frac{\gamma}{2} \frac{T_x^2}{\rho r_s^2} - \frac{T_x}{\rho r_s H_s} \right].$$  \[4.5.3\]

(the total upwelling is assumed positive in this derivation)
For the atmosphere, assuming a symmetric heating about the equator, and in regions very near the equator, equations [A.21 - A.23] and [A.33] reduce to

\[ e^2 u_a - (N^2/m^2)(u_a)_{xx} = a^*(T_s)_x \]  \[ 4.5.4 \]

where \( a^* \) is a dimensional constant. The contribution due to convergence feedback has been ignored here (this can only underestimate the local response). If a mean constant surface wind \( U_0 \) is assumed, [3.4.1] gives the following approximate form for the surface stress anomalies:

\[ \tau(x) = 2\rho_a c_D U_0 u_a \] \[ 4.5.5 \]

Finally, taking \((U_a, T_s) = (\tilde{U}_a, \tilde{T}_s) e^{i(kx + \omega t)}\), \[ 4.5.3 \] - \[ 4.5.5 \] combine to give

\[ \omega/k = \frac{2\rho_a c_D U_0 a^*)(\gamma_2 B T s H_s - r_s T_x)}{(\rho c_s H_s)^2 e^2 + N^2 k^2/m^2)} \] \[ 4.5.6 \]

Thus, because of the mean temperature structure in the equatorial ocean \((\overline{T}_Z > 0, \overline{T}_X < 0)\), the disturbances propagate westward, and with a phase speed which varies as the coupling coefficients \( \gamma_2 \) and \( a \), and inversely as the dissipation coefficients \( r_s \) and \( s \). In all the calculations, \( a, r_s, \) and \( s \) were fixed, although in Run A \( \gamma_2 \) was reduced. The phase speed of the westward propagating features was found to be smaller than in the other cases, as predicted by the above analysis.
It is important to realize that these features derive essentially from the oceanic surface layer, and not the baroclinic motions of the upper ocean. In Runs F and G, where the effects of the surface layer are removed, there are no prominent westward propagating features. In all other runs, with and without the annual cycle, and for a variety of different parameter choices, the features are clearly identifiable during at least part of the oscillation cycle.

The stationary mode interaction involves the communication between the entire upper layer of the ocean and the atmosphere. It is this interaction which fundamentally defines model ENSO events in terms of their characteristic spatial structure. The structure arises because of the influence of the mean wind, temperature, and current fields. Consider first the mean upwelling field. This is the primary means through which \( h \) (and associated subsurface temperature anomalies) can affect SST. As shown by Figure 3.4, the mean upwelling is strong in the eastern and central equatorial Pacific. Thus positive (negative) \( h \) leads to positive (negative) SST anomalies in these regions. This process is increasingly amplified toward the east because of the slope of the mean thermocline (Figure 3.2) which gives larger subsurface temperature anomalies for a given value of \( h \) (according to \([C.5]\)).

As a result of this mean field structure, there is a natural tendency for the ocean to develop SST anomalies which are largest in the east and decreasing toward the west. For such a pattern of SST anomalies, the atmospheric response is a pattern of equatorial wind anomalies which are of a single sign and extend across the entire region of SST anomalies. In the case of warm SST anomalies, the resulting wind anomalies
are westerly, and in the case of cold SST anomalies they are easterly. Since westerly wind anomalies tend to produce warm SST anomalies in the east (by increasing $h$), and easterly anomalies tend to produce cold SST anomalies in the east (by decreasing $h$), the combined pattern of anomalies can be self-sustaining (or growing) in the coupled model (for sufficient coupling strength).

In Runs F and G this interaction is isolated. The results show oscillating, but spatially fixed patterns of wind, temperature, and ULD anomalies. The anomalies are very much localized in the easternmost portion of the ocean, especially in Run G. This somewhat unrealistic feature is not found in the other coupled runs, and occurs in these cases because of the neglect of a surface layer. Consider the surface currents produced by the stationary pattern during a warm event as just described. There are westerly wind anomalies in the eastern Pacific to the west of the temperature anomaly maximum. Such wind anomalies induce anomalous eastward currents and anomalous downwelling locally. Because of the mean temperature gradients, both effects induce local warming. As a result, the temperature anomalies expand westward, and with them, the wind anomalies. Furthermore, because the mean conditions in the central Pacific are more favorable for producing large wind anomalies (the mean wind field becomes convergent, allowing convergence feedback to enhance anomalies, and the mean SST increases, giving a larger atmospheric heating anomaly for a given SST anomaly according to [2.2.5]), the maximum wind anomalies tend to occur there. This is what is found in the simulation run, which includes the surface layer. Thus, in the presence of an oceanic surface layer, the characteristic
warm event pattern is modified in a realistic fashion, although it still derives ultimately from the stationary mode interaction. In the subsequent discussion, stationary mode interaction will refer to modified anomaly pattern in cases which include a surface layer.

If the stationary mode interaction explains the spatial structure of model warm events, it does not explain the temporal characteristics. From the results it appears that in general there are two timescales to account for: the typical duration of warm events, and the preferred time period between warm events. The two do not always vary in the same fashion from one case to another, and so they must be treated separately. There is a clue here that the dynamics of warm events is different from that of 'cold events'.

4.5.1 Warm Events

The different cases suggest that there are several mechanisms that can set the timescales for the coupled oscillations, and that in each case the mechanism with the shortest timescale controls the behavior. For warm events, it is clear that the influence of surface currents is very significant. In Run G, with no surface currents, the events are very long and very slowly evolving. In this case the behavior is determined by the balance (or imbalance) between thermal dissipation and mean upwelling temperature advection. The stationary mode grows due to the positive feedback between the wind field and the h-field as described earlier. As the amplitude of h increases, however, the corresponding increase in subsurface temperature anomaly is proportionately less because of the assumed mean temperature profile.
(see [C.5]). At a certain amplitude (which depends on the coupling strength) the increase in SST anomaly due to the mean upwelling balances the loss due to dissipation, and growth ceases. At this point a small decrease in the amplitude of wind, SST, or h-anomalies leads to a decrease in the others which enhances the original perturbation, and the stationary mode thus relaxes in the same way it grew (nothing in the argument depends importantly on the sign of the perturbation). The development is slow for the assumed upwelling strength and value of $\gamma_1$ (.75). It takes a period of several years to reach the equilibrium point and then relax. Although the effect of the annual cycle is evident during this time, it is clearly of secondary importance compared with the longer timescale development.

The behavior is similar, though more rapid, in Run F which includes the baroclinic currents. The currents act to accelerate the development of the stationary mode. During warm events, the westerly stress anomalies induce eastward currents which advect warmer water from the west, and temperature anomalies are enhanced. The shorter period in this case, then, can be attributed to the more rapid amplification, which brings the system to its extreme in a shorter time.

The addition of the surface layer, however, changes the evolution qualitatively. Anomalous surface currents, and especially anomalous upwelling, become much larger, allowing the effect of the large mean temperature gradients to become more important. Aside from increasing dramatically the rate of development and amplitude of warm events, the surface layer has another very important effect: a heightened sensitivity to the annual cycle. Because anomalous currents are larger,
seasonal changes in the mean temperature gradients have a greater effect on anomaly development. As a result of this increased sensitivity, the warm event evolution appears to be strongly controlled by the annual cycle in those runs where the surface layer is effective. Examples are Runs C2, H, I, and the simulation run.

It is straightforward to see how the annual cycle modulates warm event development. In the simulation run there is a clear tendency for the major events to start in the spring, amplify sharply in the summer, peak in the latter part of the year, and then relax during the following spring (this is also what is observed during real ENSO events). In the spring, upwelling in the east is weakest, SST is warmest, and the easterly winds are weakest. All of these factors make the spring a most unfavorable time for anomalies to amplify (except very near the east coast, where mean upwelling is still strong in general). On the other hand, during summer and fall just the opposite is true. Upwelling is very strong, SST is cold, and the equatorial tradewinds are very strong. This period is ideal for anomaly growth. The coupling between h and SST anomalies is large because of the strength of the mean upwelling. The mean zonal temperature gradients are large, making the effect of zonal current anomalies large. Finally, the strength of the mean winds gives large stress anomalies in comparison with the magnitude of wind anomalies. The winter season is intermediate between the two extremes, making anomaly development less favorable than in summer and fall, and more favorable than in spring.

In runs where the surface layer is effective, this modulation is dominant in warm event evolution, and rather clearly sets the timescale
for warm events. In particular, this is true for the simulation run, which is considered to have the most realistic specification.

Runs D and D2 included a surface layer but not an annual cycle. The temporal evolution of warm events differs noticeably, as expected from the present argument. Obviously, there is no preferred time of initiation, growth or decay. In Run D (which is identical to the simulation run except for the annual cycle) the warm events are large and lengthy, and they terminate in an unrealistic fashion. The events in Run D2 are smaller, but tend otherwise to have the same unrealistic features. This is further evidence that the annual cycle is responsible for the realistic features of the warm event evolution in the model.

