On the Role of Off-Equatorial Oceanic Rossby Waves during ENSO*

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ABSTRACT

Recent theoretical and numerical modeling studies of the coupled tropical atmosphere-ocean system suggest that equatorial ocean wave dynamics may play an important role in the evolution of ENSO (El Niño/Southern Oscillation). These studies emphasize that the oceanic wave signal is confined to within a narrow equatorial band (within 6° of the equator).

In this study we use a coupled atmosphere-ocean model to investigate the role of off-equatorial Rossby waves observed in the western North Pacific Ocean during the ENSO cycle. We find that these off-equatorial Rossby waves (found poleward of 6° from the equator) are formed through both eastern boundary reflection of the equatorial Kelvin wave signal generated in a warm event (El Niño), and changes in the off-equatorial wind stress curl. Our results indicate that, independent of the generation mechanism, off-equatorial Rossby waves should be thought of as the product and not the triggering mechanism for an ENSO event.

1. Introduction

In some recent studies [White et al. 1985, 1987 (hereafter WPI); Graham and White 1988] it has been suggested that anomalous low-latitude oceanic Rossby waves may be a precursor to an El Niño/Southern Oscillation (ENSO) event. These off-equatorial Rossby waves are identified from data in a latitude band spanning about 10° to 20°N across the Pacific. All of these studies draw on data from the period 1964–85, in which ENSO events occur quasiperiodically. Although these studies do not address the origin of these off-equatorial Rossby waves, the waves are implicitly presumed to be produced by low-latitude wind stress curl during an ENSO event, and the model of McCreary (1983) is cited as the framework for this interpretation of the data. In this study we will reexamine the origin of these off-equatorial Rossby waves, and their role in ENSO. In particular, we will caution against interpretation of these off-equatorial Rossby wave signals as either a precursor or trigger for ENSO.

As WPI draw their framework for their interpretation of off-equatorial Rossby waves from McCreary (1983), it is useful to highlight first the similarities and differences of this work with that of McCreary (1976) and with more recent findings of Cane and Zebiak (1985), Schopf and Suarez (1988), Battisti (1988), and Battisti and Hirst (1988). This is done in section 2. In section 3, a numerical coupled atmosphere–ocean model is used to examine quantitatively the origin and role of off-equatorial Rossby waves during ENSO. The conclusions and discussion are presented in section 4.

2. The origin of off-equatorial Rossby waves associated with ENSO

a. Theoretical background

In an important study of the dynamics of equatorial oceans, McCreary (1976) examined the response of the ocean to changes in the zonal wind stress τ^x analogous to that occurring during El Niño, by prescribing zonal wind stress patches with a variety of latitudinal distributions. In that study, McCreary concluded that "changes in the zonal wind field outside of the equatorial band (roughly ±5° of latitude from the equator) are not important for *generating* [italics added] El Niño." We note McCreary also suggested a significant off-equatorial Rossby wave signal could be generated by the eastern boundary reflection of an incoming Kelvin wave, itself forced within the equatorial band.

Nonetheless, in a later study McCreary (1983, hereafter M83), examined the evolution of an ENSO event in a coupled atmosphere/ocean model, where the atmosphere is treated as completely parameterized by a zonal wind which extends well into the subtropics. The ocean model was identical to his first study. The spatial structure of the atmosphere consists of prescribed patches of zonal wind stress anomalies. The wind patches are switched on and off, depending on the value of the upper layer thickness perturbation h on the east-

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ern basin boundary on the equator ($h = h_E$). Here h_E is assumed to be a proxy for the SST in the eastern Pacific. In particular, when the thickness anomaly h_E exceeds some critical value h_C the zonal wind stress takes the form

$$\tau^x = \tau^H; \quad h_E > h_C \tag{2.1}$$

whereas, when h is shallower than h_C

$$\tau^x = \tau^W; \quad h_E \leqslant h_C \tag{2.2}$$

where τ^W is an equatorially symmetric, zonal wind stress patch, confined to the central equatorial basin, and is thought to be associated with the Walker circulation anomalies when the system is in a "cold" state. The τ^H, having a cosine-bell distribution in latitude with a maximum on about 15° latitude throughout the eastern Pacific, is presumed to be associated with an increase in the Hadley circulation in the eastern Pacific during a warm event due to the eastward displacement of the Indonesian convective center. However, as McCreary stated, parameterization (2.1) is not consistent with the observations (e.g., see Rasmusson and Carpenter 1982). In fact, a more realistic parameterization for the wind stress anomalies in the warm state, albeit crude, would be

