On Equatorial Waves and El Niño. II: Effects of Air–Sea Thermal Coupling

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ABSTRACT

Dynamic readjustment of a stratified ocean model to wind perturbations leads to variations in sea surface temperature (SST) related to the early phases of the observed interannual warming of the tropical Pacific known as El Niño. The role that the atmosphere plays in determining the extent and strength of the SST warming is examined through numerical experiments with varying parameterizations for the atmospheric thermal response to SST anomalies.

A priori specification of the atmospheric temperature (even as a function of space and time) amounts to assuming infinite heat capacity for the atmosphere. A zero-heat capacity atmospheric model is constructed, in which the surface air temperature is balanced between the SST and a radiative equilibrium temperature. In the latter model, SST perturbations are damped through radiative relaxation from the atmosphere, rather than through direct cooling to the atmosphere. This greatly increases the lifetime of SST anomalies and increases their areal extent.

The effect of the atmospheric parameterization has on an upper ocean model for El Niño is examined. The model tests are conducted by imposing wind perturbations on simple mean states driven by constant winds. Westerly wind perturbations in the western part of the model basin excite Kelvin waves that propagate to the east. Under southerly mean winds, this Kelvin wave propagates to the east without any signal in the SST, but large SST anomalies are generated upon reflection of the Rossby waves. Much weaker changes in the southerly winds near the eastern coast produce SST anomalies that mimic those generated by the westerly wind changes. Such a counter-example to remotely forced Kelvin wave theories for El Niño also arises when the southerly stress anomaly is held off the coast by 200 km. Sea-level changes associated with the westerly and southerly wind perturbations are markedly different. The rapid adjustment of the atmosphere to the ocean appears to be a necessary condition for successful simulation of the El Niño warming.

1. Introduction

The phenomenon of El Niño is an anomalous warming of the sea surface in the eastern tropical Pacific Ocean which occurs both annually and (with much more catastrophic behavior) interannually. Dynamical hypotheses have been put forth to explain these occurrences, although the results which have been reported to date are largely lacking in thermodynamic analysis (cf. Busalacchi and O'Brien, 1981). In a companion paper (Schopf and Harrison, 1983, hereafter SH) a thermally active model was used to examine the tropical oceans' response to wind variations in the western part of an idealized ocean basin, and it was found that the Kelvin wave processes modeled with a simple reduced gravity approach can show some essential features of an El Niño event—the first SST change to a remote wind forcing was found to occur at the eastern end of the equator when a southerly mean wind stress was imposed. The model was only successful in simulating the earliest phases of El Niño, and did not give the broad warming that is found in composite analyses of the climatological data as in Rasmusson and Carpenter (1982). That result was reached by resorting to a classical approach to air–sea interaction, the implications of which will be examined in more detail here.

Numerical models of the ocean are often driven by the specification of an atmospheric temperature and a net heat flux into the ocean parameterized as in Haney (1971)

\[ Q_{so} = K(T_a - T_o), \]

where \( Q_{so} \) is the surface flux, \( T_a \) a representative air temperature and \( T_o \) is the sea surface temperature. In Eq. 1 \( K \) is a constant, usually having a value of 30–60 W m\(^{-2}\) K\(^{-1}\) when \( T_a \) is used as the surface air temperature (Haney, 1971; Dickinson, 1981). Estimates of \( K \) have been made from atmospheric general circulation model sensitivity studies, and are also found to lie in this range (Shukla, personal communication, 1982). This parameterization can be viewed as the first order linearization of the dependence of the surface heat flux on \( T_a \) and \( T_o \).

When ocean models are run with such surface fluxes, the dynamics of the ocean drives the SST away from equilibrium with the atmosphere, but the air is not allowed to vary from its pre-determined evolution. The air temperature is often chosen to be a func-
tion of time and space, sometimes determined from observations; in which case, the SST takes on much more "realistic" patterns than obtained when modeling with $T_a$ constant. However, such modifications do not change the property that SST departures from the air temperature have no influence on the atmosphere. In cases with this parameterization (to be presented below), the ocean dynamics drive air–sea temperature differences of 5–10°C in upwelling zones and other regions of strong advective activity. This is in marked contrast to the observational evidence from Hastenrath and Lamb (1977) presented in Fig. 1: throughout the tropics, the air–sea temperature difference is on the order of 0.5–1.0°C, even in the regions of strong upwelling activity. While 5–10°C differences are not uncommon in midlatitudes, they are associated with strong horizontal advection of continental air over the ocean, particularly in the winter. For the present study of the tropical oceans, it will be advantageous to postpone consideration of these complications and study the limiting case of no atmospheric transport. Other calculations are currently underway which account for a mean atmospheric advection of heat. While the results do not indicate an insignificant effect due to advection, the solutions discussed here constitute limiting cases and are helpful for the analysis of the more complex result.

Dickinson (1981) has pointed out the essential difficulty. The atmospheric heat capacity is very much smaller than that of the ocean's mixed layer. The small atmospheric heat capacity means that the air temperature closely follows the ocean surface temperature. Instead of using the infinite heat capacity

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**Fig. 1.** Climatic data for the eastern tropical Pacific from Hastenrath and Lamb (1977). Sea surface temperature for (a) April and (b) October (°C). Ocean temperature minus air temperature for (c) April and (d) October (contour interval 0.1°C).
implicit when an \textit{a priori} specification of the atmospheric temperature is made (even spatially and temporally varying) it is more appropriate to model the atmospheric interaction as though the atmosphere had zero heat capacity.

