Seasonal Cycle Forcing of El Niño–Southern Oscillation in a Global, Coupled Ocean–Atmosphere GCM

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ABSTRACT

Recent advances in computer power and climate modeling capability have provided the opportunity for several modeling groups to undertake extended integrations with global coupled ocean–atmosphere climate models that allow the study of coupled processes thought to be important in producing Southern Oscillation and El Niño phenomena. Results are shown here from such a coupled model, developed at the National Center for Atmospheric Research (NCAR), that consists of a global, spectral (R15) atmospheric general circulation model (GCM) coupled to a global, 5° latitude–longitude, four-layer, ocean GCM. In spite of limitations of the coarse model grid, Southern Oscillation-type interannual variability of the ocean–atmosphere system is inherent in the coupled model. One of the mysteries of the Southern Oscillation cycle is how the system makes the transition from cold to warm phase and back again in the tropical Pacific in northern spring. Evidence is shown from the NCAR coupled model that a modulation of the mean seasonal cycle in the eastern Pacific drives the initiation and decay of warm and cold episodes in that region. The mechanism of this forcing in the model involves coupled seasonal anomalies of sea-level pressure (SLP), sea-surface temperature, surface wind stress, ocean upwelling, and convection–precipitation. These coupled anomalies form as a consequence of land–sea contrast in the eastern Pacific, in association with the evolution of seasonally low SLP over South America during northern winter and its movement with the seasonal cycle of solar forcing northwestward during northern spring. The anomalies become established farther west in the tropical Pacific as the year progresses and are associated with global patterns that resemble, in some ways, the phenomena observed with warm and cold events—the extremes of the Southern Oscillation. Similar sets of coupled processes exist in the observed long-term mean seasonal cycle, and the interannual events in the eastern tropical Pacific are manifested as a modulation of the mean seasonal cycle in the observations analogous to the coupled model. A reduction of coupling strength in the model (by reducing the strength of the wind-stress forcing from the atmosphere) eliminates both seasonal dependence and interannual anomaly signals. Turning off the seasonal cycle of solar forcing in the model changes the nature and regular evolution of the warm and cold events. Since the model fails to simulate any of the observed phenomena in the western Pacific, it is likely that only one of several possible sets of mechanisms involved with the observed El Niño–Southern Oscillation is simulated in the present global coupled model.

1. Introduction

Intense scrutiny of the ocean–atmosphere system in the tropical Pacific has been motivated by the probability of direct connections to anomalous weather over many parts of the globe (e.g., van Loon and Madden 1981), including most recently the North American drought of 1988 (Trenberth et al. 1988). Numerous observational and modeling studies (with both linear and nonlinear dynamical models) have attempted to explain interannual variability in these regions (e.g., McWilliams and Gent 1978; Busalacchi and O’Brien 1981; Blackmon et al. 1983; Philander et al. 1984; Barnett 1985; Meehl 1987; van Loon and Shea 1987; Zebiak and Cane 1988; Graham and White 1988; Lau and Shen 1988; Schopf and Suarez 1988; Battisti 1988). Previous investigators (Walker and Bliss 1930; Bjerknes 1969; Wright 1985; Trenberth and Shea 1987; Yasunari 1987) have identified the Southern Oscillation (SO), the large-scale interannual exchange of atmospheric mass between the tropical Indian and Pacific sectors, as associated with near-global atmospheric circulation patterns. Close linkages between the ocean and the atmosphere characterize SO phenomena, as manifested by warm sea-surface temperature (SST) anomaly events [El Niño or El Niño–Southern Oscillation (ENSO) events] and cold events (sometimes called La Niña) in the tropical Pacific (van Loon and Shea 1985). It has long been felt that an understanding of the mechanisms and forcings of the large-scale coupled ocean–atmosphere system in the tropics could lead to better seasonal forecasting capability in the midlatitudes.

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Because of the dynamically interactive nature of the tropical Pacific Ocean–atmosphere system and the near-global patterns of the associated teleconnections, one of the best tools to study these phenomena is a global coupled ocean–atmosphere general circulation model (GCM). Such a model not only can simulate the atmospheric circulation, but it can also interact with an ocean model to produce ocean temperatures and currents in much the same way as the real atmosphere–ocean system. In the last several years, more efficient models and faster computers have made possible extended integrations with global coupled ocean–atmosphere GCMs (e.g., Gates et al. 1985; Schlesinger et al. 1985; Bryan et al. 1988; Manabe and Stouffer 1988; Washington and Meehl 1989).

Many of these global coupled GCM runs have been designed to study phenomena associated with climate sensitivity, in particular, climate change associated with increases of carbon dioxide. However, interannual variability of the coupled ocean–atmosphere system resembling the Southern Oscillation was present in at least two simulations (Sperber et al. 1987; Philander et al. 1989). The reasons for its presence were not clear.

In an observational study, Meehl (1987) showed that Southern Oscillation-type signals exist in the Indian and Pacific sectors as a consequence of a modulation of the mean seasonal cycle in those regions, that the system has an inherent biennial tendency, switching from cold-to-warm phase and back again (relative to the previous and following years) in the tropical Pacific in many years, and that extremes in the oscillation are manifested by warm and cold events. However, the actual mechanisms of the transition in northern spring in the tropical eastern Pacific were not identified in that study. The purpose of this paper is to study those transition mechanisms in detail through the use of a global, coupled ocean–atmosphere GCM compared to observed data.

In spite of the inherent limitations and flaws of a global, coupled ocean–atmosphere GCM, interannual variability resembling the observed SO is simulated, the variability is manifested by a global pattern of atmospheric and oceanic anomalies, and the initiation and decay of warm and cold events in the tropical Pacific are tied to a modulation of the long-term mean seasonal cycle that triggers the events in the eastern Pacific. Evidence will be presented to show that elements of the coupled anomalies and their seasonal dependence in the eastern Pacific also appear in the observed long-term mean seasonal cycle and interannual warm and cold events and that the coupled model is simulating one subset of several different mechanisms operating in the observed system.