$$\tau^x = -\tau^W; \quad h_E > h_C.$$

Yet because the winds were parameterized as a switch in Eqs. (2.1)–(2.2), the existence of an off-equatorial τ^H is essential for model variability on interannual time scales. In M83, τ^H acts as a very efficient generator of large off-equatorial (10°–20° latitude) Rossby waves during ENSO which propagate very slowly (a few years time) across the Pacific from the generation region in the east, reflect at the western boundary sending a Kelvin wave signal back to the eastern boundary which triggers the wind switch, and thus sends the coupled system into a cold or warm state. The long time scales for ENSO in this model are solely because of the assumption of *off-equatorial* Rossby waves generated by the unrealistic, large τ^H. As τ^H is moved towards the equator, a more realistic parameterization of the observed wind field is achieved ($\tau^H \rightarrow -\tau^W$). Now the Rossby waves generated by τ^H are closely trapped about the equator. Hence, the Rossby signal propagates across the basin faster, producing a rapid (about 9 month) oscillation of the coupled system, and no interannual variability characteristic of ENSO. McCreary and Anderson (1984) discarded the unwarranted assumption of a high-latitude wind source for Rossby wave activity during ENSO, and invoked a seasonal cycle argument to obtain interannual variability in a coupled model. However, through the work of Anderson and McCreary (1985), and Cane and Zebiak (1985), it is now generally agreed that the seasonal cycle is probably not essential for generating ENSO.

Schopf and Suarez (1988) and Battisti (1988; hereafter B88) suggested that the essential dynamical aspects of the ocean during ENSO is confined within the equatorial wave guide. Both of these studies indicate that the important dynamical response for ENSO is confined to well within 8° of the equator in the form of the gravest symmetric Rossby modes and the equatorial Kelvin signal. This is demonstrated (B88) through a modal decomposition of the oceanic response of the coupled atmosphere–ocean system, and is in agreement with the conclusions of McCreary (1976). In fact, filtering the winds to zero beyond 5° of the equator (introducing a severe deviation in the model's wind stress curl field in the subtropics) produced no significant changes in the coupled model ENSOs (B88), further suggesting the dominance of low-latitude dynamics.

To some extent the dynamic development of ENSO as outlined by Schopf and Suarez (1988) and B88 would be related to that of McCreary (1983) *if* τ^H were centered on the equator and hence actually responsible for the generation of equatorially confined wave modes. However, in the latter study the ensuing oscillations were intraannual rather than interannual. The discrepancy with the former studies is due to the crude temporal nature (a switch) of the wind stress parameterization used in M83. In nature, the relationship between h (i.e., SST) and τ^x is complicated. In fact, it has been suggested by Schopf and Suarez (1988), and Battisti and Hirst (1988), using coupled atmosphere–ocean numerical models, there are two competing time scales for the coupled tropical atmosphere–ocean system: the equatorial Rossby–Kelvin wave transit time noted above [O(9 months)], and the time scale associated with the localized, coupled unstable atmosphere–ocean system in the eastern Pacific (the latter time scale is less than 6 months; see also Philander 1985; Lau 1981). The interplay between these two, short time-scale processes has been shown to produce interannual variability without invoking the need for off-equatorial oceanic Rossby waves (B88 and Battisti and Hirst 1988). In the numerical models, off-equatorial Rossby waves do in fact exist which are compatible with those observed in data, but they play a relatively passive role in the ENSO cycle, as demonstrated below.

b. Origin of the off-equatorial Rossby waves

White et al. (1987) present observational evidence for off-equatorial Rossby waves associated with ENSO events. There are at least two ways to generate these off-equatorial Rossby waves. One mechanism is direct forcing through wind stress curl at off-equatorial latitudes during a warm event (M83). However, as noted by M83, the wind stress curl in the subtropics is not consistently present in the observations during an ENSO event (see Rasmusson and Carpenter 1982). A second mechanism for off-equatorial Rossby wave generation is indirect, through reflection at the eastern ocean boundary of Kelvin waves (see McCreary 1976), which are forced by the zonal wind stress anomalies confined to within the equatorial band.

There are off-equatorial Rossby wave signals in the coupled models of Cane and Zebiak (1985), Schopf and Suarez (1987), and Philander and Lau (1988). In the Cane and Zebiak model, the signals poleward of about 12° latitude are somewhat sensitive to the off-equatorial wind stress curl (B88). However, we note the numerical model overestimates the wind stress curl forcing outside of the equatorial band (not shown).