It remains to account for the behavior of Runs B and C, which include a surface layer, but exhibit lengthy, large warm events. In Run B, the thermal dissipation is reduced by a factor of two. As a result, the stationary mode growth rate would be expected to increase sharply. The results seem to indicate that the effect is strong enough so that anomaly growth can occur even in the most unfavorable time of year, making the seasonal modulation less effective. The modulation becomes more effective at very large amplitude because (i) the stationary mode growth rate decreases with amplitude, as discussed earlier, and (ii) changes in wind stress increase nonlinearly with amplitude. The case of Run C is different, but leads to a similar result. Here, by decreasing the wave speed, the free dispersion of h is slowed. As a result, the system cannot relax as rapidly as before. For example, if the anomalous wind stress in the central regions relaxes as a result of seasonal changes in the mean wind, the associated changes in h (and
eventually SST) in the east are delayed by the slower wavespeed. Accordingly, further changes in the wind anomalies (which would be induced by the changes in SST) are delayed. Overall the system becomes considerably more sluggish. In this case, the brief period for which growth is inhibited apparently is not sufficient to terminate the event before the next period of favored growth. Again, the seasonal modulation becomes more effective at larger amplitude (for the same reasons as before), so that eventually the warm events terminate. To summarize, in both Runs B and C the results suggest that the annual cycle is still responsible for the preferred time of growth and final termination, but that the modulation is insufficient to control the behavior at modest (realistic) amplitudes.

4.5.2 'Cold Events'

The behavior in periods between warm events is not parallel to that during warm events. First of all, westward propagating disturbances occur during these periods, and not during warm events. It will be shown that this results (in the model) from constraints on the total temperature field. Secondly, it appears that the annual cycle does not control the temporal evolution as it does during warm events. Perhaps the best evidence of this are the results of runs D and D2, which exclude the annual cycle, but have inter-event periods resembling other runs.

The common characteristics of inter-event periods in Runs D, D2, and other runs (with a surface layer) are the westward propagating disturbances and the developments in the equatorial h-field. In all
of these cases, immediately following the termination of warm events the equatorial h-field is negative across the entire basin. After the arrival of each propagating disturbance at the west coast, a large positive ULD anomaly appears and then spreads eastward, tending to increase the h-field along the equator. This process is progressive (although variably rapid in the different cases), and eventually the anomalous h-field is near or above zero across the whole basin. It is only at this point that model warm events occur.

Although the propagating anomalies develop because of surface layer interactions with the atmosphere, as shown earlier, it is clear that they generate baroclinic disturbances of considerable amplitude. As each region of easterly stress propagates westward, positive ULD anomalies develop on either side of the equator. In most cases the anomalies grow and remain localized about the moving wind stress region because the propagation speed is similar to the westward phase speed of the fastest moving Rossby mode. Near the west coast the wind stress anomalies dissipate (because the mean temperature gradients and winds become very small), and h appears to reflect from the western boundary in the form of an equatorial Kelvin wave which propagates rapidly eastward, increasing the equatorial h-field. Exactly how this process leads to a progressive positive trend in the eastern ocean h-field is not obvious. It seems likely, however, that the explanation involves the unequal propagation speeds in the eastward and westward direction. Signals travel eastward at the Kelvin wave speed, and westward at the propagation speed of the coupled anomalies, which happens to be more similar to the (fastest) oceanic Rossby wave speed. Consider an initial state with h = 0 and a region of easterly wind in the eastern ocean
which starts moving westward at the Rossby wave speed. Information travels eastward faster than the anomaly moves westward, so \( h \) becomes negative in the east. On the other hand, positive \( h \) develops to the north and south of the equator without dispersing appreciably, and then reflects at the west coast as a Kelvin wave. At the east coast, the decrease in \( h \) which is forced over a considerable period of time is smaller than the increase in \( h \) associated with the single Kelvin wave pulse, so positive \( h \) results in the east. Conceivably, a sequence of such interactions could lead to a progressive increase in \( h \), as observed in the model results.

The argument is simplified, and somewhat questionable in application to the full model. Still, the available evidence suggests that it is the interaction between the surface layer-induced stress anomalies and the baroclinic motions which they force that gradually 'reconditions' the ocean for warm event development.

In Run C, with a reduced free wave speed, the inter-event periods are longer than in the other runs. This is consistent with the above argument because in this case the fastest Rossby wave speed is considerably slower than the propagation speed of the coupled anomalies. Thus, there is more dispersion, smaller relative amplitudes in the reflected Kelvin waves, and slower reconditioning.

Runs F and G depart significantly from the inter-event scenario described so far. Notice that both runs neglect the surface layer, an essential element of the proposed interaction. In these cases, the inter-event periods resemble the warm event periods, with slowly changing, stationary anomalies. The only differences are that the
cold events are slightly smaller in amplitude and longer in duration. These differences presumably derive from the nonlinearities of the model (wind stress, convergence feedback, and $T_{sub}$) which give quantitatively different results for positive and negative anomalies.

It has yet to be shown why the addition of a surface layer eliminates the symmetry between positive and negative phases. This occurs because of features of the mean fields. The mean thermocline depth is very shallow in the east. Very large positive temperature anomalies can develop by a significant deepening of the thermocline, but only small negative anomalies can develop before the thermocline surfaces. Effectively a bound is imposed on the minimum total temperature. As a result, the stationary mode pattern typical of warm events can exist only at very small amplitudes before the temperature reaches the minimum bound at the east coast, and anomaly growth ceases there. However, the easterly wind anomalies associated with the cold SST anomalies induce anomalous upwelling which allows regions to the west to cool further, so that the temperature anomaly maximum separates off the coast. At this point, the easterlies begin to relax to the east of the maximum, inducing warming, and the cold anomaly region moves westward. This does not happen during the growth of warm anomalies because the subsurface temperature anomalies remain largest at the coast during the entire development. The essential difference between cases with and without a surface layer, then, is that in the latter case anomalous upwelling effects are ignored, making it impossible for the temperature anomaly maximum to separate from the coast. The behavior is thus constrained to follow the pattern of warm events.
4.5.3 Aperiodicity

If the preferred inter-event period is set by the interaction between propagating surface layer disturbances and the baroclinic motions they excite, what accounts for the lack of perfect regularity? The results point to the effects of the annual cycle and especially the nonlinearity of the atmospheric response to SST anomalies. In the simulation run, the development during each 'aborted' warm event is similar, namely, the events start to develop at an unfavorable time of year. The situation usually is that warm temperature anomalies and westerly wind anomalies start to develop in the central ocean in early spring, when there is mean convergence at the equator in the eastern Pacific. At this time there can be significant convergence feedback, and thus strong easterly anomalies develop to the east of the warm SST anomaly. The easterlies cancel the effect of the westerlies, and h does not deepen in the eastern ocean. Thus there is no stationary mode excitation, and no warm event. In situations where the original anomalies develop in the late spring, the large easterlies do not form (there is mean equatorial divergence in the east, and thus no convergence feedback), and warm events develop. This is further evidence that the annual cycle is not determining the behavior during inter-event periods. In these cases warm events were initiated at an unfavorable phase of the annual cycle, and therefore did not develop further. If the annual cycle were controlling the behavior, initiation would not occur at such a time.
Run D2 illustrates in another way the influence of the atmospheric nonlinearity. Here there is no annual cycle, yet there is considerable nonuniformity both in warm events and inter-event periods. An examination of model output shows a variable number of feedback iterations between one event and another: i.e., a variation in the strength of the nonlinear feedback effect. Recall also that in Run J, which did not allow a variable number of iterations, the behavior became very regular and small in amplitude. From these and other cases, it is apparent that the strength of the nonlinearity has much to do with the variability in amplitude and spacing between warm events.

4.6 Comparison with Other Models

Several other investigators have undertaken modeling studies of ENSO. In most cases, the models used are somewhat simpler than the present one, and the goal is to explain some aspect of the observations rather than to produce a realistic simulation. Nonetheless, it is worthwhile to examine some of the ideas offered in these studies in light of the present results.

The highly parameterized model of McWilliams and Gent (1978) produces weak, rapidly damped oscillations following a period of imposed easterly wind forcing. Such features are found in the present model only for very weak coupling (Run A). A likely explanation for the difference is that McWilliams and Gent ignore the large horizontal gradients of mean surface and subsurface temperature in their model. It is the advection of these mean temperature gradients that is responsible for much of the amplitude of the oscillations in the present model.
The study of Lau (1981) considers only the interaction between atmospheric and oceanic Kelvin waves. Atmospheric heating is parameterized as a linear function of $h$, the oceanic upper layer depth anomaly, and no mean field effects are considered (i.e., mean temperature gradients, mean surface wind, etc.). He finds that there is a coupled mode which, for strong coupling, becomes stationary, and suggests that this is a possible explanation for the quasi-stationarity of observed ENSO anomalies. The present results point rather to the configuration of the mean wind and temperature fields, which strongly modulate the coupling between $h$ and atmospheric heating, as the source of the stationarity (section 4.5).