Section 2 describes briefly the coupled climate model developed at the National Center for Atmospheric Research (NCAR). Section 3 details aspects of interannual variability in the NCAR coupled model. A description of the transition from a composite cold to a composite warm event in the tropical Pacific appears in section 4, and section 5 shows the linkages between the interannual events and the long-term mean seasonal cycle. Section 6 contains a summary and conclusions.

2. The coupled ocean–atmosphere climate model

The NCAR global coupled ocean–atmosphere GCM is illustrated schematically in Fig. 1. The atmospheric model (CCM) has been used specifically for coupling to various ocean formulations (e.g., Washington and Meehl 1983, 1984, 1989). It is a global, spectral GCM and includes rhomboidal 15 (R15) resolution (about 4.5° latitude by 7.5° longitude), nine layers in the vertical, computed clouds, and parameterized land-surface processes. Meehl and Washington (1988) have analyzed in detail the sensitivity of land-surface processes in this model.

The ocean model has a coarse resolution (5° latitude by 5° longitude, four layers in the vertical) and includes realistic geography, bottom topography consistent with resolution, and a simple thermodynamic sea-ice formulation. Meehl et al. (1982) have tested this ocean model with a variety of parameter changes.

For the present experiment, the coarse-grid ocean model was first run by itself for 50 model years with observed atmospheric forcing. It was initialized the first year with observed vertical cross sections of temperature and salinity. The atmospheric model was also run separately for 15 model years [preceded by a spinup period (Washington and Meehl 1984; Meehl and Washington 1985)]. Then both models were synchronously coupled and run for 16 model years. Following this shakedown period, the coupled model was run for 30 model years. Because it was not originally designed to study long-term time series, only the last ten years of the run were available for the present analysis.

Flux-correction or flux-adjustment methods have been used in some coupled model simulations to ensure that the ocean simulation does not drift too far from the observed state of the ocean (Latif et al. 1988; Manabe and Stouffer 1988). With such methods, the coupled model mean climate resembles the observed climate more closely. But since the system is constrained, questions arise about the sensitivity of the model to changes in forcing and whether interannual variability simulated by such a model is consistent with what the model would have produced on its own.

Because the NCAR coupled model does not contain flux correction, simulated SSTs in the tropics are lower than observed. This cool water from the active upwelling in the model extends across the equatorial Pacific Ocean to the Indian Ocean with warmer water to the north and south in the tropical western Pacific (Fig. 2). Meehl (1989) analyzed the source of this coupled
model error for the tropical Indian and Pacific regions and concluded that the inherent limitations of the ocean model in resolution and geography contribute most to these errors. For example, as the 50-m thick surface layer in the equatorial tropics responds to easterly winds from the atmosphere, Ekman divergence occurs and cool water from the 450-m thick second layer is brought to the surface. Washington and Meehl (1989) show that weaker easterly wind stress from the atmospheric model or a shallower ocean model second layer results in warmer tropical SSTs in the coupled model. In addition, the coarse horizontal ocean model grid configuration of continental outlines does not provide a barrier in the western Pacific. Consequently, the Pacific and Indian Oceans are directly linked along the equator, and thus there is no western Pacific warm pool along the equator in the model and little east-west temperature gradient in the mean across the equatorial Pacific (Fig. 2). The effects of this SST distribution on other aspects of the simulated climate (e.g., precipitation, winds, etc.) are documented by Washington and Meehl (1989) and Meehl (1989). A slight drift of about $-0.02^\circ\text{C} \text{yr}^{-1}$ is evident in the surface layer of the ocean. For the ten-year period under consideration here, this drift is small compared to the signals of SST variability with amplitude on the order of $1^\circ\text{C}$.

3. El Niño–Southern Oscillation in the coupled model

In spite of the inherent limitations and errors in the simulation mentioned above, the interannual variability is pronounced in the tropical Pacific in the NCAR coupled model. As a first step in analyzing these fluctuations, area averages of sea-level pressure (SLP) anomalies (seasonal mean SLP minus the ten-year seasonal means in the model) for an area in the eastern Pacific ("Niño3" 10°N to 10°S, 90° to 150°W) and an area over the Indonesian region (10°N to 13°S, 110° to 155°E) are plotted for years 21 to 30 in the coupled model (Fig. 3b). This can be compared to a similar plot of observed SLP anomalies for Tahiti and Darwin (Fig. 3a). The familiar fluctuation of atmospheric mass, characteristic of the SO (as indicated by the opposite sign of SLP anomalies between the eastern and western Pacific), is evident in both model and observations, with the amplitude of the oscillations in the model less than the observed.

A measure of variability associated with the SO is the Southern Oscillation index (SOI). The SOI calculated by the Climate Analysis Center (CAC) is shown in Fig. 4a. Extremes of the index are often associated with warm events (negative SOI) and cold events (positive SOI) in the Pacific. A similar SOI calculated from model grid points near Tahiti and Darwin appears in
Fig. 2. SSTs (°C) for the global tropics: (a) DJF from coupled model. (b) DJF, observed from Alexander and Mobley (1976).
(c) Same as (a) except for JJA. (d) Same as (b) except for JJA.
Fig. 3. (a) SLP anomalies (five-month running mean) at Tahiti (solid line) and Darwin (dashes) from CAC. (b) Seasonal SLP anomalies (three-season running mean) from an area in the tropical eastern Pacific ("Niño3," 10°N to 10°S, 90° to 150°W) and an area over Indonesia in the far western Pacific ("Indonesia," 10°N to 13°S, 110° to 155°E).

Fig. 4b. The magnitude of the model SOI is about half of the observed in Fig. 4a. However, this plot suggests that warm and cold SST events in the Pacific are associated with the positive and negative excursions of the model SOI in a similar fashion to the observed SOI.