If we make the assumption that the tropical atmosphere is on or near the moist adiabat, there is one degree of freedom in the vertical with which to describe the thermal profile. For convenience, we will choose this to be the surface temperature, although any other choice will give the same results. Dickinson's (1981) analysis can be simplified by considering the following coupled ocean–atmosphere model

\begin{equation}
C_a \dot{T}_a = -K(T_a - T_o) - K'T_a + Q_a,
\end{equation}
\begin{equation}
C_o \dot{T}_o = K(T_a - T_o) + Q_o.
\end{equation}

Here, $C_a$ and $C_o$ are the heat capacities of the atmosphere and the ocean mixed layer; $K'$ is the atmospheric radiative feedback constant. The $Q_a$ and $Q_o$ stand in for atmospheric and oceanic heating sources, including radiation and advection. Here $Q_o$ includes upwelling and the connection with the deep ocean, but it excludes those surface fluxes parameterized through $K(T_a - T_o)$. Dickinson (1981) reports values for $C_a = 0.45$ W m$^{-2}$ year K$^{-1}$, $C_o = 10$ W m$^{-2}$ year K$^{-1}$, $K = 45$ W m$^{-2}$ K$^{-1}$ and $K' = 2.4$ W m$^{-2}$ K$^{-1}$.

Since $C_a/C_o \approx 0.045 \ll 1$, the solutions to (2)–(3) may be constructed by expanding $T_a$ and $T_o$ in this small parameter. To first order, the following equations represent the system with a zero heat capacity atmosphere

\begin{equation}
K(T_a - T_o) = Q_a - K'T_a,
\end{equation}
\begin{equation}
C_o \dot{T}_o = K(T_a - T_o) + Q_o.
\end{equation}

With constant atmospheric heating and without the ocean driving the system away from equilibrium ($Q_o = 0, Q_a$ constant), the outgoing infrared flux matches the incoming short wave at radiative–convective equilibrium, which has the radiative equilibrium surface temperature $T_e$ given by

\begin{equation}
T_e = \frac{Q_a}{K'}.
\end{equation}

Under time-varying conditions (6) serves to define $T_e$, and (4)–(6) give

\begin{equation}
C_o \dot{T}_o = KK' \frac{T_e - T_o}{K + K'} + Q_o
\end{equation}
\begin{equation}
= K(T_e - T_o) + Q_o,
\end{equation}

where

\begin{equation}
K_r = \frac{KK'}{K + K'}.
\end{equation}

We have therefore replaced (1) with

\begin{equation}
Q_{oa} = K_r(T_e - T_o).
\end{equation}

Thus, under the assumption of zero heat capacity for the atmosphere, the ocean is coupled with the radiative equilibrium temperature for the atmosphere, by a coefficient $K_r$. With a mixed layer of 25 m depth, $C_o/K_r$ — the e-folding time for SST response to perturbations in $Q_o$ — is $\sim 500$ days. The infinite heat capacity approach had the ocean coupled to the specified atmospheric temperature with the coefficient $K$ and a time scale of 50 days. Thus, models using a zero heat capacity atmosphere should show substantially larger changes in SST caused by ocean dynamics. Perturbations should be longer lived and have more time to cover significant surface area.

The atmospheric temperature has been eliminated from the system and is free to seek its own value, based on a balance between the ocean temperature and the radiative imbalance at the top of the atmosphere. To first order the air–sea temperature difference can be evaluated as

\begin{equation}
(T_a - T_o) = (T_e - T_o) \frac{K_r}{K}.
\end{equation}

The air–sea temperature difference is on the order of $\gamma_0$ of the difference between the SST and the imposed equilibrium atmosphere. If the dynamics of the ocean causes a 5°C cooling by upwelling or advection, the air–sea temperature difference should be on the order of 0.5°C. These values are consistent with the data given in Hastenrath and Lamb (1977): tongues of cold water are seen along the Peru coast with SST $\sim 19–21°C$. In the mid-Pacific the SST is $\sim 28–29°C$, probably close to equilibrium. The coastal water is therefore $\sim 8–10°C$ cooler than the equilibrium value. Air–sea temperature differences are on the order of 1°C.

In this treatment of the atmosphere for ocean modeling, the air temperature is determined by the fluxes across the ocean surface and not imposed by the boundary conditions. The balance temperature ($T_e$) is a convenient description of the atmospheric heating: in essence the specification of its value is a specification of the atmospheric heating rate $Q_{oa}$.

In previous ocean modeling studies of El Niño, we have found difficulty in establishing large areas of anomalously warm water (SH). Attempts at simulating the seasonal cycle in the tropical Pacific with an ocean circulation model had to resort to a seasonal specification of surface air temperature as a function of latitude and longitude in order to achieve adequate results for SST (Siegel and Philander, personal communication, 1982). In both cases the surface flux was parameterized as in (1), and in view of the above analysis, these results are not surprising. To explore the implications of the low heat capacity of the at-
mosphere, we have undertaken a repetition of the previous calculations, but with zero heat capacity for the atmosphere.

The first two cases examine the effect of the low thermal inertia of the atmosphere on the warming which occurs after the excitation of a Kelvin wave in the western portion of the basin. The very realistic SST anomaly generated under the assumption of zero atmospheric heat capacity is found to be due to the Rossby wave propagation from the coast and the anomalous advection associated with it. Since this mechanism can be excited by changes in the meridional component of the wind stress near the eastern boundary, a case is examined in which the strength of the southerly mean stress is weakened slightly. This too, produces strong, broad warming. There is some indication that the southerly stress along the South American continent does not significantly change during an El Niño. Therefore, a fourth case is examined, in which the southerly wind is relaxed, but the perturbation vanishes near the coast. Finally, the influence of the atmospheric parameterization on the case with mean easterly winds is presented.

2. Model results

The model used is a nonlinear primitive equation model for the upper few hundred meters of the ocean as described in SH (an extension of Schopf and Cane, 1983). The surface layer is treated as the ocean’s well-mixed layer, and its depth is determined by advection, upwelling and entrainment. Mixing is calculated as in extensions of the bulk energy balance models of Kraus and Turner (1967). SST is determined by advection, surface heating and entrainment cooling.