McCready (1983) adopts a dynamical ocean model similar to this one, except without a surface layer. The atmosphere, however, is highly parameterized, and consists of a prescribed anomaly pattern which is turned on or off according to the magnitude of $h$-anomalies in the eastern ocean. It should be noted that the pattern contains a subtropical component for which there is little or no observational support. It was found that the model could produce oscillations at arbitrary intervals, depending on the scale and position of the prescribed wind patterns. The intervals were determined by the westward propagation of a subtropical Rossby wave across the basin after the ENSO-state winds switched off. In the present model, no such behavior was found. The same is true of all other models which allow a continuously variable atmosphere that depends on the entire tropical SST anomaly field. In all of these cases, subtropical ULD anomalies do not develop and survive long enough
to propagate (slowly) across the basin, reflect as a Kelvin wave, and initiate another warm event, as occurs in McCreary's model.

McCreary and Anderson (1984) relax some of the questionable aspects of the McCreary (1983) model by eliminating the unrealistic dependence of subtropical wind anomalies on the eastern ocean h-field. They include a tropical component designed to mimic equatorial tradewind fluctuations during ENSO, and an annual cycle in the tropical tradewinds (which is highly idealized). Again, though, the anomaly pattern is fixed, and has only two states: on and off. The transition from one state to the other occurs when the eastern ocean thermocline moves above or below a critical depth. The model gives interannual oscillations which depend on the amplitude of the annual cycle and the critical thermocline depth. For some parameter choices, warm events have a duration and inter-spacing which is similar to observations. However, the mechanism controlling the behavior depends sensitively on the unrealistic atmospheric specification. The anomalous wind field has only two configurations. During the period it is in either state, the ocean approaches its equilibrium state for that wind stress field. When the eastern ocean thermocline becomes sufficiently close to the equilibrium state, the superimposed annual oscillation causes it to pass through the critical depth, and the wind field switches to the alternate state. The timescale of the oscillation thus depends fundamentally on the action of free waves which bring the ocean toward equilibrium. This process can occur only because the wind field is not allowed to change continuously with the SST anomaly field. In the present model, the equatorial h-field during the lifetime of a warm event exhibits the
character of a forced oscillation, with no evidence of free wave propagation. Here it is the modulation of the strength of the stationary mode interaction by the annual cycle which determines the lifetime of a warm event. Thus, the present results suggest that, with a less idealized atmosphere model, the role of the annual cycle is more important, and the role of free wave modes is less important than predicted by the McCreary and Anderson (1984) model.

Philander et al. (1984) develop a model more analogous to the one used here. The atmosphere dynamics are the same. The thermodynamics are, however, much simpler, with no convergence feedback, and a heating which depends directly on the oceanic h-field. The ocean model is the same, except for the neglect of a surface layer. Finally, the effect of the mean fields is not included. The model gives unstable growth of coupled anomalies for sufficient coupling strength. This behavior can be compared with the growth of stationary mode anomalies in the model used here, except that in this case the anomalies are spatially fixed due to the influence of the mean fields which are ignored in Philander et al. (1984). The stationary mode grows as a result of unstable interactions between the large scale oceanic h-field and the atmosphere, similar to the development in the Philander et al. model. However, other aspects of the ENSO cycle cannot be explained by their model (e.g., termination and regeneration of warm events). In the present model, these features depend on the additional effects of the annual cycle, the prescribed mean field, and the oceanic surface layer.

Very recently Anderson and McCreary (1984) have investigated a coupled model in which the oceanic component includes explicit thermo-
dynamics. The ocean model consists of a single layer with variable temperature. In some respects the model is more elaborate than that used in this study (e.g., nonlinear terms are retained, and the thermo-dynamics influence the dynamics). However, the model does not consider a separate mixed layer near the surface which is distinct from the oceanic upper layer. That is, the effect of the surface wind stress is distributed uniformly over the entire upper ocean, rather than preferentially near the surface (as in the present model). Furthermore, the effects of entrainment and upwelling are parameterized as independent of the surface stress anomalies. These assumptions make SST much less sensitive to surface wind stress than in the present model. With respect to the atmosphere component, the model is equivalent to the one used here. An important difference, though, is that the model attempts to simulate the total SST and wind fields, rather than just the perturbation fields, and no account is taken of the annual cycle. The model produces large amplitude, very slowly eastward propagating anomalies for a wide range of parameter choices. In the closed basin calculations, the results become periodic with a period of several years, which is also the period required for a single perturbation to travel across the basin. For the most part, the results do not resemble observations. The time scales are much too long, and the behavior much too regular. There is no evidence of stationarity of anomalies or westward expansion of anomalies, both of which are characteristics of typical ENSO events during at least part of their life cycle. The slow timescales and regularity of anomalies were also found with the present model when the surface layer was ignored (Run F). In that case, as with the Anderson
and McCreary model, current anomalies are weak, and the mechanism for rapid SST changes is absent. Also, it is apparent that the model does not well simulate the mean fields or the annual cycle, which were found to determine the spatial structure of anomalies in the present model. This is not particularly surprising. The configuration of the mean fields almost certainly involves many more effects than are included in the model (e.g., orography, radiation, neglected nonlinear terms). It was for this reason that the mean fields were specified in this study. The eastward propagation found by the authors does resemble the development during part of the 1982-3 ENSO event. However, the propagation is in some doubt in the model because it uses a linear stress formulation which overemphasizes westerly stress anomalies relative to easterly stress anomalies.

Another, simpler model has been examined by Gill (1984). This model assumes a prescribed meridional structure for all anomalies, and assumes that SST anomalies are produced only by the zonal advection of mean temperatures by anomalous zonal currents. Additionally, the annual cycle is ignored. The model produces weak, westward propagating coupled anomalies with a period slightly greater than one year. Thus, the behavior does not resemble observed interannual fluctuations. It can, however, be understood from the present results. Since there is no coupling between h and the atmosphere (i.e., the mean upwelling has no effect), there is no stationary mode excitation in the model, and therefore no characteristic ENSO events. The behavior is similar to Run E, where h was ignored. In that run there were only westward propagating disturbances, and no ENSO-like events. In Gill's model
the anomalies are much weaker because there is no surface layer. Therefore current anomalies are smaller, implying smaller temperature anomalies and smaller wind anomalies.

The above references cover the spectrum of simple (dynamical) coupled models that have been developed and discussed in the literature to date. It is felt that the present study offers a considerable advancement in the simulation of observed interannual variability, and in particular the life cycle of warm events, although there are still unrealistic features of the model deserving further attention. In the following chapter, the strengths and weaknesses of the model will be reviewed, and discussed in terms of their physical implications.
5.1 Summary

The goal of this work has been to develop a coupled model which is as simple as possible, and yet can capture the salient features of observed ENSO variability in the tropical Pacific. The two component models were developed and tested separately, in each case using observations to specify the forcing and to evaluate the results. Two fundamental hypotheses underlie the work, and are substantiated by the results: (i) that the surface wind anomalies can be explained largely in terms of (tropical Pacific) SST anomalies, and (ii) that the SST anomalies can be explained largely in terms of (tropical Pacific) surface wind stress anomalies.

In Chapter II the atmosphere component was developed. By assuming a specific vertical structure for the flow, the governing equations were reduced to the steady-state, linear shallow water equations on an equatorial beta-plane. Dissipation assumes the form of Rayleigh friction and Newtonian cooling. Two versions of the model were examined, one in which heating was parameterized solely in terms of local SST, and another which additionally included a low-level convergence feedback mechanism. The parameterization is intended to approximate the effect of convective heating on the large scale flow. That is, in regions where the total wind field is convergent, the low-level convergent flow associated with a heating anomaly is assumed to lead to an anomalous influx of moisture which is released as latent heat, enhancing the initial heating anomaly. In regions where the total wind field is
divergent, it is assumed that there is no organized convection, and thus no convective enhancement of heating anomalies. The version of the model which includes the feedback gives results much more comparable to observations, and yields equatorial wind anomalies which, for the most part, agree well with observations. In particular, the model well simulates the large westerly anomalies in the central Pacific which characterize the major ENSO signature. The major disagreement was found during a brief period encompassing the termination phase, when the strength of equatorial easterly anomalies was underestimated in the western Pacific and overestimated in the eastern Pacific.