Figure 5a shows a time–longitude plot of SST anomalies in the model from 50°E in the Indian Ocean to 80°W in the Pacific, averaged from 10°N to 10°S. For this ten-year period in the model, SST anomalies generally appear in the eastern Pacific and become established progressively farther west. Maximum amplitudes are roughly ±1°C. The magnitude of the SST anomalies in the coupled model is about 30% less than observed, composite warm-event SST anomalies shown, for example, by Rasmusson and Carpenter (1982). The establishment of SST anomalies greater than ±0.5°C between about 110°W and the date line (e.g., the warm event starting in MAM of year 26 and ending after MAM of year 27). Two-year warm and cold events also occur (e.g., a warm event lasting from MAM year 22 to MAM year 24 and a cold event lasting from MAM year 24 to MAM year 26) and are analogous to single-year and multiyear events observed in the tropical Pacific (van Loon 1984).

Warm-water events in the model eastern Pacific (Fig. 5a) correspond with low SLP in the Niño3 region, high SLP over Indonesia (Fig. 3), and a negative SOI (Fig. 4), and vice versa for cold events. This is similar to the association between Tahiti and Darwin SLP and warm and cold events observed in the Pacific. Like the observed ENSO phenomenon, the model SOI involves processes of the dynamically coupled ocean–atmosphere system.

Figures 5b–d show time–longitude plots similar to the SST plot in Fig. 5a for SLP, precipitation, and sur-
face wind stress from the model. There is a close association in the model in time and space among positive SST anomalies, negative SLP anomalies, positive precipitation anomalies, and westerly wind-stress anomalies to the west (i.e., wind-stress convergence in the areas of warmest SSTs), and vice versa for negative SST anomalies.

4. Composite warm and cold events in the coupled model

Two questions arise at this point. First, why do the anomalies move from east to west? Second, why do the transitions from cold-to-warm phase or warm-to-cold phase near 110°W and the date line in the tropical Pacific depend on the seasonal change from northern spring to northern summer? To address these questions, composites are formed for warm and cold events in the model. The year of initiation of a warm event (year zero) is defined as a year with SOI (calculated from grid points nearest Tahiti and Darwin in the model—Fig. 4) less than −0.5 and SST in the Niño3 area greater than 0.25°C. Year zero for a cold event in the model is defined by SOI greater than 0.5 and SST in Niño3 less than −0.25°C. Two warm-event (years 22 and 26) and two cold-event initiations (years 24 and 27) are defined from years 21–30 in the coupled model run. Examination of the seasonal composite anomalies
shows that, as in the observed events (Meehl 1987; Kiladis and van Loon 1988), the warm- and cold-event patterns are near-mirror images of each other. To emphasize the composite patterns, differences of warm-event seasons minus cold-event seasons are computed. The sign of the difference, a transition from a cold phase to a warm phase, will be discussed in that context. The designations of warmer, colder, higher, lower refer to relative anomalies in the model.

Negative SST differences characteristic of the waning stages of a cold event cover most of the central and western tropical Pacific during DJF0 (Fig. 6a). Small-amplitude warming is just starting in the far eastern Pacific. Positive SLP anomalies (Fig. 6b) overlie the cool water in the tropical central and western Pacific. In the far eastern Pacific, the small-amplitude positive SSTs are associated with negative SLP anomalies that extend across most of tropical South America. This anomalous SLP gradient is associated with ageostrophic westerly wind-stress anomalies directed from high to low pressure in Fig. 6c along the equator between about 100° and 150°W. Similar relationships exist between SST and low-level winds in the observed system as well (Lindzen and Nigam 1987), particularly in the eastern Pacific (Gutzler and Wood 1989). The westerly wind-stress anomalies in the model (weakened easterly trade winds) along the equator in that region result in a decrease in upwelling in the ocean (Fig. 6c). Most areas under the influence of westerly anomaly wind stress and suppressed upwelling lie just to the west of the warm SST anomaly.

In the model, strong trades along the equator cause Ekman divergence in the upper layer of the ocean model, and the cooler water from the second layer is then brought to the surface. This upwelling process, as noted previously, maintains low SSTs in the equatorial Pacific. When the easterly trade winds in the model slacken (evidenced by the westerly wind-stress anomalies in Fig. 5c), the upwelling decreases. This means that less cool water from the second layer in the model reaches the surface and SSTs are higher. Simplistically, this can also occur in the observed ocean–atmosphere system in the tropical Pacific. However, the relationship between upwelling and SST in the real system depends critically on the depth of the thermocline. Since the thicknesses of the upper two layers of the ocean model are 50 and 450 m, the very close association between upwelling and SST in the model is not analogous in all respects to the real ocean. This will be discussed further in relation to Fig. 12.

When warmer water appears in the equatorial eastern Pacific as a result of the weakened easterlies and reduced upwelling in the model, lower pressure overlies this warm water, partly due to a hydrostatic relationship between the warm surface and low SLP associated with thermally direct circulation similar to that noted in the observed system in this region (Lindzen and Nigam 1987). Associated increases in convection and precipitation also occur over the warmer water, as suggested by Fig. 5d and noted in the observations in the eastern Pacific (Gutzler and Wood 1989). The SLP gradient between the positive SLP anomalies in the central equatorial Pacific and the negative SLP anomalies in the eastern Pacific is established west of the positive SST anomalies, the westerly wind-stress anomalies set up by this surface pressure gradient extend farther west, upwelling is weakened, and SSTs become higher farther west during MAM0 (Fig. 7). Smaller-magnitude meridional component wind-stress anomalies are present in Figs. 6–8 just to the north and south of the warm-water anomalies. Though less extensive, these meridional anomalies also contribute to convergence in the warm-water regions as noted by observational studies (e.g., Philander 1985).