Five cases are reported for runs in a rectangular equatorial basin that spans 75° of longitude and 40° of latitude. As in SH, a constant mean wind of 0.5 dyn cm⁻² is used to spin up the model for 360 days, at which time the solutions have reached a relatively steady state—the changes induced by the wind perturbations clearly dominate subsequent evolution. In the first four cases, this wind is southerly; the fifth case has easterly mean winds. At the end of the spin-up period, the mean wind is held steady, and a perturbation wind field is superimposed on the mean. Table 1 lists the cases and their forcing parameterization.

<table>
<thead>
<tr>
<th>Case</th>
<th>Mean wind</th>
<th>Perturbation wind</th>
<th>Heat flux equation</th>
<th>( T_d ) ( (°C) )</th>
<th>( K_v ) ( (W m^{-2} K^{-1}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>S</td>
<td>Westerly 0.5 dyn cm⁻² confined to western half of basin</td>
<td>(1)</td>
<td>[29]</td>
<td>[30]</td>
</tr>
<tr>
<td>2</td>
<td>S</td>
<td>Westerly 0.5 dyn cm⁻² confined to western half of basin</td>
<td>(9)</td>
<td>29</td>
<td>2.5</td>
</tr>
<tr>
<td>3</td>
<td>S</td>
<td>Westerly 0.5 dyn cm⁻² confined to western half of basin</td>
<td>(1)</td>
<td>variable</td>
<td>[30]</td>
</tr>
<tr>
<td>4</td>
<td>S</td>
<td>Northerly 0.125 dyn cm⁻² over entire basin</td>
<td>(9)</td>
<td>29</td>
<td>2.5</td>
</tr>
<tr>
<td>5</td>
<td>S</td>
<td>Northerly 0.125 dyn cm⁻² reduced to zero at east coast</td>
<td>(9)</td>
<td>29</td>
<td>2.5</td>
</tr>
<tr>
<td>6</td>
<td>E</td>
<td>None</td>
<td>(9)</td>
<td>29</td>
<td>2.5</td>
</tr>
</tbody>
</table>

FIG. 2. Case 1 SST and mixed layer depth after 360 days of spin-up to 0.5 dyn cm⁻² southerly winds, \( K = 30 \ W m^{-2} K^{-1} \) (contour intervals 1°C and 5 m).
an additional wind component was imposed—a westerly stress of 0.5 dyn cm\(^{-2}\) confined to the region from 10° to 37.5°E longitude and from the southern to northern walls. The surface flux was parameterized as in (1) with \(K = 30 \, W \, m^{-2} \, K^{-1}\), and \(T_a = 29^\circ C\). During spin-up, cold upwelled water formed a very short tongue extending out 5° from the upwelling.

**FIG. 3.** Case 1 SST changes 40, 80 and 180 days after the onset of a westerly wind perturbation. Fig. 2 gives the initial conditions for these changes.

**a. Mean southerly winds: Westerly remote perturbation.**

In SH, we examined the response of the SST to anomalous westerly wind stress in the far western part of an equatorial basin. That run constitutes Case 1 for this study. The basin extended over 75° of longitude and spanned the equator from 25°S to 15°N. A constant mean wind of 0.5 dyn cm\(^{-2}\) blew from the south. After 360 days of spin-up to this forcing, SST

**FIG. 4.** Case 2 (a) SST, (b) mixed layer depth, and (c) surface currents after 360 days of spin-up to 0.5 dyn cm\(^{-2}\) southerly winds, \(K = 2.5 \, W \, m^{-2} \, K^{-1}\) (contour intervals 1°C and 5 m). Current speed is proportional to arrow length with single arrows representing speeds from 2 to 20 cm s\(^{-1}\); double arrows give currents greater than 20 cm s\(^{-1}\).
region south of the equator. Kelvin waves excited by the perturbation propagated to the east, and upon reaching the cold tongue, caused an advective increase in SST as the cold water was displaced. Although the initial cold tongue was quite small, the temperature along the coast was reasonable, ~22°C. Fig. 2 gives the SST and mixed layer depth after 360 days of spin-up to the steady southerly winds. These fields give the initial conditions for the perturbation run. Fig. 3 presents the change in SST at 40, 80 and 180 days after the onset of the westerly wind perturbation.

Case 2 is a repetition of the model calculation using (9) with an air–sea coupling constant $K_r = 2.5 \text{ W m}^{-2} \text{ K}^{-1}$, and with $T_r = 29.0^\circ \text{C}$. In order to provide approximately the same diagnostic depth for the mixed layer, the friction velocity was reduced from 0.864 to 0.74 cm s$^{-1}$. In this run, the spin-up of the initial state is markedly different from the previous run. Fig. 4 shows the SST, mixed layer depth and surface currents achieved after 360 days of spin-up with constant southerly winds of 0.5 dyn cm$^{-2}$. The tongue of cold water is much larger, with a substantial area covered with water at less than 24°C. It is also interesting to note that the temperature along the coast is only slightly different—a minimum of ~20°C.

Heat budget calculations for the surface layer reveal that the primary balance establishing the cold tongue is advection of cold water from the upwelling region balancing surface heating. The initial stratification and the strength of the wind stress determines the temperature of the water that reaches the surface at the upwelling zone, and establishes the minimum temperature for the tongue. Once the water comes in contact with the surface, the interactive flux warms it back to the equilibrium value. This balance is the same as was obtained in Case 1, but the much longer decay time associated with $C_{0}/K_r$ means that the advective effects are carried much farther from shore. It was also previously found that the primary temperature change mechanism associated with the Kelvin waves was advective. In Case 2, there is a much greater opportunity for advective warming, since the tongue is much larger.