In Chapter III the ocean component was developed. The dynamics of the model consist of an oceanic upper layer governed by the time-dependent, linear, shallow-water equations, plus a near-surface frictional layer designed to mimic the turbulent mixed layer. Surface temperature anomalies are predicted using a complete three-dimensional advective equation. Additionally there is an assumed surface heat flux which is a linear function of the local SST anomaly, acting to damp anomalies. Mean temperature fields are specified from observations, and the mean currents are determined by running the model to equilibrium with observed monthly mean winds. The model well reproduces the major features of observed SST anomaly evolution during ENSO events. There are however, differences in certain details. The model underestimates the amplitude of the initial coastal peak and overestimates the amplitude of the second peak. Also, the spatial pattern of cold anomalies during the termination and post-event phases is slightly different. It was suggested that the lack of a seasonal cycle in subsurface temperature in the
model, and possibly inadequate spatial resolution near the east coast, may account for these differences. In any case, the differences are quantitative rather than qualitative, and the general success of the model is good.

By examining the mechanisms producing SST anomalies in the model, it was determined that: (i) very near the east coast the dominant mechanism producing SST anomalies during ENSO is the mean upwelling of anomalously warm subsurface temperatures, (ii) in the eastern Pacific the dominant mechanisms are zonal advection and vertical advection due to anomalous upwelling, and (iii) in the central and western Pacific the dominant mechanism is zonal advection. Temperature anomalies in regions away from the equator are produced by poleward advection associated with the mean Ekman flow.

In Chapter IV, the component models were combined in a series of coupled calculations. A modification in the atmospheric feedback procedure was introduced in order to reflect a finite atmospheric response time for continuously changing SST anomalies. Using parameter values taken from the forced calculations, and starting off from an imposed initial state with westerly wind anomalies in the central ocean, the model was run for 90 years of simulated time. The results display many realistic features. Among them are (i) the spatial configuration of anomaly fields during warm events, (ii) the recurrence of warm events of variable amplitude at variable intervals, though with a strong preference for three years, (iii) the duration of warm events (slightly more than one year), and (iv) a strong tendency for warm event development to be phase-locked to the annual cycle, with peak amplitudes typically
around the end of the calendar year. Also, the long-term mean of all predicted anomalies is nearly zero. The most significant difference between the model and observations relates to warm event termination, which is delayed by a few months in the model. This difference is traceable to the failure of the model to predict the large easterly anomalies in the western Pacific, a problem also evident in the atmospheric forced calculations. In addition, the model tends to overestimate wind anomalies in the eastern Pacific somewhat, and underestimate the initial coast warming. These also were characteristic of the individual component models. During inter-event periods the model displays surges in the equatorial tradewinds which propagate rather rapidly from east to west with associated regions of SST anomaly.

To elucidate the mechanisms underlying the model behavior, the results from a number of additional coupled calculations were examined. These calculations were designed to isolate particular processes and to test model sensitivities. The combined results pointed strongly to the importance of the mean fields, the annual cycle, and the surface layer in determining the characteristics of the model variability.

Two distinct types of interaction were identified. The first involves coupling between the atmosphere and the upper ocean thermal structure through the action of the mean upwelling. This mode of interaction is stationary with a fixed spatial pattern resembles that of ENSO events (large SST anomalies in the eastern ocean, and large westerly wind anomalies in the central ocean). Warm events in the model are associated with the excitation of this mode. The second mode of interaction involves only the oceanic surface layer and the atmosphere. This is
a local mode interaction which, due to the mean oceanic temperature gradients, produces only westward propagating disturbances. In the model, this mode of interaction is responsible for the surges of the tradewinds which characterize inter-event periods. The combination of these tradewind surges and the baroclinic motions that they excite (Kelvin wave pulses) leads to an oscillatory, but generally deepening thermocline in the eastern ocean during the inter-event period. If, through this process, the eastern ocean thermocline deepens sufficiently at a favorable time of year, the stationary mode interaction is excited and a warm event follows. If the excitation occurs at an unfavorable time of year, the potential warm event aborts quickly. The influence of the annual cycle is thus seen to be an important element of the aperiodicity of interannual oscillations in the model. It was also shown to be dominant in the evolution of warm events once they are initiated. Finally, the nonlinearity of the atmospheric feedback was shown to contribute to the irregularities in both amplitude and spacing of warm events.

5.2 Conclusions

The model results suggest answers to a number of the fundamental questions surrounding ENSO which were posed earlier. If ENSO behaves as in the model, then the following statements can be made.

- Extratropical forcing — ENSO is fundamentally a tropical event. This is not to say that extratropical effects can exert no influence, but only that the influence may not be critical or essential for interannual ENSO-like variability.
o random forcing — this also is not essential for interannual ENSO-like variability. Well defined tropical Pacific atmosphere-ocean interactions can account for the observed behavior.

o the characteristic 'phase-locked' development of warm events — this is clearly identifiable with the role of the annual cycle, which strongly modulates the positive feedback between atmospheric and oceanic anomalies. The typical observed pattern of anomaly development during (N.Hem.) spring, growth during summer and fall, and termination in the following spring/summer can all be explained in terms of seasonally dependent feedback which varies with the normal seasonal changes in the mean fields (see section 4.5.1).

o the characteristic spatial pattern of ENSO anomalies — The configuration of the mean fields determines that the sensitivity of SST to thermocline displacements and local wind effects is highest in the eastern Pacific, and the sensitivity of the winds to local (and remote) SST anomalies is largest in the central and western Pacific. Models which ignore the effect of the mean fields do not reproduce the characteristic structure.

o coupled instability — the amplification of the stationary mode in the model represents an unstable interaction between atmosphere and ocean. However, this describes only one phase of warm event evolution, and cannot, for example, account for the decay of anomalies during the same period of the following year. Warm events are better described as a preferred mode of oscillation of the coupled system.

o initiation of warm events — in the model, initiation was traced to the action of the tradewind surges and the equatorial mass adjustments
that they force. The tradewind surges arise from local interactions between the atmosphere and the oceanic surface layer, and propagate from east to west. Although exactly comparable features are not conspicuous in the data, less well-defined fluctuations in the tradewinds (and SST) are clearly evident. In the model, the local coupling between atmosphere and ocean is exaggerated because of the parameterization of atmospheric heating, which includes a component proportional to the local SST anomaly. As discussed in Chapter II, the local coupling is expected to be less direct in reality because of the possibility of moisture advection. Through this process, the effect of a SST anomaly in a region of mean divergence is felt downstream rather than locally. The results, then, lead to the following speculation: that the initiation of warm events may occur in reality as a result of either directly or indirectly coupled fluctuations of the equatorial tradewinds and surface layer SST anomalies. The most likely region to find such interaction is the equatorial western Pacific, a region of intense moisture convergence and very warm SST which additionally experiences large annual variability associated with the Southeast Asian Monsoon. In this region small fluctuations in SST at certain times (e.g., the monsoon transition periods when the center of convergence crosses the equator) could conceivably induce sizeable anomalies in the equatorial winds, which, in a favorably conditioned ocean, would trigger major warm events. A similar idea has been proposed by Nicholls (1984).

- preferred inter-event duration — in the model this was related to the progressive influence of the tradewind surges on the large-scale equatorial h-field. There is a very strong preference for a regular
period (three years) presumably because of the regularity and uniformity of these propagating features in the model. In the actual system, where the coupling depends on moisture transport, natural fluctuations between equatorial winds and surface layer temperatures may also tend to influence the upper ocean in a similar fashion, but if so, almost certainly the influence would be less regular and uniform. This might account for the greater variability of inter-event duration in the observations (two years and four years are as common as three years). It must be emphasized, however, that there are other possible mechanisms for giving a preferred inter-event period in a more sophisticated model that considers moisture transport. It will be necessary to consider such a model in order to answer this question more definitively.

- aperiodicity — aperiodicity can result from a positive feedback which depends strongly (i.e., nonlinearly) on the annual cycle, together with a natural variability which is largely independent of it. Potential ENSO precursors can be produced at random times with respect to the annual cycle, but are able to develop only when that time corresponds with the favorable growth period. A major source of the nonlinearity involves the atmospheric response to SST anomalies. This leads to an effective coupling strength which varies with the magnitude of both the anomalies and the mean fields.

- predictability — there is no obvious criterion which allows model warm events to be predicted before they are already well underway. The implication is that the difference between states which later develop into warm events and others that do not may be very subtle, and not observable. Much more work is required to substantiate this result.
5.3 Outlook

The results of this study suggest that it will be possible to capture the essentials of ENSO variability with a coupled model that is much simpler than a GCM. On the other hand, certain elements have been shown to be essential in approximating the observed anomalies. These include

- Mean wind and temperature fields
- Annual cycle in wind and temperature fields
- Oceanic surface layer
- Three-dimensional oceanic temperature advection
- Atmospheric (horizontal) moisture convergence effects.

On the basis of the present results, it is suggested that no model which ignores any of the above will produce a realistic simulation of the observations.