This set of coupled processes involving the atmosphere and the ocean continues to establish the positive SSTs farther west in the model until, by JJA0, the positive SSTs almost reach the date line and positive SLP extends all the way to India (Fig. 8). There, the positive SLP is associated with a suppression of the Indian Monsoon, as evidenced by the weakened southwestward flow (easterly wind-stress anomalies) in the Indian Sea. This is analogous to the observed relationship between warm events in the Pacific and weak Indian Monsoons (Rasmusson and Carpenter 1983).

The mechanism in the model, then, for the movement of the coupled anomalies from east to west is the establishment of an SLP gradient (consistent with the SST anomaly pattern) that is associated with westerly anomaly wind stresses and suppressed ocean upwelling to the west of the positive SST anomalies. Decreased upwelling means that warmer SSTs form to the west in the model and the entire set of coupled anomalies continues to the west.

Associated with these patterns of SST, SLP, and wind stress are corresponding changes in convection, precipitation, and atmospheric circulation in the troposphere (not shown). As mentioned before, high SSTs and low SLP are generally associated with increased convection and precipitation in the model tropics (Fig.

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**Fig. 5.** Time–longitude plot of seasonal mean differences from the long-term seasonal means for years 21–30 in the coupled model, 50°E to 80°W, averaged from 10°N to 10°S. Vertical line near center is date line; horizontal lines demarcate seasonal boundary between northern spring (MAM) and northern summer (JJA): (a) SST. Stippling greater than +0.5°C; hatching less than −0.5°C. (b) SLP. Stippling less than −0.5 mb, hatching greater than +0.5 mb. (c) u-component wind stress. Stippling greater than +0.1 dyn cm⁻² (+0.01 N m⁻²), hatching less than −0.1 dyn cm⁻² (−0.01 N m⁻²). (d) Precipitation. Stippling greater than +0.5 mm day⁻¹, hatching less than −0.5 mm day⁻¹.
Fig. 6. For DJF<sub>0</sub> at the end of a composite cold event beginning the transition to a warm event. (a) Composite SST anomaly differences for warm-event composites minus cold-event composites in the coupled model. Solid lines are positive differences (stippled areas), dashed lines are negative differences. (b) Same as (a) except for SLP differences; negative differences are stippled. (c) Same as (a) except for wind-stress vector differences. Stippling indicates decreased upwelling in the Pacific between 10°N and 10°S. Scaling arrow at bottom right is 0.3 dyn cm<sup>-2</sup> (0.03 N m<sup>-2</sup>).

5). This affects vertical motion and latent-heat release in the atmospheric model. As a result, the changes in SST and SLP at the surface are indicative of large-scale fluctuations of thermally direct atmospheric circulation throughout the tropical troposphere, such that low SLP and increased convection in the tropical Pacific are associated with increased vertical motion, large-scale sinking, and high SLP to the east and west.

The SO-type interannual variability in the tropics involves near-global changes in atmospheric circulation and SST patterns in the coupled model as well. Figure 9 shows composite global patterns of (a) SST and (b) SLP anomalies corresponding to the waning stages of a cold phase in the tropical Pacific (DJF<sub>0</sub>) as a warm phase is about to begin. (Figure 6 shows the tropical anomalies for this season.) The analogous anomalies from the observed system appear in Fig. 9c (SST anomalies from December 1988 indicative of general SST anomaly patterns for that time of year during a cold event) and Fig. 9d (composite SLP anomalies for DJF<sub>0</sub> before the inception of observed warm events when low SSTs dominate the central equatorial Pacific) after van Loon (1986). For the SST anomaly patterns, both computed and observed patterns show low SSTs
in the subtropical northwest and southwest Pacific, low SSTs south of the Aleutians, and a band of low SSTs near 50°–55°S at most longitudes. The corresponding SLP anomaly patterns for computed and observed show high SLP over the areas of cool water in the tropical Pacific just east of the date line, low SLP to the north and south in the subtropical Pacific, high SLP near the Gulf of Alaska, low SLP over southern North America, and a band of high SLP near 45°S with low SLP to the south. This latter feature implies a strengthening of the north–south pressure gradient in the southern midlatitudes, an increase of the mean westerly low-level winds (not shown), and lower SSTs at 50°–55°S in both model and observations. There is less agreement between model and observations over northwestern North America, the North Atlantic, Europe, and western Asia.

For the season at the peak of a warm event (DJF+1), both model and observations (Fig. 10) show high SSTs in the equatorial Pacific, low SSTs in the subtropics to the north and southwest, and a band of warm SSTs at most longitudes south of about 45°S. Model and observed composite SLP anomaly patterns for this season indicate low SLP in the tropical Pacific east of the date line, high SLP to the northwest and southwest, a center of low SLP in the North Pacific, low SLP in the Indian and Pacific oceans near 40°S (model, Fig. 10b) and 45°S (observations, Fig. 10d) with positive SLP anom-
Fig. 8. As in Fig. 6 except for JJAs.

alias around 55°S (model, Fig. 10b) and 60°S (observations, Fig. 10d). This indicates a weakening of the SLP gradient, reduced westerlies, and warmer water near 50°S in both model and observations. Clearly, the SST anomaly patterns in the tropical Pacific are associated with global atmospheric and oceanic anomalies in both model and observations. The evolution of these global patterns in the model is under further scrutiny.

Returning to the seasonal evolution of composite anomaly patterns in the tropics, Figs. 11a and 11b show warm-minus-cold-event, composite, time-longitude plots for SST anomalies and u-component, wind-stress anomalies. As before, the sign of the anomalies is a warm event. Both are averaged from 10°N to 10°S.