After the onset of the westerly stress anomaly, the changes in the surface height, the thermocline depth and the velocity shear are similar to those found with the stronger air–sea coupling. Fig. 5 shows equatorial longitude–time sections of these quantities. As in Case 1, (cf., SH, Fig. 5) the sea level changes with the gravest mode Kelvin wave propagating from east to west in ~30 days. The shear mode wave is seen in $u_1 - u_2$, taking nearly 100 days to cross the eastern half of the basin. The pycnocline is depressed by the gravest mode and is drawn up by the shear mode. The dynamical changes induced by the wind perturbation are relatively unaffected by the air–sea coupling—although the SST behaves quite differently. The results in SH concerning these fields remain valid, particularly the distinction between dynamic height changes and changes in sub-surface isotherms.

The Kelvin waves and the Rossby wave reflection can be seen in the surface height field. Fig. 6 shows the changes in surface height at 40, 80 and 180 days after the onset of the perturbing wind. At 40 days, the Kelvin wave front has just reached the coast and has begun to propagate down the coast. By 80 days, the reflected Rossby waves are apparent, and after 180 days of perturbing winds, the response is quite mature. These dynamic signals are quite symmetric about the equator, giving evidence that the asymmetric mean flow has little nonlinear influence on the gravest mode waves.

The changes in SST are given in Fig. 7. As in the previous case, the first indications of SST change due to the waves occurs at the eastern coast, as the first Kelvin wave arrives. Subsequent evolution of the anomaly is quite different, however. The gravest mode Kelvin and Rossby wave pair provides substantial warming, propagating from the coastal region out into the basin. Large areas are now seen with warming in excess of 1°C, much more like those anomalies reported in the composite data of Rasmusson and Carpenter (1982). It should be noted that
these patterns look like only the early phases of El Niño, not giving the broad mid- and late phases that straddle the equator and extend much farther westward.

Heat budget analysis has been carried out on the mixed layer over 5° by 3.3° subregions south of the equator, extending out from the eastern boundary. Since the heat content \( (h_i T_i) \) can change through a warming of the surface temperature, and through changes in the mixed layer depth, these effects can be separated. Fig. 8a shows the time history of \( h_i \partial T_i / \partial t \) for three regions. Fig. 8b gives \( T_i \nabla \cdot (v_i h_i) \). In the latter, the gravest mode Kelvin waves cause a mass convergence that can be seen propagating to the east, starting on about day 20 in the westernmost region, and arriving at the coast on day 29. The reflected Rossby wave is seen to propagate back across the basin, arriving at the westernmost bin in about 56 days. The travel time back westward is about three times as long as that taken by the incoming Kelvin wave. Fig. 5 allows the identification of the shear mode Kelvin wave arriving at the easternmost box at about 100 days. Since the boxes extended from 1.7° to 5.0°S, the narrow shear mode Kelvin wave does not show up while propagating

\[ \text{FIG. 6. Changes in surface dynamic height 40, 80, and 180 days after onset of westerly perturbation (Case 2). Contour interval 2 cm.} \]

\[ \text{FIG. 7. SST changes 40, 80 and 180 days after the onset of a westerly wind perturbation (Case 2). Fig. 4 gives the initial conditions for these changes.} \]
to the east, and it is its coastally trapped southern extension that is seen in Fig. 8b. Propagation of the shear mode Rossby waves is not seen within the 180 days of the run examined here.

The temperature behavior in Fig. 8a is interesting because it shows no evidence at all of the incoming Kelvin wave. Warming is only accomplished by the westward propagating first-mode Rossby wave. The initial temperature field seen in Fig. 4 has only a meridional temperature gradient in the region considered, and the Kelvin wave has no meridional velocity component. Thus the Kelvin wave is able to propagate through the region undetected in SST. The changes in the horizontal currents induced by the gravest-mode Kelvin–Rossby response persist long enough and with enough organization to significantly change SST over 2000–3000 km.

In Case 2, the warming is due to the advective change in surface temperature that is induced on the mean state by the wave response. Two features contribute to the enlarged size of the anomalous region: first, the anomaly that is generated is not so quickly damped back to the external boundary temperature and, therefore, can be longer lived (spread further). Second, the mean state has a much larger region of surface thermal gradient so that the effects of $v \cdot \nabla T$ can be seen over a greater distance. The relative importance of these two contributing effects is an important question related to the specification of the upper boundary condition, because with suitable manipulation of the pre-specified air temperature, the mean thermal structure obtained in a model run with (1) can be identical to that obtained in a model run with (9). If the anomaly is independent of the coupling (dependent primarily on the ocean’s mean state), the use of (1) with variable air temperatures can be viewed as an acceptable boundary condition for the study of SST anomaly evolution.

b. Southerly mean wind: Spatially varying fixed air temperature—Case 3.

To obtain a mean thermal state using (1) that is identical to that obtained from (9), the only requirement is that when the model is in equilibrium, the heat flux $Q_{oa}$ be the same in the two runs. Equating $Q_{oa}$ in (1) and (9) gives a diagnostic equation for the spatially variable air temperature; this temperature is then held constant while a wind change is introduced. The advective perturbation will then go through the same initial state, but under the different coupling law. If the atmospheric parameterization is unimportant, the anomaly generated in such a modified model should look like that found in Case 2.

An analytic examination of the system under uniform advection provides information on the results to be expected from a full calculation with the numerical model. First, assume that the radiative equilibrium temperature is constant, so that along the streamlines of the flow the thermal balance as given with (9) is

$$UCo T_{oa} + K T_o = 0,$$

where the temperature is referenced to the radiative equilibrium value. For a constant mean speed $U$, the solutions are simply

$$T_o(x) = T^* \exp(-K_x/xUCo),$$

$$T_o(x) = T^*(1 - K_x/K) \exp(-K_x/xUCo).$$

We suppose that some external process sets $T^*$, the boundary condition at $x = 0$. The use of an infinite heat capacity atmosphere that would have the same state as given in (12) gives an equation for the SST (denoted here as $T_2$) which is

$$UCo T_{2a} + K T_2 = T^*(K - K_o) \exp(-K_o x/UCo).$$

The $T_2$ is given by a general and forced solution.
The general solution has exponential decay with a length scale \( L_2 = UC_0/K_1 \), while the forced solution decays over the longer scale characteristic of the zero heat capacity model: \( L_o = UC_0/K_2 \). When the two models have the same boundary temperature and velocity, the general solution has zero amplitude, and the solutions for the two models are identical.