Our calculations, consistent with the uncoupled ocean model results of WPI, demonstrate that off-equatorial Rossby waves observed (e.g., WPI) in the band spanning 6° to 12° latitude are indeed associated with ENSO events, but are formed primarily via Kelvin reflection at the eastern boundary rather than by off-equatorial wind stress. On interannual time scales the forcing of a Kelvin wave is done most efficiently during the peak of an ENSO event (largest zonal wind stress anomalies). The off-equatorial Rossby signals should appear immediately after each warm (ENSO) event in the eastern Pacific near the coast and somewhat later farther west into the Pacific basin. The time between the peak of the ENSO event and the peak off-equatorial Rossby wave signal at any longitude will increase as one goes poleward from the equator.

3. Interpretation of off-equatorial Rossby waves in ENSO

In a periodic regime for ENSO the off-equatorial Rossby waves generated by a (warm) ENSO event would be exactly correlated with any index for ENSO at some point of the cycle, with time lags of periodic interval. However, this does not necessarily imply that the off-equatorial Rossby waves are the trigger for ENSO, as suggested by WPI. To demonstrate this, a model of the coupled atmosphere/ocean system will be used which follows closely that described by Cane and Zebiak (1985) and Zebiak and Cane (1987), and is briefly described below. The ocean component of the model consists of an upper layer, topped by a fixed depth surface (mixed) layer, overlaying a deep motionless layer. SST is calculated separately and does not directly affect the ocean dynamics. The surface currents are driven by the wind stress and retarded by Rayleigh friction. SST is changed by nonlinear advection, upwelling, and heat fluxes to the atmosphere. The upwelling is prescribed in terms of the divergence of the mixed layer currents and represents the entrainment into the surface layer. The upper layer, which includes the surface layer, is governed by linear shallow water wave dynamics. The atmospheric component of the model is a simple linear reduced-gravity model of a thermally forced tropical atmosphere (Gill 1980). The forcing of the atmosphere depends on the total atmospheric convergence and the initial SST perturbation and is calculated iteratively (Zebiak 1986). The ocean is forced by the anomalous wind stress and the atmosphere is forced by latent heat release, which is a function of the convergence of the wind field and the SST anomaly. Both of the forcing terms are nonlinear. The ocean model domain is a rectangular basin (30°N to 30°S, 124°E to 80°W); the atmosphere is modeled on an equatorial β-plane. For a complete model description, see Zebiak and Cane (1987) or B88.

We will focus on the seasonless cycle run with the standard model physics (experiment 3 of B88). Hovmöller diagrams of the SST and upper layer thickness anomaly (h, crudely interpreted as the pycnocline or upper layer heat content) from this experiment are displayed in Figs. 1 and 2. The model ENSO events are periodic at 3.47 years, with h anomalies in the eastern Pacific leading the SST anomalies by roughly 70 days. A plan view of the upper layer thickness anomalies at various model times is presented in Fig. 3. At the peak of a model ENSO event (t = 11 550 days; Fig. 3a) the off-equatorial Rossby waves at 15° north and south are well pronounced, coming off the eastern boundary due to the reflection of the Kelvin signal. A little over two years later (t = 12 330 days; Fig. 3b) this pycnocline depression has propagated across the basin, and is arriving at about 160°E, when the eastern Pacific is still cold from the collapse of the ENSO event that peaked at t = 11 550 days. About 15 months later (t = 12 810) the next warm event is almost at a peak (Fig. 1). Thus, for the choice of free wave speed in this model (c = 2.98 m s^{-1}), the pycnocline response at about 10°N and S, 160°E is exactly correlated with the maximum SST anomaly in the eastern Pacific with either a 2–2.5 year lag *or* 1–1.5 year lead. This correlation is also seen in Fig. 4, which displays a latitude–time plot of h along 160°E. The exact lag or lead of h (160°E) with SST in the eastern Pacific (Fig. 1) is very sensitive to the latitude at which h is monitored (hence the value of c). These waves are generated by both reflection and local wind stress. The results for this experiment (relevant to ENSO) are very similar to the same experiment done with the winds outside of 5° of the equator set to zero (introducing severe changes in the off-equatorial wind stress curl).