Cool water just to the east of the date line early in the cycle (JJAs through MAMs) gives way to the large-amplitude continuous positive anomalies (stippled areas) in the eastern Pacific in MAMs (Fig. 11a). These positive SST anomalies with narrow longitudinal extent move west as warm water is established farther west. This band of warm water is associated with the zero line separating westerly and easterly wind-stress anomalies in Fig. 11b (i.e., u-wind-stress convergences). The establishment of easterly anomaly wind stresses in the far eastern Pacific in JJAs (Fig. 11b) is associated with a reduction in amplitude of the positive SST anomalies there. By SONs, the sign of the SST anomalies is negative around 90°W in association with the strong easterly wind stresses near that longitude.
Comparative time–longitude plots of composite observed warm events (1957, 1965, 1972 events from Rasmussen et al. 1986) for SST and surface $u$-wind anomalies are shown in Fig. 11c and 11d (composite SST and wind anomalies for six warm events—1951, 1953, 1957, 1965, 1969, 1972—are given by Rasmussen and Carpenter, 1982, Fig. 22, and show a similar pattern). The seasonal timing between model and observed is comparable in the eastern Pacific with the appearance of large-amplitude positive SST anomalies in northern spring of year zero. These positive anomalies then become established farther west to the date line by the following northern winter (JAN+1). The pattern of low-level $u$-wind (observed) and $u$-windstress (model) anomalies is also similar in the eastern Pacific with the zero line dividing westerly anomaly and easterly anomalies (indicative of $u$-wind-stress convergence) coinciding with the establishment in the eastern Pacific (and movement westward) of large-amplitude SST anomalies beginning in northern spring of year zero. (For comparison, largest $u$-component, westerly surface wind anomalies from the coupled model are about 2.0 m s$^{-1}$.)

Two interesting differences between the composite model and observed warm events emerge, however. First, observed positive SSTs and westerly wind anomalies in the western Pacific at the beginning of year zero migrate eastward (Fig. 11c, d) and meet the anomaly signals from the east in the central Pacific around northern summer of year zero. The coupled model shows no comparable features in the western Pacific. Second, the warmest water in the composite events near 100°–110°W in the eastern Pacific in the observations (Fig. 11c) occurs in northern fall of year zero at a time when easterly anomaly winds are established there (Fig. 11d). In the model, easterly anomaly wind stress appears with similar seasonal timing in the far eastern Pacific as noted earlier, but the ocean surface layer immediately cools under the influence of these easterly anomaly wind stresses in the model. This confines the positive SSTs in the model warm events to a narrow longitudinal band (Fig. 11a) compared to the wide extent (80°W to the date line) of the observed positive anomalies (Fig. 11c).

Several factors explain the lack of simulation of processes in the western Pacific in the coupled model. Primarily, the coupled model does not simulate the western Pacific warm pool or any of the processes thought important to maintain it there (Fig. 2). The model’s failure to maintain warm SST anomalies late in year zero of an event in the eastern Pacific is probably related partly to the lack of vertical resolution and inability to simulate properly some of the processes involving mixed-layer dynamics. To better illustrate this point, Fig. 12 is a schematic of coupled processes in the eastern Pacific in the present model and those thought to be occurring in the observed system (e.g., Wyrtki 1975). In the model, easterly wind stress in the equatorial eastern Pacific causes Ekman divergence in the surface layer, and the cool water of the second layer is brought to the surface (Fig. 12a). This cool water is associated with high SLP and weak convection, precipitation, and upper-level divergence. Westerly anomaly wind stress causes weakened upwelling and warming of the surface layer (Fig. 12b). This is associated with low SLP and active convection, precipitation, and upper-level divergence. With the reestablishment of easterly wind stress (Fig. 12c), upwelling of the cool second layer water resumes, SST is lower, and the associated SLP is again higher with weakened convection, precipitation, and upper-level divergence as in Fig. 12a.

For the observed system, strong easterly wind stress causes vigorous upwelling of cool water from below the thermocline (Fig. 12d). SSTS are low, SLP is high, with weak convection, precipitation, and upper-level divergence. As anomalous westerly wind stresses appear, upwelling is reduced, the surface layer warms and the thermocline deepens (Fig. 12e). During this period, the arrival of downwelling Kelvin waves generated by westerly anomaly winds in the western Pacific several months earlier could contribute to deepening the thermocline. Finally, when the easterly wind stress is re-established in October, upwelling resumes, but its vertical extent is confined to the warm layer above the deepened thermocline. Therefore, warm water is brought to the surface and the SSTS continue to rise.

The coarse vertical resolution of the ocean model’s upper ocean makes the SSTS very sensitive to upwelling and the associated wind-stress forcing from the atmosphere. A weakening of the wind stress from the atmospheric model to 0.25 of its original value results in a reduction in equatorial upwelling by an order of magnitude and an increase in tropical SSTS by about 4°–5°C (Washington and Meehl 1989). However, the timing of the phenomena in the eastern Pacific and the westward movement of the anomalies there point to some interesting correspondence between the model and observed. Both appear to be linked to the seasonal cycle in the eastern Pacific.

It is likely that the coarse resolution of the ocean model either distorts or does not adequately resolve internal ocean waves. The westward mode 1 Rossby wave speed in the ocean model [deduced from mean vertical temperature profiles in the equatorial Pacific; refer to Gent et al. 1983, their Eq. (8)] is about 1.0 m s$^{-1}$. This is close to the theoretical phase speed of about 0.9 m s$^{-1}$, but is only true for sufficient latitudinal and longitudinal resolution. The mean westward phase speed of SST anomalies in the coupled model (warm and cold anomalies deduced from Fig. 5) is 0.20 m s$^{-1}$, much slower than the calculated mode 1 westward Rossby wave phase speed, but about twice as fast as the mean westward surface current speed in the model of about 0.10 m s$^{-1}$. As mentioned earlier, no well-defined western boundary in the equatorial western Pacific exists in the ocean model because of the coarse
FIG. 9. (a) Global SST differences for the warm-minus-cold-event composites in the coupled model for DJF₀, as discussed in text. Sign of differences is representative of a cold episode in the central equatorial Pacific prior to the beginning of a warm episode. Stippling indicates cooler water. (b) Same as (a) except for SLP. Stippling indicates higher pressure. (c) Observed global “blend” SST anomalies from the CAC for December 1988. Stippling indicates cooler water. This period is representative of a cold episode in the central equatorial Pacific. (d) Mean SLP anomalies for the northern winter season prior to the inception of composite warm events. Stippling indicates higher pressure. Period is representative of a cold phase in the central equatorial Pacific. Event composites are differenced from mean SLP that does not contain those event years (van Loon 1986).
model grid and simplified continental outlines. Wave reflection and returning eastward Kelvin waves, therefore, are not factors in the model. Westward advection of the SST anomalies alone evidently does not control the westward propagation. However, Hirst (1986) shows in his simple mechanistic model that westward
Fig. 10. As in Fig. 9 except for computed and observed SST and SLP anomalies representative of a warm event in the central equatorial Pacific.
c) OBSERVED SST ANOMALIES, DECEMBER, 1987 (WARM EVENT)