There are two ways of introducing anomalies in this system: either the boundary temperature can be changed, or the advection speed can be altered. When the boundary temperature is changed, the zero heat capacity atmosphere leads to an anomaly with exponential decay over length scale \( L_o \). In the infinite heat capacity model, the general solution takes on a non-zero amplitude, and the anomaly decays over a length scale \( L_2 \)—the forced solution is unchanged. The anomalies are not influenced by the mean state; they have the same length scales as the mean states would have under spatially uniform atmospheric temperatures (Fig. 2a or Fig. 4a).

When the current is perturbed, the anomalies behave in slightly different fashion, but the anomalies under the infinite heat capacity atmosphere are much weaker. If the speed is perturbed to \( U(1 + u') \), the two models will give new temperatures \( T_o + T'_o \) and \( T_2 + T'_2 \). The anomalies are given by

\[
T'_o = T^* \exp[-x/(1 + u')L_o] \\
\times \{ 1 - \exp[-xu'/(1 + u')L_o] \},
\]

\[
T'_2 = \frac{u'rT^*}{(1 - r - ru')} \exp(-x/L_o)[-1 - \exp(-x/L^*)],
\]

where \( r = K_o/K \). Here \( u' \) may be \( O(1), r \ll 1 \). A short length scale \( L^* \) is given by

\[
L^* = \frac{L_0r(1 + u')}{(1 - r - ru')},
\]

\[
= \frac{L_0(1 + u')}{(1 - r - ru')}.
\]

Fig. 9 presents the solutions to (15) and (16) for small and large perturbations to \( U \). Note that the anomalies under the infinite heat capacity atmosphere are not only of smaller length scale, but have a substantial reduction in their magnitude.

The extension of the foregoing analysis to the numerical model results is straightforward. A run was made, (Case 3), which was identical to that in Case 2 for the first year of spin-up (weak coupling, steady southerly winds) giving the same SST as shown in Fig. 4a. The surface air temperatures were then computed diagnostically from (10). The model integration was continued for an additional year with the surface flux law (1), but with the diagnostically-determined air temperatures. Because of small imbalances still present in the system when the air temperatures were determined, the SST did change slightly over this second year, giving the pattern shown in Fig. 10.

A westerly wind anomaly of 0.5 dyn cm\(^{-2}\) was then introduced as previously, and the resulting SST anomalies are shown in Fig. 11 at 180 days after the onset of the anomalous winds. The results exhibit behavior which is strongly consistent with the foregoing linear analysis. Not only are the anomalies much smaller than those arrived at with the weak coupling (Case 2), but because the gradients in Fig.
wind change but takes at least 80 days to show significant areas with water 0.5°C warmer than its initial value. The distribution of the SST anomaly is consistent with the propagation of the gravest-mode Rossby wave front from the coast, its effect modulated by the initial thermal gradients.

In this case, changes in the SST are sought by altering the strength of the upwelling, and hence the depth and temperature of the water that reaches the

10 are much smaller than those found from Case 1, the anomalies generated here are much weaker than those shown in Fig. 3.


The change in parameterization of the air–sea exchange of heat led to dramatic changes in the SST distribution, and the use of (9) produced an initial distribution of SST which was much closer related to the observed mean state of the tropical eastern Pacific (cf. Hastenrath and Lamb, 1977). The SST was cooled 8°C by a 0.5 dyn cm⁻² southerly wind component. This change of 16°C/dyn cm⁻² indicates the possibility of an SST change of 1–2°C with a small change in the southerly component of the wind stress (~0.5 dyn cm⁻²). To investigate this possibility in greater detail, the model was run for 360 days with a southerly wind stress of 0.5 dyn cm⁻². The southerly stress was then reduced to 0.375 dyn cm⁻² over the entire domain, and the run was continued for another 180 days. This case is undertaken as a mechanistic study of the response of the ocean—the available surface wind data is insufficient to assess how realistic this wind anomaly is.

The state of the ocean at the onset of the wind perturbation was identical to that shown in Fig. 4. Subsequent changes in the SST are shown in Fig. 12. The magnitude and shape of the SST changes due to the perturbation in southerly wind seen here is the same as that induced by the stronger change in the westerly wind component acting remotely. After 40 days the first strong warming signals are seen, near the coast at 50°S where a 1.9°C increase is seen in SST. By the end of the 180 day perturbation period, a large tongue of warming is seen, with a maximum warming of 2.1°C and a large area covered with water more than 1°C warmer than the original state. The warming does not occur simultaneously with the

FIG. 11. Case 3. Change in SST 180 days after onset of perturbing westerly winds. Fig. 10 gives the initial conditions for these changes.

FIG. 12. Case 4. SST changes 40, 80 and 180 days after the onset of a northerly wind perturbation of 0.125 dyn cm⁻² applied over the entire domain. Fig. 4 gives the initial conditions for these changes.
surface. Advective changes also occur which slow down the water's travel away from the upwelling region. This deceleration means that the water receives heat from the atmosphere for a longer period of time before it reaches a particular longitude; the cold tongue can be reduced by this mechanism as well.