The present model can be interpreted in terms of the mutual interaction of three independent components:

1. Stationary (ENSO) mode
2. Disturbance generator
3. Annual cycle in coupled feedback

Warm events are characterized by the excitation of the stationary mode. This mode has a well-defined spatial structure which is determined by the mean fields. The disturbance generator of the model is the atmosphere-oceanic surface layer interaction, which produces anomalies that can excite that stationary mode, but only during that phase of the annual cycle in which positive feedback is strong. Once excited, the stationary mode development is controlled in a well-defined manner.
by the annual cycle modulation. It is suggested that these are essential components of a model which can produce aperiodic oscillations with realistic characteristics.

Many features of the model behavior have been explained only tentatively. In these cases, considerable further work is required to substantiate the conclusions reached. This is particularly true with regard to inter-event periods and their duration. There is likely much more that can be learned from this model, even in its present state, through additional, carefully designed experiments.

Although the model is successful in reproducing many features of the observations, systematic discrepancies are also evident. In particular, warm events do not terminate as observed because of the failure to reproduce easterly wind anomalies in the western Pacific, and the serious overestimation of easterly anomalies in the eastern Pacific. Also, model warm events do not show the typical pattern of large coastal SST anomalies prior to the large central Pacific warming.

There are several ways in which the model could be improved. Among the most important is the inclusion of an explicit boundary layer moisture budget which allows for moisture transport between regions of divergence and convergence. This will reduce the correlation between local SST anomalies and atmospheric heating in regions of divergence, such as the equatorial eastern Pacific, where the present model clearly overestimates heating anomalies. Associated with the decrease in heating anomalies in regions of divergence will be an increase in heating anomalies in the regions of convergence. Most importantly, this includes the western and central Pacific, where the present model underestimates
the anomalies. The failure to reproduce the large easterly anomalies in the western Pacific almost certainly derive from the failure to account for the non-local contribution to the local heating, i.e., the reduced inflow of moisture into the western Pacific during the mature phase of ENSO when there is large anomalous convergence in the central Pacific. In order to successfully simulate boundary layer moisture transport it may be necessary to consider a separate atmospheric boundary layer. This is especially true in regions near the ITCZ, where observations show considerable vertical structure at low levels.

Another model improvement involves the atmospheric dissipation parameterization. A very large value of dissipation is used in the model, a value that can be attributed only to the effect of cumulus-scale motions on the large-scale flow. Given this interpretation, a more consistent formulation would assign the large dissipation only to the regions of intense convection (i.e., the regions of large net heating). Other regions would remain relatively inviscid, as is generally assumed for the atmosphere. In Chapter II it was argued that, with such a parameterization, the disagreement regarding easterly anomalies in the eastern Pacific may be eliminated.

The failure to produce a large initial coastal warming in the model was attributed in part (Chapter III) to the lack of a seasonal cycle in the oceanic subsurface temperature field. This was ignored simply because of the lack of data. It will probably improve the model to parameterize the subsurface temperature field in terms of the thermocline depth, and then to specify an annual cycle in thermocline depth from an equilibrium model run which is forced by observed monthly mean winds.
Since the parameterization for subsurface temperature anomalies already is done in an analogous way, this will additionally make all calculations of subsurface temperature consistent. Also, an increase in spatial resolution in the narrow upwelling zone near the east coast will probably improve the model.

One other simple change would make the model energetically more consistent: to equate the equivalent heat gain in the atmosphere (due to increased local evaporation) to the heat loss in the oceanic surface layer. At present, these parameterizations give this result only approximately.

While complicating the model to some extent, these modifications will likely allow an even better simulation of ENSO. If so, yet further progress can be made in answering the many questions surrounding this intriguing phenomenon.
References


APPENDIX A

Dimensional Estimation of Atmosphere Model Parameters

The governing equations for the flow are the following (on an equatorial beta-plane):

\[(u_a)_t - \beta \omega y v_a = -\frac{p_x}{\rho_o} + F \quad [A.1]\]

\[(v_a)_t + \beta \omega x u_a = -\frac{p_y}{\rho_o} + G \quad [A.2]\]

\[p_z = -\rho_o g \quad [A.3]\]

\[T \frac{ds}{dt} = \dot{Q} \quad [A.4]\]

\[\rho_t + \nabla \cdot (\rho u_a) = 0 \quad [A.5]\]

\[p = \rho RT \quad [A.6]\]

Equations [A.1] and [A.2] are the horizontal momentum equations (neglecting the nonlinear momentum terms), where \(F\) and \(G\) are external forcing functions. Equation [A.3] is the hydrostatic relation. Equation [A.4] is the thermodynamic energy equation, where \(s\) is the specific entropy, and \(\dot{Q}\) is the heating rate per unit mass. Equation [A.5] is the equation for conservation of mass, and equation [A.6] is the equation of state.

Additionally, the flow is assumed to be Boussinesq. That is, variations in density are neglected except in producing buoyancy. With this approximation, [A.5] becomes a relation for incompressibility:

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(u_a)_x + (v_a)_y + (w_a)_z = 0 \hspace{1cm} [A.7]

Also,

\[ T \, ds = c_p \, dT|_p - v' \, dp|_T \approx c_p \, dT|_p. \hspace{1cm} [A.8] \]

where \( c_p \) is the specific heat at constant pressure, and \( v' \) is the specific volume. Thus [A.4] becomes

\[ c_p(dT/dt) = \dot{Q} \hspace{1cm} [A.9] \]

From the state equation, one can write

\[ dp = 1/RT \, \left( dp|_T - \rho/T \, dT|_p \right) \approx -\rho_0/T_0 \, dT \hspace{1cm} [A.10] \]

where the approximate relation again derives from the Boussinesq assumption, with \( \rho_0 \) the basic state density and \( T_0 \) the basic state reference temperature. Using [A.10], [A.9] becomes

\[ dp/dt = -\rho_0 \alpha_o \frac{\dot{Q}}{c_p} \hspace{1cm} [A.11] \]

where \( \alpha_o = 1/T_0 \).

Retaining only the buoyancy term in [A.11] gives

\[ \rho_t + w_a(dp_0/dz) = -\rho_0 \alpha_o \frac{\dot{Q}}{c_p} \hspace{1cm} [A.12] \]
Defining \( N^2 = -(g/\rho_o) \, d\rho_o/dz \) and combining with [A.3] gives

\[
\left( \frac{p}{\rho_o} \right)_{zt} + N^2 w_a = a_o g Q/c_p \tag{A.13}
\]

The assumed vertical structure for the flow is:

\[
(u_a', v_a', p', F', G') \cos mz = (u_a, v_a, p, F, G) \tag{A.14}
\]

\[
w_a' = (w', Q') \sin mz \tag{A.15}
\]

Applying [A.14] and [A.15] to [A.1], [A.2], [A.13], and [A.7] gives

(dropping primes)

\[
(u_a)_{t} - \beta_o y v_a = -(p/\rho_o)_x \tag{A.16}
\]

\[
(v_a)_{t} + \beta_o y u_a = -(p/\rho_o)_y \tag{A.17}
\]

\[-m(p/\rho_o)_{t} + N^2 w_a = a_o g Q/c_p \tag{A.18}
\]

\[(u_a)_x + (v_a)_y = -mw_a \tag{A.19}
\]

[A.18] and [A.19] can be combined to form one equation:

\[
\left( \frac{p}{\rho_o} \right)_{t} + N^2/m^2 \left( (u_a)_x + (v_a)_y \right) = -a_o g Q/m c_p \tag{A.20}
\]

The nondimensional equations are formed by making the following assignments

(dimensional variables denoted with asterisks):

\[
u_a^*, v_a^* = U_a(u_a, v_a)
\]

\[p^* = \rho_o U_a c_a p\]
\[(x^*, y^*) = (c_a/2\beta_o)^{1/3} (x, y) \]
\[t^* = (2\beta_o c_a)^{-1/3} t \]
\[\varepsilon^* = (2\beta_o c_a)^{1/3} \varepsilon \]
\[(c_a = N/m) \]

The result is

\[\varepsilon u_a - yv_a/2 = -p_x \quad [A.24] \]
\[\varepsilon v + yu_a/2 = -p_y \quad [A.25] \]
\[\varepsilon p : (u_a)_x + (v_a)_y = -Q_1, \quad [A.26] \]

where

\[Q_1 = \frac{a_o g}{(2\beta_o c_a)^{1/3} U_a m c_p} \frac{\dot{Q}}{Q} \quad [A.27] \]

Taking \(U_a = 1m/sec, \ g = 9.8 \ m/sec^2, \ a_o = 1/300^0K, \ m = (5 \times 10^3 \ m)^{-1}, \ c_a = 60 \ m/sec, \ c_p = 10^3 \ m^2/sec^2/^0K, \) and \(\beta_o = 2.26 \times 10^{-11} \ sec^{-1}m^{-1} \) in

\[Q_1 = 52.3 \dot{Q}, \quad [A.28] \]

which relates the dimensional heating rate/unit mass to the nondimensional heating.