d) OBSERVED MEAN DEVIATION OF SLP (WARM EVENT)

FIG. 10. (Continued)
FIG. 11. Time–longitude plots showing the evolution of composite warm events: (a) Warm-minus-cold-event composite SST differences (°C) from the coupled model (10°N–10°S). Stippled areas greater than +0.75°C, hatching less than −0.75°C. (b) Warm-minus-cold-event composite u-component wind-stress differences (10°N–10°S). Stippling greater than +0.5 dyn cm⁻² (+0.05 N m⁻²). (c) Observed composite SST anomalies along the equator for 1957, 1965, and 1972 events minus the long-term mean (after Rasmusson et al. 1986). Stippling indicates positive SST anomalies. (d) Same as (c) except for u-component surface wind anomalies. Stippling indicates positive (westerly) wind anomalies.
propagation occurs when temperature advection is included. Therefore, the present unstable coupled mode may also be influenced by mean advection in the ocean surface layer.

In any case, equatorial upwelling is a dominant process in the present coupled model. The coupled propagation mechanisms outlined in Figs. 7–9 that depend critically on modification of upwelling and have to do with the westward movement of coupled anomalies also play similar roles in the long-term mean seasonal cycle as well and are examined in the next section.

5. The role of the seasonal cycle

In attempting to explain observed interannual variability in the tropical Pacific, previous investigators have postulated sets of interacting processes similar to those described above (Bjerknes 1969). The intriguing aspect of the present results is that the coupled model system is producing them internally, entirely as a consequence of the coupled interactions of the two media. This type of seasonally dependent interannual variability seems to be inherent in the model.

To investigate the role of the seasonal cycle, time-longitude plots averaged from 10°N to 10°S for long-term seasonal means from the model are shown in Fig. 13 for SLP, u-component wind stress, and SST. The heavy dashed line running diagonally from east to west traces a relative maximum of SST in (c) which appears in the far eastern Pacific in the model during December-January-February (DJF) and MAM and moves out toward the date line as the seasonal cycle evolves. The SLP plot (a) shows higher pressure just to the west of the highest SST. The pressure gradient set up by the
higher pressure to the west and lower pressure to the east produces an area of weaker easterly wind stresses just to the west of the highest SST (b). The result is wind-stress convergence in the region of highest SST.

Similar evolution of mean seasonal SST and wind-stress convergence in the equatorial eastern Pacific has been documented in the observed system (Horel 1982). Figure 14a shows the annual cycle of observed SLP (10°N–10°S) from atlas data, and Fig. 14b shows a similar time-longitude plot for the long-term mean seasonal cycle of SST along the equator (Horel 1982). The annual mean has been removed from both. The heavy dashed line traces a maximum of SST that appears in the far eastern Pacific in northern spring in Fig. 14b and becomes established farther west as the seasonal cycle evolves. This same line is traced on Fig. 14a to show that, as in the coupled model, the mean seasonal cycle is characterized by higher SLP to the west (and lower SLP to the east) associated with wind-stress convergence that follows the SST maximum in its annual westward progression (Horel 1982). Also similar to the coupled model is the fact that modulations of the observed mean annual cycle in the eastern Pacific produce or trigger the observed transition of warm and cold events that have a comparable seasonal timing and anomaly signature. For example, the mean westward phase speed of the SST maximum taken from Fig. 14b is about 60 cm s⁻¹, while the westward phase speed of the +1.0 SST anomaly contour (Fig. 11c) is also about 60 cm s⁻¹. In this area, the mean observed westward surface current is about 30 cm s⁻¹ during northern spring (Meehl 1982), suggesting that advection is probably not the controlling mechanism, but, when taken together with the zonal temperature gradients in that region, may play a role in the mean westward propagation (e.g., Hirst 1986). In the model (Fig. 5), the westward propagation speed of the anomalies is about 20 cm s⁻¹, while mean westward surface currents are about 10 m s⁻¹. Meanwhile, the mean westward phase speed of the SST maximum is about 20 cm s⁻¹ (Fig. 13c). This suggests that similar sets of coupled processes in both model and observations, as well as the interannual warm and cold events, are involved with the long-term mean seasonal progression. The westward movement of anomalies in the model is slower than the observed for both the interannual events and mean seasonal cycle, and may be related to the coarse horizontal and vertical resolution in the ocean model.