Finally, the issue of causality for ENSO in relation to off-equatorial Rossby waves is examined. WPI and Graham and White (1988) have suggested these off-equatorial Rossby waves are a trigger for ENSO. In a periodic or quasiperiodic regime, the above results suggest that correlation of off-equatorial Rossby wave activity with the onset of ENSO is not a rigorous test of causality. The question "Can the trigger mechanism for ENSO originate from Rossby signals at off-equatorial latitudes" is most easily addressed in the context of the numerical model. Here we pose the question: "Where is the primary contribution to the Kelvin waves, generated by western boundary reflection,

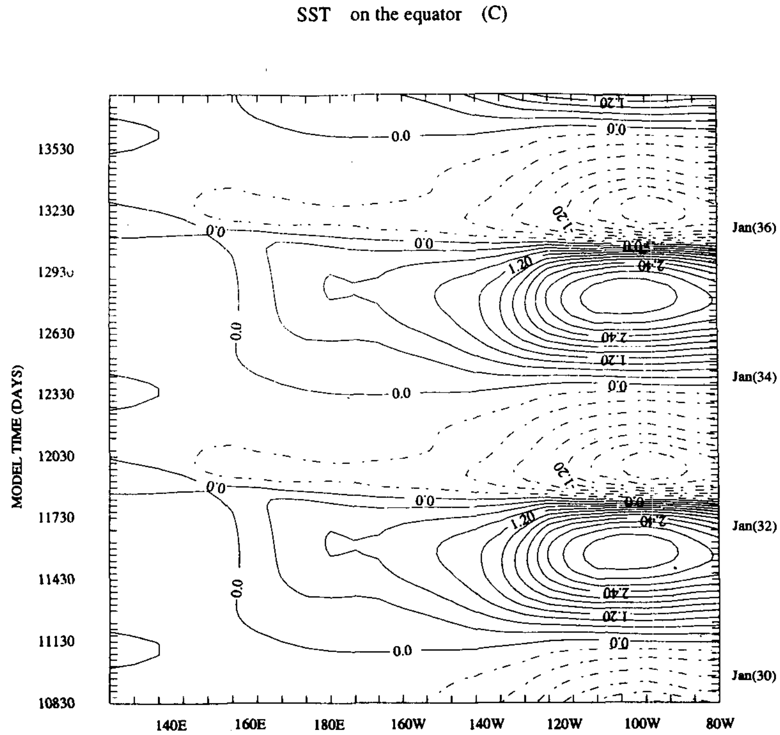


FIG. 1. Evolution of equatorial anomaly SST field (averaged between 2°N and 2°S) for equilibrated nonlinear model between the indicated times, from the coupled numerical model experiment 3 described in Battisti (1988). The time is in model days; January of model year 32 is denoted Jan (32), etc.

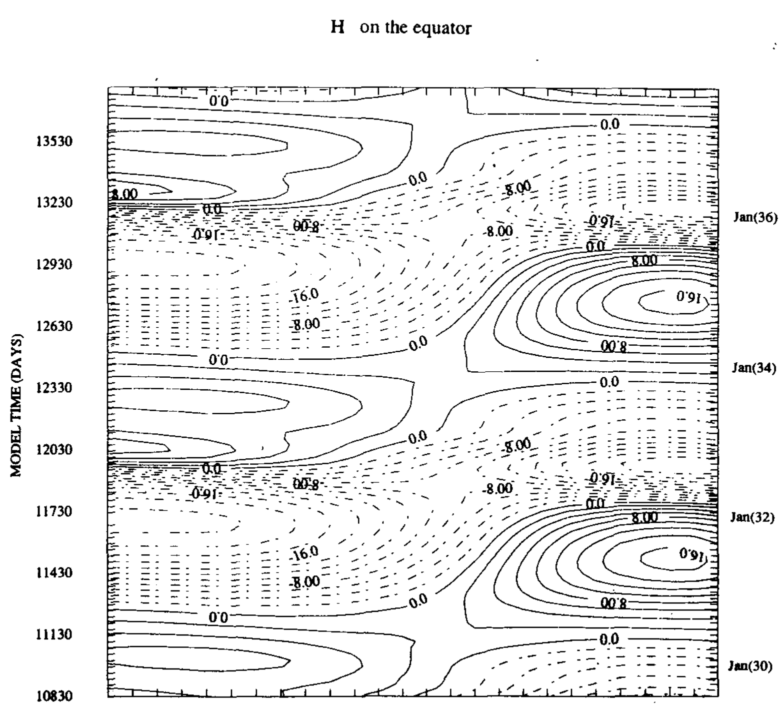


FIG. 2. As in Fig. 1, except for the model upper layer thickness anomaly. Units are in meters.