The sea level changes due to the northerly wind perturbation are much smaller than those seen from the westerly perturbation above. Fig. 13 shows the changes in sea level at 40, 80 and 180 days. Whereas the incoming Kelvin wave causes surface height anomalies greater than 10 cm in a clear linear propagation (cf. Figs. 5 and 8), the northerly perturbation causes no surface height changes in excess of 4 cm. This may be understood as being due to two causes. First, the linear response to a symmetric meridional stress is projected onto the higher-order, anti-symmetric equatorial Hermite modes (Moore and Philander, 1977). These modes have relatively smaller amplitudes distributed over a longer meridional extent. At the equator they have zero amplitude. In addition, the northerly perturbation used to produce these SST changes is much smaller than that required with the westerly stress anomaly.

d. Southerly mean winds: Modified northerly wind perturbation—Case 5.

In the above result, the wind changed at the coast, and the coastal upwelling was modified. This altered the source temperature of the water that was advected outward to form the tongue and thereby led to an apparent warming of the SST. Observational evidence for such a change in the coastal wind field related to El Niño is unconvincing and does not appear to be supported by the data reported in Rasmussen and Carpenter (1982). On the other hand, a wind stress change of this magnitude represents a wind speed change of only 10–12%, which is difficult to verify from the limited observations in the region. This is particularly true if the coastal wind data is neglected. South of the equator, just off the American coast, there is a substantial data void. To see whether changes confined to the central ocean area can have similar influences on SST, the above case was repeated but with the wind remaining steady along the coast at 0.5 dyn cm⁻². To connect the coastal region to the central ocean, where the wind was reduced to 0.375 dyn cm⁻², the wind field was linearly interpolated over a strip 2° wide. As before, the initial condition for this experiment was a spin-up for 360 days to a southerly wind of 0.5 dyn cm⁻², with resulting SST as shown in Fig. 4. The changes in SST that result following the perturbation are shown in Fig. 14.
Within 2° of the coast, the results in Fig. 14 are significantly altered from those in Fig. 12. There is almost no change in the coastal temperature, indicating that the coastal upwelling is strongly trapped and that the same source of cold water is present right at the wall. Once outside the coastal zone, however, the two figures are remarkably similar. The changes in SST that have the broad, tongue-like character between 45 and 70° longitude are almost identical. The processes that lead to the SST changes respond to the larger-scale wind field perturbations, rather than the details of the coastal wind activity.

e. Easterly mean winds—Case 6.

The southerly wind case does not produce the broad equatorial warming seen in Rasmussen and Carpenter's (1982) analysis of El Niño. Although there is cross-equatorial flow, the currents from the upwelling region take on a strong zonal component before reaching the equator, and the cold water is advected out across the basin and warmed before crossing the equator. Another process is apparently needed to cool the equatorial surface waters. Horizontal transport of heat by the atmosphere through the trades could be one candidate mechanism, as could the influence of an easterly component of the surface stress. In SH and in Schopf and Cane (1983) strong cooling of the equatorial SST was produced with easterly mean winds. Before attempting to analyze the combined response to the total tropical wind field, we will consider the modifications to our previous result induced with easterly mean driving.

When the model is spun up with easterly winds, the basic state is substantially different from that obtained in either SH or Schopf and Cane (1983). Fig. 15 shows the SST and mixed layer depth attained after 360 days of spin-up. Fig. 16 presents the equivalent figure from SH. The first feature that is noticed is that the SST pattern with the modified atmospheric parameterization (9) is much broader and smoother. The fronts reported in Schopf and Cane (1983) appear here to be much further poleward, very much weaker, and do not break down into the series of mesoscale eddies identified there. The front in mixed layer depth is also on this broad scale. The front in Schopf and Cane (1983) was identified with two features of the combined mixed layer–ocean dynamics system: in the presence of mean vertical currents, the equilibrium depth is one half of the Monin–Obukhov diagnostic depth, and the surface cooling leads to increased surface heating which shallows the diagnostic depth itself. In the present model the SST reaches a value quite close to that in the previous study, but the much smaller surface heat flux calculated using (9) decreases the frontogenesis due to the second mechanism. Apparently this effect acted more strongly in the earlier study than was expected.

It is also interesting to note that the SST developed along the equator is only 3–4°C colder than that in the previous model. It is not the surface flux that is determining the SST, but rather the degree of upwelling that can be supported and the depth from which the surface water comes. In this model, under steady conditions, there is no net mass flux when integrated over depth. The Ekman divergence at the equator is compensated by inflow in the second layer, and the water that arrives at the surface has its source in the upper part of the second layer. The conditions at higher latitudes set the thermal structure in this layer in the tropics, and it is this structure, coupled with the wind-driven circulation that essentially establishes the SST. Changing the surface flux parameterization changes the degree to which this subsurface water is warmed after it is incorporated into the mixed layer, but it has little to do with its source condition. Thus the ocean drives the SST, the air temperature is determined by (10), and the surface heat flux as in (9).

3. Discussion

These results are in a certain sense our first attempt at a coupled ocean–atmosphere system model, albeit with perhaps the simplest atmospheric specification.
For analyzing short time-scale changes in the tropical ocean, the approximation of a zero heat capacity for the atmosphere seems a more meaningful one than the infinite heat capacity model commonly used. Under this new modeling assumption, the resulting SST is more strongly dependent on the internal ocean model dynamics because it is weakly coupled to an atmosphere radiative equilibrium temperature rather than strongly coupled to an apparent air temperature representing the surface. This leads to quite different interpretations of the important ocean physics that play a role in such features as El Niño. Previously, we found the model's second (or shear) mode to play a significant role in the SST response to variable wind stress, and found significant complications when trying to predict its behavior, since it was so susceptible to nonlinear destruction. The gravest mode response was rather linear, but only changed the SST by way of its horizontal advective effects. While we find the same properties for the dynamic features of the modes, their relative roles in modifying SST change dramatically. Because the advective perturbations can be much longer lived, the first mode leads to the dominant SST change, and the Rossby wave reflection enters as a primary actor. The shear-mode Rossby waves are much too slow and small to play a significant role. Under these circumstances, the change in tropical SST appears to be much more strongly connected to the dynamic quantities calculated in linear, reduced-gravity models (cf., Busalacchi and O'Brien, 1983).