Maps of mean SLP (Fig. 15) can also illustrate the role of the seasonal cycle. In the model and the observed

**Fig. 13.** Time-longitude plots of long-term seasonal means from the coupled model: (a) SLP in millibars. Stippling indicates relatively higher SLP. (b) u-component wind stress in dyn cm⁻² (0.1 N m⁻²). Stippling indicates weaker easterly wind stress. (c) SST in °C. Stippling indicates higher SST. Heavy dashed line traces seasonal evolution of axis of higher SSTs in eastern Pacific in (c).
long-term mean SLP fields, seasonally low pressure over South America is evident during northern winter (Figs. 15a, d) when the sun is most directly overhead in the southern tropics. In northern spring (Figs. 15b, e), this low pressure and the associated convection and precipitation then weaken and move northwestward (also noted by Horel et al. 1989) until, by northern summer (Fig. 15c, f), low pressure is established well to the north over southern North America. During this seasonal transition of low pressure and associated convection and precipitation from South America to southern North America, low pressure is established in the far eastern tropical Pacific in northern spring; SSTs are highest along the equator at that time of year; and a weak east–west SLP gradient is set up. This results in a weakening of the equatorial easterlies with associated wind-stress convergence (Horel 1982), implied suppression of the ocean upwelling and weakened westward (or slight eastward) surface currents (Halpern 1987), and higher SSTs in the far eastern Pacific (Horel 1982). Once initiated by the passage of the seasonal cycle, the whole set of coupled anomalies moves west in the observed long-term mean (Fig. 14) similar to the coupled model (Fig. 13), with the axis of highest SST just to the west of lowest SLP, as described earlier. The evolution of the seasonal cycle of SLP and the conditions set up in the ocean by the state of SSTs in the tropical Pacific at that time of year produce warm and cold events through the same set of coupled anomalies, depending on the strength of the surface pressure gradient due to the contrast of land and sea in the far eastern Pacific in northern spring. That is, if SSTs in the tropical eastern Pacific are cool as northern spring approaches, higher SLP will occur there compared to lower SLP to the east over South America, and vice versa for SSTs that are warm as northern spring approaches.

 Processes in the Indian Ocean and western Pacific involving the annual cycle and transitions from cold to warm events (including the eastward movement of SLP, precipitation, and wind anomalies coincident with the annual cycle there) (Meehl 1987; Gutzler and Harrison 1987; van Loon and Shea 1987) are not well simulated in the coupled model. This is probably related to the lack of a western Pacific warm pool and absence of the South Pacific convergence zone (Washington and Meehl 1989). In effect, the eastern Pacific regime in the model extends to the Indian Ocean. Yet the model is able to make transitions from cold-to-warm phase and back again in the eastern Pacific with seasonal timing dependent on modulations of the annual cycle similar to the observations. The fact that the coupled model has no western Pacific warm pool or SST anomalies that form there and move eastward, but is still able to produce SST anomalies that form in the eastern Pacific and move west, could shed light on the different character of the onset of observed El Niños, some having stronger SST anomalies that originate in the east or in the west (Rasmusson and Wallace 1983; Fu et al. 1986). For example, the 1982–83 and 1986–87 warm events show larger SST anomaly signals originating in the western Pacific and moving east (Kousky and Leetmaa 1989), compared to the composite warm events (Fig. 11c) by Rasmusson and Carpenter (1982) that show larger SST signals originating in the eastern Pacific and moving west.

 To investigate the sensitivity of these coupled processes to alterations of forcing, the coupled model was first run for five years with the wind-stress forcing from the atmospheric model reduced to 0.25 of the original values. A time–longitude plot of the seasonal SST anomalies produced in that experiment (Fig. 16a) can
Fig. 15. Long-term mean SLP from the coupled model: (a) DJF, (b) MAM, (c) JJA. Stippling indicates areas less than 1012.5 mb. Dashed line in the eastern Pacific is the 1013.5-mb contour. Long-term mean observed SLP from atlas data (Taljaard et al. 1969; Crutcher and Meserve 1970) for: (d) December. (e) March. (f) June. Stippling indicates areas less than 1010 mb. Contour interval is 2.5 mb.
be compared to a similar plot of SST anomalies from the original model experiment in Fig. 5a. Virtually no interannual SST variability and no longitudinal movement of anomalies exist in the reduced wind-stress experiment. Plots of SLP and SST for the long-term mean (five-year) seasonal cycle in Fig. 17 show little longitudinal variation with time of either SLP or SST. Results from this experiment suggest that the concept of
with the solar insolation left at its mid-March value during year 25 of the coupled model run. The subsequent evolution of SST anomalies is shown in Fig. 16b where the six-year averages of 90-day means with the perpetual March solar forcing are subtracted from individual 90-day means during the six-year period. The low SST anomalies in the earlier years compared to the later years are indicative of a warming of the equatorial tropics with time as an adjustment to the sun being perpetually over the equator with the cessation of the seasonal cycle. The pattern of the SST anomalies, compared to the post-year 25 plot in Fig. 5a, indicates an alteration of the development and behavior of tropical interannual SST anomalies. The cool anomalies present in year 25 of the original experiment east of the date line continue to evolve and move slowly to the west (Fig. 16b). However, positive SST anomalies in the east are not established until midway through year 2 of the perpetual March case, then linger near 90°W for a year, and finally move slowly west during year 4. In the original run, the one-year warm event that began early in year 26 and ended one year later is not represented in the perpetual March run. In years 5–6 of the perpetual March run (Fig. 16b), there is no evidence of persistent SST anomalies of either sign that move in any direction.

This experiment indicates that, without the seasonal cycle of forcing, anomalies in the coupled model still occasionally form and tend to move from east to west. This is similar to what was noted by Philander et al. (1989) in their coarse-grid coupled model with annual mean solar insolation. They attributed such signals to forcing from “random” atmospheric disturbances. But without the seasonal cycle in the eastern Pacific as a trigger, the regular formation and progression of anomalies seen in the present coupled model with the seasonal cycle are not reproduced.