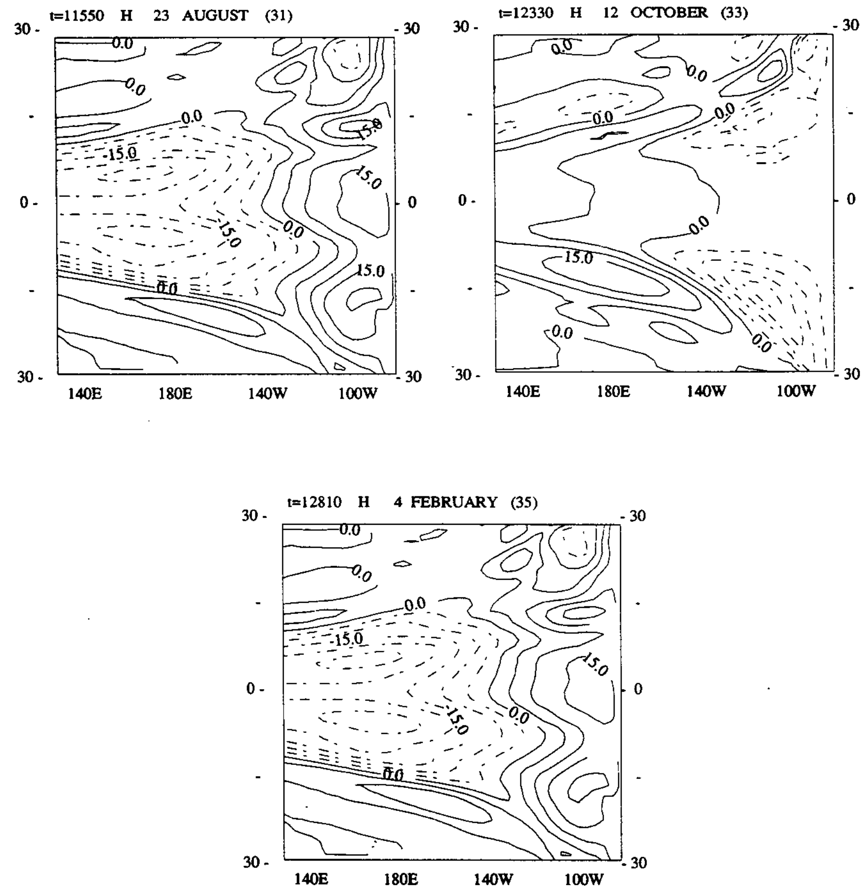


FIG. 3. Plan views of the model upper layer thickness anomaly for various times in the model ENSO cycle from the coupled numerical model experiment 3 of Battisti (1988). The time is in model days; the contour interval is 5 meters.

coming from?" From the model we can calculate the (nondimensional) amplitude of the Kelvin wave signal, q_0, generated from the reflection of a collection of Rossby waves coming into the western boundary, described by the zonal velocity field, u_R (y, x = western boundary). In the longwave approximation (Cane and Sarachik 1977)

$$q_0 = -2 \int_{y_s}^{y_N} u_R\,(y, x = \text{western boundary})\,dy$$

where u_R (y, x = western boundary) is the zonal velocity due to the westward propagating long wave Rossby signal. To examine the contribution of different latitude bands to the reflected Kelvin wave signal q_0, we have plotted in Fig. 5 the running integral

$$K(y) = -H_0 \int_{y_s}^{y} u_R dy \qquad (3.1)$$

starting from the southern boundary, $y = y_s$ (H_0 is the fixed upper layer depth 150 m). $K(y)$ represents the total Kelvin wave upper-layer thickness (pycnocline) perturbation on the equator due to the incoming accumulated Rossby wave signal from y_s to the latitude of interest. At all times in the integration, the entire Rossby contribution to q_0 is contained within 8° of the equator. There is never any significant Kelvin wave signal generated from Rossby waves incident on the western boundary poleward of 8°N or 8°S.[1] Therefore, during a period of highly irregular ENSO activity, the off-equatorial Rossby wave activity may better be thought of as a result of the previous event than as a precursor to future events. This result is consistent with the experiments of Battisti (1988), where severe modification of the wind stress curl outside of the equatorial band was found to have no effect on the physics of model ENSO event evolution.

To demonstrate the passive role of off-equatorial Rossby waves during ENSO we ran the standard phys-

[1] This conclusion is also reached by repeating the coupled model calculation, explicitly filtering out the off-equatorial signals by considering only the contribution to q_0 between 8°S and 8°N, whereupon there is no significant difference in the model response.