The model heating parameterization, defined by equation [2.2.5],

is

\[Q_1 = a (SSTA) \frac{T_{ref}^2}{T_M} \exp \left( b \left( \frac{1}{T_{ref}} - \frac{1}{T_M} \right) \right). \quad [A.29] \]

To get an estimate for \(a, \) consider the case where \(T_M = T_{ref}, \) for which
\[ Q_1 = \alpha \text{ (SSTA)}. \]  

For perturbation in temperature about \( T_{\text{ref}} = 303^0\text{K} \), the Clausius-Clapeyron equation gives

\[ \Delta e = \left( \frac{de_s}{dT} \right)_{303^0\text{K}} \frac{\Delta T e_s (303^0\text{K})L}{R(303^0\text{K})^2}, \]

where \( e_s \) is the vapor pressure of water vapor, \( e_s \) is the saturation vapor pressure, \( L \) is the latent heat of condensation \( \approx (2.45 \times 10^6 \text{m}^2/\text{sec}^2) \), \( R \) is the gas constant \( \approx (461.5 \text{m}^2/\text{sec}^2/\text{K}) \), and \( \Delta T \) is the change in temperature, here identified with the sea surface temperature anomaly. From a standard table one can find that \( e_s (303^0\text{K}) \approx 40\text{mb} \). Using the relation \( \Delta q = \Delta e/p_o \), where \( q \) is the specific humidity, and \( p_o \) is the atmospheric pressure \( \approx 1000\text{mb} \), [A.31] can be written

\[ \Delta q = \frac{\Delta T e_s (303^0\text{K})L^2}{p_o R(303^0\text{K})^2} \approx \text{ (SSTA) (5.7 \times 10^3)}. \]

To relate this to the model heating one must account for the vertical distribution of heating. If the homogeneous boundary layer is taken to be 1 km. deep, and it is assumed that the vertical mode extends to a height of 10 km, then the amplitude of the model heating is approximately \( \pi/20 \) times that of the equivalent heating due to humidity changes in the boundary layer. Finally, the assumption is made that this anomalous accumulation or deficit of equivalent heat is 'felt' over a period
of several hours (specifically, seven hours used here, based on observational results during GATE).

Therefore the dimensional model heating is, (using [A.32])

\[
\dot{Q} = \frac{L \Delta q}{7 \text{ hrs}} \frac{\pi}{20} = (\text{SSTA})(.031) \, \text{m}^2/\text{sec}^3
\]

[A.33]

The nondimensional model heating is obtained by using [A.33]

\[
Q_1 = (52.3)(.031)(\text{SSTA}) = 1.6 \, (\text{SSTA})
\]

[A.34]

Thus the estimate for \( \alpha \) in [A.30] is 1.6; this is the value used in the simulations of Chapter II and Chapter IV.

The estimate for the parameter \( \beta \) in the convergence feedback model (see [2.4.1]) is now easily made. In the boundary layer, the rate of heat input per unit mass due to horizontal mass convergence is

\[
Lq_t = -Lq_s (\nabla \cdot \mathbf{u})
\]

[A.35]

The value of \( q_s \) is taken to be .04, which is appropriate for \( T = 30^\circ \text{C} \) (the reference temperature). The nondimensional divergence is scaled by \( U_a/L = 1 \, \text{m/sec}/10^6 \text{m} = 10^{-6} \text{sec}^{-1} \). Accounting again for the vertical distribution factor, and using [A.28], the model heating is

\[
Q_1 = (52.3) \frac{\pi}{20} (L)(.04)(10^{-6})(\delta_{\text{non}}) = 0.7 \, (\delta_{\text{non}})
\]

[A.36]
where $\delta_{\text{non}}$ is the nondimensional convergence. Comparing [A.36] with equation [2.4.1], the estimate for $\beta$ is seen to be 0.7. In the simulations of Chapter II and Chapter IV, $\beta$ was taken to be 0.75.

Many questionable assumptions have been made in deriving these estimates. The results should be considered at best order-of-magnitude estimates for the parameter values. In practice, however, it was found that the dimensional model results compared most favorably with the observations for values of $\alpha$ and $\beta$ close to the estimated values.
APPENDIX B

Selected Monthly Mean Maps of SST, Wind and Convergence

Monthly mean maps of SST, surface wind, and surface wind convergence for March, May, August, and November are presented. The data represent the 30-year climatology from which the Rasmussen and Carpenter (1982) composite ENSO anomaly fields were derived (spanning 1946-1976). A more complete description of the annual cycle in the Tropical Pacific may be found in Horel (1982).

a. March Conditions (Figs. B1(a) - B1(c))

The SST is at its maximum in the eastern equatorial Pacific at this time. In the west the maximum SST is near 10°S. The wind field shows weak Southeast Trades (compared to the annual mean), and strong Northeast Trades. Near the maximum SST the wind field is weak, cyclonic and convergent. The convergence in the ITCZ is at its weakest, and the zone spans the equator at this time. Poleward of 10°N and 10°S there is strong divergence in the eastern Pacific. The SPCZ is poorly defined, with weak convergence from the equator to 25°S.

b. May Conditions (Figs B2(a) - B2(c))

The SST is now cooler in the eastern equatorial Pacific; in the western Pacific, the broad zone of maximum SST spans the equator. The equatorial tradewinds in the eastern Pacific are stronger than in March. The region of small winds now sits on the equator at the same longitude (150°E) where the belt of maximum SST crosses the equator; the region around 10°S and 160°E now is characterized by sizeable south-easterly winds. The ITCZ is well north of the equator in the east,
Fig. B.1 Long term means of (a) SST, (b) surface wind, and (c) divergence, for March.
Fig. B.2 Long term means of (a) SST, (b) surface wind, and (c) divergence, for May.
and stronger than in March. On the equator there is weak divergence. The SPCZ is weak as in March, but narrower, with maximum convergence near $10^\circ$S.

c. **August Conditions** (Figs. B3(a) – B3(c))

The cold tongue in the eastern equatorial Pacific is most pronounced at this time. In the west the warmest water reaches its northernmost extent, at around $15^\circ$N. The Northeast Trades are correspondingly weak, and the Southeast Trades are very strong. The far western Pacific has southerly flow across the equator, and weak, convergent flow in the vicinity of the warmest SST. The ITCZ is strong and considerably north of the equator, and sizeable divergence exists on the equator from the dateline eastward. The SPCZ is strong and well defined.

d. **November Conditions** (Figs. B4(a) – B4(c))

Now the region of warmest water in the western Pacific is again spanning the equator, as in May, but in this case moving southward rather than northward. The cold tongue is still pronounced in the east, although slightly less so than in August. The wind field is characterized by intermediate strength tradewinds in both hemispheres, and again a region of weak, cyclonic, convergent flow near the maximum SST (at the equator in November). To the north and south of the equator, the winds become strong. The same regions had weak winds when the maximum SST was nearby. At this time the ITCZ is centered at around $9^\circ$N in the eastern Pacific, and the equatorial divergence east of $180^\circ$ is at its maximum strength. The SPCZ is quite strong, and at its southernmost location, centered around $15^\circ$S.
Fig. B.3 Long term means of (a) SST, (b) surface wind, and (c) divergence, for August.
Fig. B.4: Long term means of (a) SST, (b) surface wind, and (c) divergence, for November.
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APPENDIX C