**6. Summary and conclusions**

Results from the coarse-grid global coupled model (global, spectral R15, atmospheric GCM coupled to a global, 5° latitude-longitude, four-layer ocean GCM) suggest that the model simulates one aspect of the observed system having to do with interannual variability in the tropical eastern Pacific manifest by a modulation of the long-term mean seasonal cycle there. The coupled processes identified in the model appear to have analogues in the observed system that provide one possible set of mechanisms to explain the transition and seasonal dependence of the Southern Oscillation-type signals involving the coupled system in the eastern Pacific, as outlined by Meehl (1987). Figures 18a–c summarize the processes taking place in the coupled model transition from cold-to-warm SSTs in the tropical Pacific. If the system starts with relatively cold water in the tropical Pacific in DJF (Fig. 18a), active convection and low SLP to the east will be associated with
Fig. 18. Schematic of the transition from a cold event to a warm event in the eastern tropical Pacific. Transition in the coupled model is shown for: (a) DJF. (b) MAM. (c) JJA. Likely sets of interactions for the observed system are shown for: (d) DJF. (e) MAM. (f) JJA. Cold/high areas refer to low SSTs and high SLP. Warm/low denotes high SSTs and low SLP. Large arrow at surface indicates direction of surface wind-stress anomalies. Ribbon arrows above the surface denote relative large-scale rising and sinking associated with thermally direct convection/precipitation enhancement or suppression. Dashed ribbon arrows show weak large-scale rising/sinking.
a west–east SLP gradient at the surface. This gradient drives ageostrophic, near-equatorial, surface westerly anomaly wind stresses that suppress ocean upwelling to the west. The reduced upwelling allows the ocean surface layer to warm and positive SST anomalies to appear to the west. By MAM$_0$ (Fig. 18b), the warm water in the far eastern Pacific is associated with active convection, low SLP, upper-level divergence, and an east–west atmospheric thermally direct circulation indicated schematically by the ribbon arrows. Large-scale sinking and high SLP are maintained to the west over the cool water; the surface SLP gradient is associated with westerly surface-wind anomalies and suppressed upwelling in the ocean to the west of the warm water; SSTs warm to the west; and the whole system of coupled anomalies moves west. By JJA (Fig. 18c), warm water is established well to the west in the tropical Pacific, and the stage is set for the transition back to negative SST anomalies the following northern spring due to the inverse set of processes described for the cold-to-warm transition.

Even though the coupled model simulates, in a simplified way, some of the coupled processes and seasonal dependencies, the observed system appears to be much more complicated. Figures 18d–f show coupled processes that could participate in a cold-to-warm-water transition in the tropical Pacific. The eastern Pacific operates much the same as the coupled model scenario in Figs. 18a–c. In particular, the association between cold water in the tropical Pacific, positive SOI, and lower pressure and increased convection/precipitation over tropical South America (Fig. 18d) has been noted in the observations for the southern summer season by Aceituno (1988) and Ropelewski and Halpert (1989). But the observed system involves another set of processes and interactions in the western Pacific and South Pacific convergence zone (SPCZ) regions that are not present in the coupled model. In particular, the annual cycle of the convective maximum and its movement from the Indian monsoon region in northern summer into the Pacific in northern winter and spring (Meehl 1987) is not well-simulated by the coupled model. As outlined by Meehl (1987) and shown in Figs. 18d–f, the maximum of convection in the Australian monsoon region in DJF$_0$ (Fig. 18d) is intensified by the relatively warm SSTs set up in that region from the previous annual cycle and leaves cool SSTs in its wake as it moves out to the SPCZ and the tropical western and central Pacific in MAM$_0$ (Fig. 18e). The warm water in the western Pacific is probably moved east by advective processes associated with the westerly wind-stress anomalies in the western Pacific (e.g., Gill and Rasmusson 1983). The two positive anomaly areas meet in the central equatorial Pacific by JJA (Fig. 18f, compare to Figs. 11c, d). The transition is then complete and ends up similar to the model pattern in Fig. 18c for that season. Different combinations of these processes could produce different types of transitions as manifested by the various observed warm and cold events. The common theme that emerges is that all of the relevant processes occur in the mean seasonal cycle and that the interannual events are modulations of the mechanisms associated with the seasonal cycle.

The principal conclusions are:

1) The coarse-grid, global, coupled ocean–atmosphere model displays internally generated active interannual variability in the tropical Pacific that resembles some aspects of observed ENSO phenomena.

2) Sets of coupled anomalies in the model (SST, SLP, wind stress, convection/precipitation) appear in the far eastern Pacific during northern spring and move west.

3) The movement of the anomalies does not easily relate to or seem controlled solely by westward Rossby wave propagation (too fast, possibly due to distortion from the coarse grid) or, taking into account the weak zonal temperature gradients, mean westward advection (too slow). Instead, the anomalies depend on a kind of coupled propagation whereby SLP gradients, set up at the surface and associated with the precondition of SSTs in the tropical Pacific and the land–sea contrast in the eastern Pacific, produce ageostrophic near-equatorial, westerly wind-stress anomalies and suppressed upwelling to the west. This causes SSTs to warm to the west and the whole system of coupled anomalies is pulled to the west. Horel (1982) has suggested this type of coupled interaction to explain the long-term mean seasonal cycle in the eastern Pacific. This unstable coupled mode may be a version of what is seen in the more simple mechanistic models.

4) The mean seasonal cycle of SST, SLP, wind stress, etc. in the eastern Pacific is modulated in the observations and the model to produce the interannual events via the same sets of coupled processes.

5) By weakening the coupled strength (reducing the wind-stress forcing), the interannual variability and seasonal dependence vanish. By turning off the seasonal cycle, the nature and regular evolution of the interannual events are altered.

Because a transition from warm to cold phase or back again in the model (and the observations) depends only on the precondition of the SSTs in the tropical Pacific and the passage of the forcing from the annual cycle in northern spring, the system should have an inherent biennial tendency as is observed (Meehl 1987; Kiladis and van Loon 1988). That is, if there is cool water in the equatorial Pacific in northern spring, the system should make a transition to warm water there, and vice versa if the system starts out with warm SSTs in the tropical Pacific. Yet, as in the real ocean–atmosphere system, multiyear events disrupt the biennial cycle. The cause of these multiyear events is under investigation and appears to involve heat storage in the upper ocean.
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