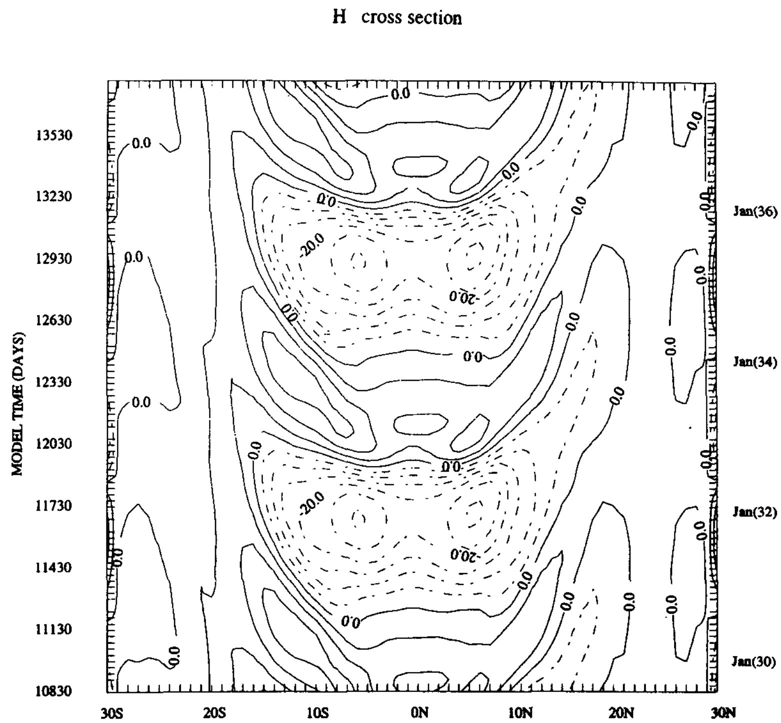


FIG. 4. Latitude–time plot for the upper layer thickness anomaly along 160°E, from the numerical experiment 3 of Battisti (1988). Time is in model days; the contour interval is 5 meters.

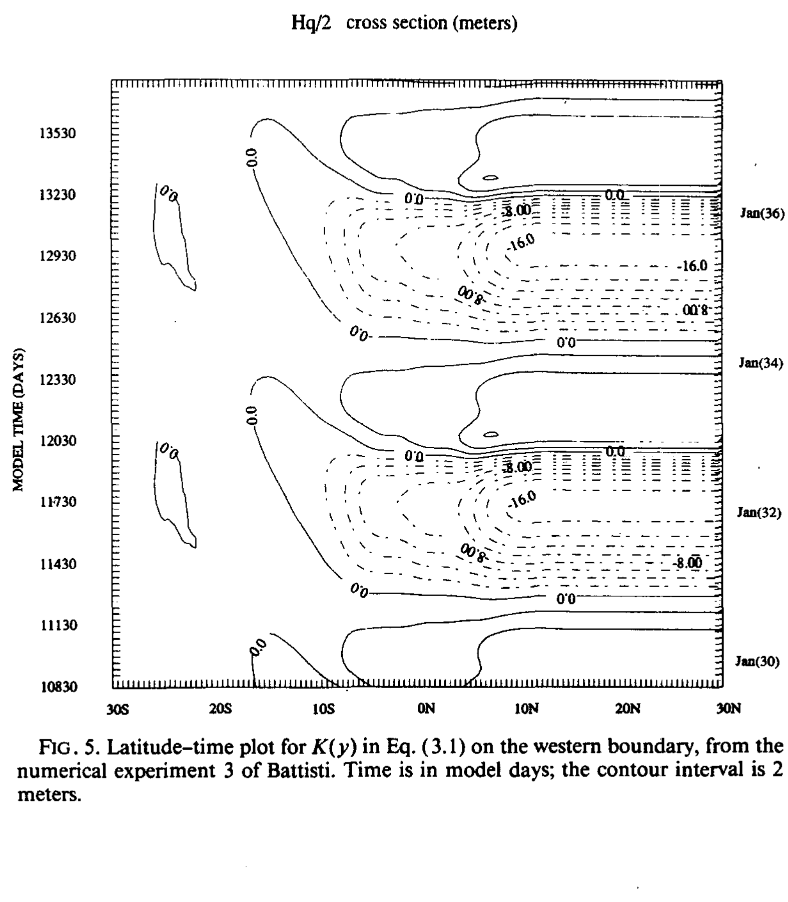


FIG. 5. Latitude–time plot for $K(y)$ in Eq. (3.1) on the western boundary, from the numerical experiment 3 of Battisti. Time is in model days; the contour interval is 2 meters.