Surface Temperature Anomaly Model—Governing Equations

Dimensional forms of the governing equations are developed

III. They are reproduced below:

\[ \begin{align*}
\frac{\partial t}{\partial x} - \beta_0 y V &= -gH_0 h_x + \frac{\tau(x)}{\rho} - rU \quad \text{[C.1a]} \\
\beta_0 y U &= -gH_0 h_y + \frac{\tau(y)}{\rho} - rV \quad \text{[C.1b]} \\
\frac{\partial}{\partial t} (U_x + V_y) &= -\rho h \quad \text{[C.1c]} \\
\end{align*} \]

\[ u, v = \frac{(U, V)}{H_1} \quad \text{[C.2]} \]

\[ \begin{align*}
J_s &= (r_s \tau(x) + \beta_0 y \tau(y)) / \rho (\beta^2 y^2 + r_s^2) \\
\frac{\partial T_s}{\partial x} &= \frac{r_s \tau(x) - \beta_0 y \tau(y)}{\rho (\beta^2 y^2 + r_s^2)} \quad \text{[C.3a]} \\
\frac{\partial T_s}{\partial y} &= \frac{r_s \tau(y) + \beta_0 y \tau(x)}{\rho (\beta^2 y^2 + r_s^2)} \quad \text{[C.3b]} \\
\end{align*} \]

\[ \begin{align*}
\nu_{sfc} &= u - U_s/H_1 + U_s/H_s \quad \text{[C.4a]} \\
\nu_{sfc} &= v - V_s/H_1 + V_s/H_s \quad \text{[C.4b]} \\
w &= H_s ((u_{sfc})_x + (v_{sfc})_y) \quad \text{[C.4c]} \\
\end{align*} \]

\[ w = A(\tanh[B(\bar{h} + h)] - \tanh[B\bar{h}]) \quad \text{[C.5]} \]

\[ T_z = \frac{T_s - T_{sub}}{H_s} \quad \text{[C.6]} \]

\[ \begin{align*}
t_{sfc}^{-1} \cdot \nabla (\bar{T}_s + T_s) - \nabla T_s &= -\frac{\nu_{sfc} \cdot \nabla T_s}{\nu_{sfc}} - \gamma_1 F_1(\bar{w}) T_z - \gamma_2 F_2(\bar{w}, \bar{w}) (\bar{T}_z + T_z) - \alpha_s T_s \quad \text{[C.7]} \\
\end{align*} \]
This is the most complete text of the thesis available. The following page(s) were not included in the copy of the thesis deposited in the Institute Archives by the author:

pp 249 - 251
Algorithm for Convergence Feedback in the Coupled Model

The total heating anomaly field is calculated once per month, using the same scheme as for the forced calculations:

\[
Q_{n+1} = \begin{cases} 
Q_0, & \text{if } (\delta_M + \delta_n') > 0 \text{ and } \delta_M > 0 \\
Q_0 - \beta(\delta_M + \delta_n'), & \text{if } (\delta_M + \delta_n') \leq 0 \text{ and } \delta_M > 0 \\
Q_0 + \beta(\delta_M), & \text{if } (\delta_M + \delta_n') > 0 \text{ and } \delta_M \leq 0 \\
Q_0 - \beta\delta_n', & \text{if } (\delta_M + \delta_n') \leq 0 \text{ and } \delta_n' \leq 0,
\end{cases} \tag{D.1}
\]

where \( Q_0 \) is defined by [2.2.5], \( \delta_M \) is the prescribed mean divergence, and \( n \) is the iteration index.

At intermediate timesteps, the change in heating anomaly is calculated in a similar fashion:

\[
Q_{n+1} = \begin{cases} 
Q_0', & \text{if } (\delta_T + \delta_n') > 0 \text{ and } \delta_T > 0 \\
Q_0' - \beta(\delta_T + \delta_n'), & \text{if } (\delta_T + \delta_n') \leq 0 \text{ and } \delta_T > 0 \\
Q_0' + \beta(\delta_T), & \text{if } (\delta_T + \delta_n') > 0 \text{ and } \delta_T \leq 0 \\
Q_0' - \beta\delta_n', & \text{if } (\delta_T + \delta_n') \leq 0 \text{ and } \delta_n' \leq 0,
\end{cases} \tag{D.1}
\]

where

\[
Q_0'(t) = Q_0(t) - Q_0(t - \Delta t), \tag{D.3}
\]

and

\[
\delta_T(t) = \delta_M(t) + \delta'(t - \Delta t). \tag{D.4}
\]
In [D.3], \( Q_o \) is defined by [2.2.5], and in [D.4], \( \delta_M \) is the prescribed mean divergence, and \( \delta' \) is the total divergence anomaly from the previous timestep.
**APPENDIX E**

**List of Symbols**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>c</td>
<td>oceanic free wave speed [=\left(gH_o\right)^{1/3}]</td>
</tr>
<tr>
<td>$c_a$</td>
<td>atmospheric free wave speed [=N/m]</td>
</tr>
<tr>
<td>$c_p$</td>
<td>specific heat at constant pressure</td>
</tr>
<tr>
<td>$e_s$</td>
<td>saturation water vapor pressure</td>
</tr>
<tr>
<td>$f$</td>
<td>Coriolis parameter</td>
</tr>
<tr>
<td>$F$</td>
<td>external atmospheric zonal momentum forcing</td>
</tr>
<tr>
<td>$g$</td>
<td>acceleration of gravity</td>
</tr>
<tr>
<td>$G$</td>
<td>external atmospheric meridional momentum forcing</td>
</tr>
<tr>
<td>$h$</td>
<td>oceanic upper layer depth anomaly</td>
</tr>
<tr>
<td>$\bar{h}$</td>
<td>mean oceanic upper layer depth</td>
</tr>
<tr>
<td>$H_o$</td>
<td>oceanic equivalent depth</td>
</tr>
<tr>
<td>$H_1$</td>
<td>reference oceanic upper layer depth</td>
</tr>
<tr>
<td>$H_s$</td>
<td>oceanic surface layer depth</td>
</tr>
<tr>
<td>$k$</td>
<td>wavenumber in $x$</td>
</tr>
<tr>
<td>$l$</td>
<td>wavenumber in $y$</td>
</tr>
<tr>
<td>$L$</td>
<td>latent heat of condensation</td>
</tr>
<tr>
<td>$m$</td>
<td>wavenumber in $z$</td>
</tr>
<tr>
<td>$N$</td>
<td>reference atmospheric buoyancy frequency</td>
</tr>
<tr>
<td>$p$</td>
<td>pressure</td>
</tr>
<tr>
<td>$q$</td>
<td>specific humidity</td>
</tr>
<tr>
<td>$q_s$</td>
<td>saturation specific humidity</td>
</tr>
<tr>
<td>$Q$</td>
<td>dimensional heating/unit mass</td>
</tr>
<tr>
<td>$Q_o$, $Q_l$</td>
<td>nondimensional heating</td>
</tr>
</tbody>
</table>
\( r \) oceanic dissipation coefficient
\( r_s \) oceanic surface layer momentum dissipation coefficient
\( R \) gas constant
\( s \) entropy
\( t \) time
\( T \) temperature
\( T_o \) tropospheric reference temperature
\( T_s \) oceanic surface temperature anomaly
\( T_s \) mean oceanic surface temperature
\( T_{sub} \) subsurface temperature anomaly
\( T_z \) mean oceanic vertical temperature gradient
\( u \) baroclinic zonal current
\( u_a \) zonal wind
\( \bar{u} \) baroclinic horizontal current vector
\( u_a \) horizontal wind vector
\( \bar{u}_{sfc} \) total surface layer horizontal current vector
\( \bar{u}_{sfc} \) mean surface layer zonal current
\( U \) baroclinic zonal transport
\( U_s \) frictional zonal transport
\( U_s \) frictional horizontal transport vector
\( v \) baroclinic meridional current
\( v_a \) meridional wind
\( \bar{v}_{sfc} \) mean surface layer meridional current
\( V \) baroclinic meridional transport
\( V_s \) frictional meridional transport
\( w \) vertical velocity at the base of the oceanic surface layer
\( w_a \) atmospheric vertical velocity
\( \bar{w} \) mean oceanic upwelling
\( x \) distance in eastward direction
\( y \) distance in northward direction
\( z \) distance in upward direction
\( z_T \) reference tropopause height
\( a \) coefficient relating local evaporation to local atmospheric heating
\( a_0 \) reference thermal expansion coefficient \([= 1/T_0]\)
\( a_s \) thermal dissipation coefficient of SST anomalies
\( \beta \) atmospheric convergence feedback efficiency factor
\( \beta_0 \) meridional gradient of the Coriolis parameter at the equator
\( \gamma_1 \) efficiency factor for surface temperature tendency related to mean upwelling advection
\( \gamma_2 \) efficiency factor for surface temperature tendency related to anomalous upwelling advection
\( \delta' \) anomalous wind divergence
\( \delta_M \) mean wind convergence
\( \varepsilon \) atmospheric dissipation coefficient
\( \rho \) oceanic reference density
\( \rho_0 \) atmospheric reference density
\( \tau(x) \) zonal surface stress
\( \tau(y) \) meridional surface stress
\( \zeta \) horizontal surface stress vector
\( \omega \) temporal frequency
\( \nabla \) horizontal gradient operator \([\partial/\partial x \hat{i} + \partial/\partial y \hat{j}]\)
\( \nabla^* \) horizontal divergence \([\partial/\partial x + \partial/\partial y]\)
APPENDIX F

List of Figures

Figures 2.1-2.3 Prescribed heating and resultant model winds, for (a) $\varepsilon = 0.1$, and (b) $\varepsilon = 0.2$.

Figure 2.4 (a) initial wind field, and (b) initial divergence field, for heating of the form $Q = \cos(\pi x/8)\cos(\pi y/8)$. (c) wind field, and (d) divergence field, after convergence feedback. Mean divergence field is assumed to be zero.

Figure 2.5 As in Fig. 2.4, except for a small-scale heating: $Q = \cos(\pi x/2)\cos(\pi y/2)$.