The comparison between the two results indicated the magnitude of the change in model ocean results which may be expected by going from fixed air temperature to a full air-sea coupled model. While the oversimplification used here leads to net surface heat fluxes that are quite a bit smaller than those observed in Hastenrath and Lamb (1978), the parameterization is intended to represent a linearization of the surface heat balance about some mean state, and to give the change in surface heating associated with a change in the SST and air temperatures. Additional heating sources that are not influenced by the SST anomalies are probably active; these will have no impact on the longevity and extent of the SST anomalies, except insofar as they alter the initial state through which dynamically-induced perturbations act.

The atmospheric model used here is most limited in its neglect of horizontal transports. The above parameterization for the atmosphere will have properties essentially similar to more complete atmospheric models in the tropics, provided these atmospheric transports are ignored. The Held and Suarez (1978) model is one example. In their model, the tropical atmosphere is treated with two layers and with parameterizations of the radiative, convective, latent and sensible heating. Perturbations in the atmospheric temperatures are damped through radiative feedback from each layer. The surface fluxes of heat are parameterized as linearly dependent on the difference between the surface air temperature and the ocean temperature. Because the tropical atmosphere is undergoing convective adjustment, the vertical temperature profile is determined by the moist adiabat. That is, there is actually only one degree of freedom for the temperature profile which describes the surface temperature and the two-layer temperatures. If we ignore the change in the adiabatic lapse rate with variations in the mean atmospheric temperature, the three temperatures which operate in the Held and Suarez model are linearly dependent on one thermal variable, which we may take to be $T_a$, the surface temperature as used in (1). Therefore, their atmospheric model reduces to Eq. (2).

The damping of SST perturbations is made by the radiative relaxation of the atmosphere in its response to the constant external radiative input. The above results show sensitivity to the size of this feedback parameter, at least when varied by an order of magnitude. The value used for $K_b$ by Dickinson (1981) of 2.4 W m$^{-2}$ K$^{-1}$ was based upon the results of the CO2 study of the NAS (1979). They reported a value of 1.7 ± 0.8 W m$^{-2}$ K$^{-1}$. Dickinson (1981) increased this value to account for the fractional coverage of the globe by the ocean. North and Coakley (1979) performed a least squares fit to satellite data on infrared flux (Ellis and Vonder Haar, 1976). Based on the large
scale Fourier–Legendre modes for the atmosphere, they found best agreement for $K_r = 2.09 \text{ W m}^{-2} \text{ K}^{-1}$. They were able to obtain good global distributions of temperature with the entire outgoing IR flux parameterized by $A + K_r T$. The number used here is consistent with these values, if perhaps too large. For these results, the important distinction is between the 30 W m$^{-2}$ K$^{-1}$ used previously and the small value used here.

The results with varying values of the feedback constant have shed light on the nature of the El Niño warming and its relationship to the atmospheric response. The initial spin-up to southerly winds with the two values of $K_r$ (2.5 and 30) supports the idea that the surface temperature in the coastal upwelling zone is set by the depth from which the upwelled water is drawn and the thermal profile in the deeper ocean. For these cases, run over no more than a few years, the primary process determining this deep profile is the specification of the initial profile and the sloshing motion of the second model layer as it comes into adjustment. Over longer integrations, the maintenance of the second layer profile is of concern and is a question beyond the scope of this investigation. The coastal temperature in the upwelling zone was roughly equivalent in the two cases, indicating that the heating from the surface does not have significant impact on the SST before it leaves the region. The size of the cold tongue, however, is critically dependent on this feedback. If the surface water moves away from the upwelling region without thermal interaction with the surrounding water, it will warm due to the air–sea flux. The perturbation temperature will have a time scale of $C_p/K_r$, which gives a length scale for the cold water tongue of $V C_p/K_r$, where $V$ is a representative advection velocity. If the velocity is determined by the wind stress, independent of $K_r$, it is clear that the stronger air–sea coupling will give a much smaller tongue of cold water.

In both cases, the Kelvin wave, excited by the westerly wind perturbation, propagates along the equator and down the coast and changes the SST in the coastal zone by roughly the same amount. The Kelvin wave acts to change the temperature of the upwelled water and hence changes the SST. Since the coupling has had little time to influence the SST, the effects in the two runs are much the same. Two effects tend to make the offshore region of anomalous warming much smaller in the case with infinite atmospheric heat capacity. The initial cold tongue is very small, and so any advective change in SST can only be seen over a small region. In addition, if the warming signal produced by the coastal Kelvin wave is to propagate out into the basin, the same rapid decay process happens to the SST perturbation, and a very small region of warming is expected. The run with $K_r = 2.5 \text{ W m}^{-2} \text{ K}^{-1}$ produced a much larger warming region, both because the initial state has a larger zone of cold water, and because the SST perturbation is longer lived. When the model was modified to allow a large tongue of cold water, but with strong coupling to the atmosphere, the two effects separated, and it appears that the influence of the relaxation time dominates the model’s response.

Models have been run with the observed annual cycle of atmospheric air temperature used with (1). The observed surface air temperature has large horizontal gradients, and can drive ocean models to have better looking mean SST distributions. However, the physics implicit in (1) means that the ocean can have no influence on the atmosphere. If this is so, then it becomes difficult to explain why the air temperature gradients exist in the first place. However, the radiative–convective approach used here sheds light on this problem. In (9), we have assumed that the radiative properties are independent of longitude, yet this simple box ocean model produces horizontal SST gradients which duplicate the climatological Pacific SST. Eq. (10) indicates that the surface air temperature will then be in a balance between the constant radiative temperature and the variable SST, and this model produces its own surface air temperature distribution which shares many features with climatology.