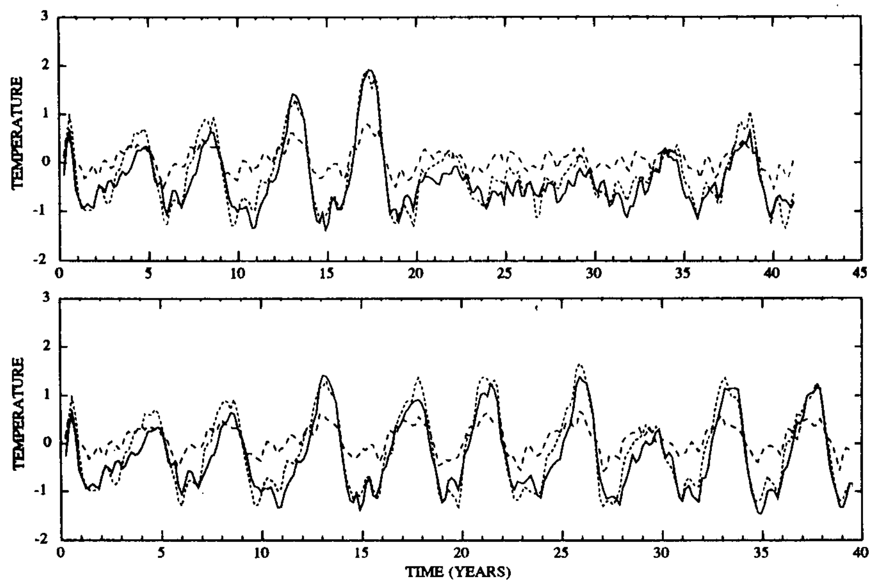


FIG. 6. SST indices for the first 45 and 40 years of two runs of the standard physics model including a parameterized 30–60 atmospheric signal and some additional noise in the zonal wind. In the model, Niño 1 (solid line) extends from 10° to 5°S, and 6° offshore (equivalent of 87°W), Niño 3 (dotted line) covers the area (5°S to 5°N, 90° to 150°W), and Niño 4 (dashed line) spans (5°S to 5°N, 150°W to 160°E).

ics model which includes a basic state with a seasonal cycle (see experiment 3 of B88), adding some external high frequency "noise" to the atmosphere zonal wind signal to make the interannual variability in the model irregular. This external high-frequency noise is of the form

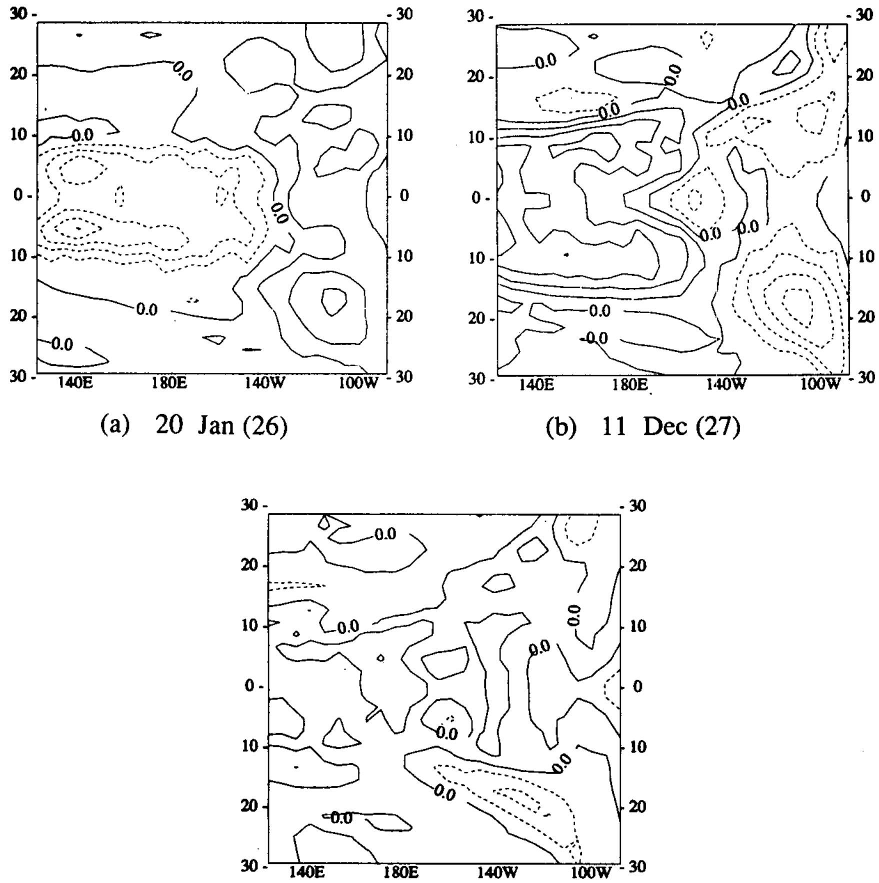
$$U_a{}^n = \sqrt{2}U_{MJ}\frac{x'}{L_x}e^{(1/2)-(x'/L_x)^2-(y/L_y)^2}, \quad x' = x - ct \tag{3.2}$$