Figure 2.6 (a) windfield, and (b) divergence field, after convergence feedback, with prescribed mean convergence between $y=0.5$ and $y=1.5$, and divergence elsewhere (see text). Initial heating, wind, and divergence are the same as for Fig. 2.4.

Figure 2.7 (a) wind field, and (b) divergence field, after convergence feedback, with prescribed mean convergence for $x<0$, and divergence for $x>0$ (see text). Initial heating, wind and divergence are the same as for Fig. 2.4.

Figure 2.8 El Niño composite of eastern Pacific SST anomalies (after Rasmussen and Carpenter (1982)). Arrows indicate the times for which model results will be presented.

Figure 2.9 (a) composite SST anomalies, (b) composite wind anomalies, and (c) composite divergence anomalies, for AUG(-1). (d) predicted wind anomalies, and (e) predicted divergence anomalies, for the same time, using model I. (f) predicted wind anomalies, and (g) predicted divergence anomalies, for the same time, using Model II (see text). Scale for wind vectors and contours for divergence are the same as for the composites.

Figure 2.10 As in Fig. 2.9, except for DEC(-1).

Figure 2.11 As in Fig. 2.9, except for MAY(0).

Figure 2.12 As in Fig. 2.9, except for DEC(0).

Figure 2.13 As in Fig. 2.9, except for APR(1).

Figure 2.14 As in Fig. 2.9, except for AUG(1).

Figure 2.15a TW1 indices (see text), computed from composite data, model I output, and model II output. Time is indicated in years, from the beginning of year (-1) to the end of year(+1).

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Figure 2.15b  As in Fig. 2.15a, except for TW2 index (see text).

Figure 2.16a  Pressure anomalies at the location corresponding to Darwin, Australia, computed from the model I output and Model II output.

Figure 2.16b  Model pressure anomalies for the location corresponding to Tahiti.

Figure 2.16c  Composite pressure anomalies (after Rasmussen and Carpenter (1982)).

Figure 3.1  Isotherms along the equator in the Pacific Ocean (after Colin et al. (1971)).

Figure 3.2  Model upper layer depth departures from the mean depth (150m), for (a) March, (b) June, (c) September, and (d) December. x is measured in degrees of longitude east of the Greenwich meridian, and y is measured in degrees of latitude from the equator. Contour intervals are 20m.

Figure 3.3  Model surface currents for (a) March, (b) June, (c) September, and (d) December. Largest vector in (c) corresponds to about 75 cm/sec.

Figure 3.4  Model upwelling velocities for (a) March, (b) June, (c) September, and (d) December. Contour intervals are 10 cm/sec (heavy line is the zero contour).

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Figure 3.6  (a) NINO1 indices (°C), from model output (light line) and from composited data (heavy line) (b) same for NINO3 indices, (c) same for NINO4 indices.

Figure 3.7  The equatorial h-field across the basin, between months 132 and 168 (i.e. year(11)–year(14)) of the model run. Contours are 5 m.

Figure 3.8  (a) composite SST anomalies for AUG(-1), and (b) model SST anomalies for the same time. Contours are .25°C (heavy line is the zero contour).

Figure 3.9  As in Fig. 3.8, except for DEC(-1).

Figure 3.10  As in Fig. 3.8, except for MAY(0).

Figure 3.11  As in Fig. 3.8, except for DEC(0).
Figure 3.12  As in Fig. 3.8, except for APR(1).

Figure 3.13  As in Fig. 3.8, except for AUG(1).

Figure 3.14  Contributions to model NINO1 variability. In (a), $Q$ refers to the anomalous heat flux, $\bar{w}$ refers to vertical advection associated with the mean upwelling, and $\bar{w}$ refers to vertical advection associated with the anomalous upwelling. In (b), $u$ refers to the combined effect of all zonal advection terms.

Figure 3.15  As in Figure 3.14, except for NINO3.

Figure 3.16  As in Figure 3.14, except for NINO4.

Figure 3.17  $b(x,t)$, computed from the observed wind anomalies between August, 1981 and March, 1983.

Figure 3.18  Model predictions for NINO1, NINO3, and NINO4 (in °C). Time is in years, where year(0) corresponds with 1982.

Figure 3.19  NINO1, NINO3, NINO4, and other indices computed from observations between 1979 and 1983 (after Arkin et al., 1983).

Figure 3.20  (a) model SST anomalies corresponding to May 1982, and (b) observed SST anomalies for March–May, 1982 (after Arkin et al., 1983).

Figure 3.21  (a) model SST anomalies corresponding to December, 1982, and (b) observations for December 1982 – February 1983 (after Arkin et al., 1983)

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Figure 4.1  Coupled model predictions for NINO1 (heavy solid line), NINO3 (light solid line), and NINO4 (dotted line), during 90 years of simulated time.

Figure 4.2  Indices of eastern and central Pacific SST anomalies, computed from observations. (after Rasmussen and Carpenter, 1982).

Figure 4.3  Coupled model predictions for TW1 (heavy line), and TW2 (light line), during the 90-year simulation run.

Figure 4.4  Sequence of model SST anomalies (left panels) and corresponding wind anomalies (right panels), during the 90 year simulation, starting with December of year(30) and proceeding to December of year(35).
Figure 4.5  Equatorial h-anomalies (left panels), and b(x,t) (right panels) for the period between month 360 (year 30) and month 540 (year 45) of the 90-year simulation run. Contours for h are 5m in the lower two panels, and 10m in the upper panel. Contours for b are 5 nondimensional units in the upper two panels, and 10 units in the lower panel.

Figure 4.6  Contributions to the NINO1 variability between years(31) and (60) of the 90-year simulation run. CODES are the same as for Figure 3.14.

Figure 4.7  Upper panel: NINO1 (heavy solid line), NINO3 (light solid line), and NINO4 (dotted line) for year(0) to year(30) of RUN A. Lower panel: TW1 (heavy line), and TW2 (light line) for the same period of RUN A.

Figure 4.8  b(x,t) (left panels), and the equatorial h-field (right panels), for the period between year(20) (month 240) and year(30) (month 360) of Run A. Contours for h: .02m (lower), .01m (upper). Contours for b: .02 units (lower), .01 units (upper).

Figure 4.9  As in Fig. 4.7, except for RUN B.

Figure 4.10  As in Fig. 4.8 except for RUN B. Contours for h and b: 20m and 20 units, respectively.

Figure 4.11  As in Fig. 4.7, except for RUN C.

Figure 4.12  As in Fig. 4.8, except for RUN C. Contours for h and b: 20m and 20 units, respectively.

Figure 4.13  As in Fig. 4.7, except for RUN C2.

Figure 4.14  As in Fig. 4.8, except for RUN C2. Contours for h: 1m (lower), 2m (upper). Contours for b: 2 units (lower), 1 unit (upper).

Figure 4.15  As in Fig. 4.7 except for RUN D.

Figure 4.16  As in Fig. 4.8 except for RUN D. Contours for h and b: 10m, and 10 units, respectively.

Figure 4.17  As in Fig. 4.7, except for RUN D2

Figure 4.18  As in Fig. 4.8, except for RUN D2. Contours for h and b: 10m, and 10 units respectively.

Figure 4.19  As in Fig. 4.7, except for RUN E.
Figure 4.20  As in Fig. 4.8, except for RUN E. Contours for h and b: 10 m, and 10 units, respectively.

Figure 4.21  (a) SST anomalies, and (b) wind anomalies for December of year(20) in RUN E.

Figure 4.22  As in Fig. 4.7, except for RUN F.

Figure 4.23  As in Fig. 4.8, except for RUN F. Contours for h and b: 2 m, and 2 units, respectively.

Figure 4.24  As in Fig. 4.7, except for RUN G.

Figure 4.25  As in Fig. 4.8, except for RUN G. Contours for h: 1 m (lower). 2 m (upper). Contours for b: 2 units.

Figure 4.26  (a) SST anomalies, and (b) wind anomalies, for June of year(13) in RUN G. (c) SST anomalies, and (d) wind anomalies, for June of year(18) in RUN G.

Figure 4.27  As in Fig. 4.7, except for RUN H.

Figure 4.28  As in Fig. 4.8, except for RUN H. Contours for h and b: 10 m, and 10 units, respectively.

Figure 4.29  As in Fig. 4.7, except for RUN I.

Figure 4.30  NINO1, NINO3, and NINO4 indices for RUN J.

Figure B.1  Long-term means of (a) SST, (b) surface wind, and (c) divergence, for March.

Figure B.2  Long-term means of (a) SST, (b) surface wind, and (c) divergence, for May.

Figure B.3  Long term means of (a) SST, (b) surface wind, and (c) divergence, for August.

Figure B.4  Long-term means of (a) SST, (b) surface wind, and (c) divergence, for November.