SST anomalies are strongly dependent upon the initial conditions in which they develop, and also upon the way in which the atmosphere responds to them. If perturbations in the ocean conditions are felt by the surface layers of the atmosphere, the duration and strength of the anomalies is extended. Models which produce highly realistic mean states through use of pre-specified air temperatures severely truncate the physics associated with this feedback. The dynamically induced changes in the SST will be damped back to the climatological air temperature with the rapid decay time $C_p/K$ (about 50 days). The results of Case 3 indicate that this process eliminates much of the strong warming caused by the Kelvin waves. It should be expected that a run with the strong coupling to the annual air temperature should fail to produce significant regions of anomalous SST. If the atmosphere does not respond to the SST anomalies [i.e., Eq. (1) is correct] then these results provide a very discouraging assessment of our ability to model El Niño—the ocean dynamics associated with the Kelvin and Rossby waves can not give rise to enough of the warming to be the cause, and the source must be entirely within the atmosphere itself.

The results in Case 2 lead to the view that the processes that determine the cold water tongue observed in the Pacific SST may be due to oceanographic features describable with the wind-driven model used here—i.e., the SST is not overly dependent on subtle atmospheric motions, cloudiness, and radiative properties that are beyond the abilities of the present model, but rather the strength of the upwelling is of vital importance. This leads to the examination of the role of changes in the longshore component of the wind stress. Small changes in the southerly wind gave substantial changes in the SST and in the entire...
tongue of cold water. The stress was reduced by \( \frac{1}{4} \) dyn cm\(^{-2} \) or 25\%—equivalent to a 12\% reduction in the surface wind speed. This stress variation produced changes in the SST which were largely indistinguishable from the changes due to the remotely forced Kelvin waves with the perturbation stress of 0.5 dyn cm\(^{-2} \). Both the shape of the warm anomaly and its temporal evolution are similar, due to the same process for establishing the offshore region of warming—propagation of the warming signal from the upwelling region via the surface currents. This propagation happens with the speed of the gravest-mode Rossby waves, a relatively slow process. While the remote westerly and local northerly perturbations should appear at the coast at different times, the Kelvin wave which brings the signal from the west is relatively fast, and this offset is small in comparison with the time that it takes to set up the offshore warming.

When the southerly wind stress is not changed right at the coast, but decreased in strength some 200 km offshore, the changes in SST are distinguishable only within the coastal zone. What appears to make a difference is the integrated change in the upwelling mass flux. In the modified case there is an upwelling at the coast with a weaker downwelling between the coast and 200 km offshore. Even though the water right at the coast remains at its cold temperature, the water leaving the 200 km zone has warmed in a fashion similar to that found when the stress relaxed all the way to the coast.

The implications of these results for analysis of El Niño are clear. It is not just the zonal wind component along the equator which matters for the SST response, but the entire wind field—at least within the region covered by the gravest-mode Kelvin and Rossby waves. The longshore wind south of the equator is particularly effective in this model result, even when the coastal wind remains steady. In the data of Rasmussen and Carpenter (1982), this is the region of poorest data coverage, and it is difficult to ascertain whether such changes do occur during El Niño events. If southerly stress changes of this magnitude do occur over sustained periods, the model suggests that significant warming should also occur. Whether the stress does change is a question beyond the scope of this investigation, but in light of these experiments, a question of significance for understanding El Niño. The same caveat applies to the remotely forced Kelvin wave fronts, although the better data density gives greater confidence in the assertion that such westerly stress perturbations do occur in concert with some El Niño events. The relationship and role of these features can most profitably be studied through a closer examination of the forcing fields to determine whether the necessary stress variations do occur.

The northerly wind perturbations produce much smaller surface height changes than the westerly wind case, even though they produce roughly equivalent SST changes. This is particularly true along the equator, where there is almost no change in the surface height when the southerly stress component is relaxed. The procedure of relating the model results to the changes in sea level at the Galapagos Islands, as done in Busalacchi and O'Brien (1981), will be a measure of the correlation of these changes with the westerly-induced Kelvin waves. It adds little information on the role of changes in the southerly wind component. Investigation of this difference in greater detail may provide a means of differentiating the types of ocean response which are occurring—does every El Niño show a sea-level signal along the equator?

The changes due to the alteration in the treatment of the atmosphere are not confined to the process of El Niño, but are seen to act on the evolution of the flow in the easterly mean wind case as well, changing the frontogenesis and stability of the oceanic flow in the model result. Whereas Schopf and Cane (1983) found strong baroclinic instability and sharp thermal and mixed layer depth fronts with a coupling constant of 30 W m\(^{-2} \) K\(^{-1} \), the weaker coupling led to a decrease in the frontogenesis, and a strong suppression of the baroclinic instability. Fronts still exist, and are formed by the same process as previously found, but the weaker coupling reduced one of the generation factors—the shallowing of the mixed layer due to strong heating of the upwelled water.

Future efforts will investigate the role of atmospheric transports in establishing the surface air temperature. In the tropical Pacific, the air moves to the west or northwest, and will carry heat out across the ocean. This will mean that the atmosphere will not be in balance with the local SST and radiative temperature through (10), but will instead be balanced between the radiative temperature and a weighted average of the upstream SST. Unfortunately, it is not easy to determine what the upstream path will be, since such substantial changes in the atmospheric thermal properties will, no doubt, significantly change the flow. Such changes would also occur in a coupled ocean–atmosphere model if the ocean were to generate SST changes as produced here. This of course means that the present treatment of the wind stress as an externally prescribed forcing is a simplification which makes this study an investigation of ocean mechanics in an abstract sense. El Niño derives from a more complex, ongoing interaction between the ocean and atmosphere which functions effectively in the tropics.

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