where $L_x = 30°$, $L_y = 8°$, and $c = 8$ m s^{-1}; $U_a{}^n$ is meant to be a crude proxy to some synoptic scale variability neglected in the model atmosphere. In this case, a modest surface wind perturbation of amplitude U_{MJ} is assumed to be associated with the 30–60 day wave phenomenon. The amplitude of this noise is $(1.5 + N)$ m s^{-1}, where $N(t)$ is a random perturbation, $|N(t)| \leq 0.5$ m s^{-1}. The SST indices for the first 40 years of integration for two such runs are displayed in Fig. 6. By adding the high frequency noise to the atmosphere, the interannual variability in the coupled atmosphere–ocean system is now very irregular (Fig. 6).[2] This is consistent with the results of Schopf and Suarez (1988), who concluded irregular interannual variability in the coupled system can be attributed to "weatherlike" events in the atmosphere. We will now focus on the nonevent in year 29 of Fig. 6b. As in the periodic (non-external noise) run (Fig. 3a), the warm event peaking in Jan(26) produces "blobs" of Rossby wave activity (positive pycnocline depression) at about 15°N and S of the equator in the eastern basin (Fig. 7a). About two years later this Rossby wave signal has moved into the western basin, indicated by the large pycnocline depressions at about 10° off the equator in Dec(year 27) (Fig. 7b, see also Fig. 3b). However, this Rossby wave signal is not associated with a warm event in the fall of model year 28 (Fig. 6b; cf. Figs. 7c and 3c). Rather, the atmosphere–ocean state is near to climatology from the end of year 28 through model year 29, whereafter the system plunges into a cold state! Hence, the off-equatorial Rossby wave activity *fails* as a precursor for an upcoming ENSO event. Alternatively, the passivity of the extratropical Rossby waves in ENSO evolution is demonstrated by examining the quasiperiodic epochs (e.g., from model year 10 through year 27 in Fig. 6b). Not all ENSO events in the quasiperiodic regime are preceded by extratropical western Pacific Rossby wave activity; there is significant wind stress curl associated with (3.2) to modify these off-equatorial wave signals.

[2] It is interesting to note that adding noise to the seasonless model does not create irregular interannual variability. Both atmospheric noise *and* a time interval where the basic state for the coupled system is stable (January–February) are required for irregular interannual variability (see Battisti and Hirst 1988).

4. Conclusions and discussion

In this paper a coupled atmosphere–ocean numerical model was used to investigate the role of off-equatorial Rossby waves during the ENSO cycle, motivated by the interpretation of these signals seen in observations



(a) 20 Jan (26)

(b) 11 Dec (27)

(c) 4 Feb (27)

FIG. 7. Snapshots of the model h field for the run displayed in Fig. 6b. The time for each snapshot is: (a) 20 Jan (26), (b) 11 Dec (27) and (c) 4 Feb (29). Contour interval is 5 meters.

by White et al. (1987). We conclude, to a large extent, that these off-equatorial Rossby waves are best described as originating from both an eastern boundary reflection of the equatorial Kelvin wave signal generated in a warm event (El Niño), and changes in the off-equatorial wind stress in the central-eastern Pacific. The Rossby wave signals outside of 8° latitude of the equator provide virtually no contribution to the returning Kelvin wave which is essential in the model ENSO evolution. The essential dynamical processes for ENSO are contained within the equatorial band (within 6° of the equator). Caution is required, therefore, in interpreting off-equatorial signals in data. In a quasiperiodic regime (e.g., the last two decades), it is shown using a numerical coupled atmosphere–ocean model that off-equatorial Rossby wave activity should be associated with the ENSO cycle as the observations indicate, and be passive in the cycle. Our results indicate that, in general, these off-equatorial Rossby waves, upon western boundary reflection, should not be thought of as the triggering mechanism for an ENSO event.

These conclusions are based on calculations using a simple ocean model assuming linear equatorial ocean dynamics, a single dominant vertical mode, ignoring wave–wave and wave–mean flow interactions, etc. It is possible that these neglected processes may, in more realistic geometry and oceanic conditions, conspire to allow large off-equatorial Rossby wave signals which trigger ENSO events. However, these conclusions are not inconsistent with the result from the coupled atmosphere–ocean general circulation model (GCM) being run at GFDL (Philander et al. 1989). This model is essentially the GFDL tropical Pacific ocean GCM (see Philander et al. 1987), coupled to the global atmosphere GCM of Lau (1985). After 20 years of integration, at least one ENSO event (and one nonevent) occur in the model which evolve in a manner consistent with our conclusions using the simpler coupled model. Specifically, one ENSO (cycle) appears to evolve in a manner consistent with that described by Schopf and Suarez (1988) and Battisti (1988), with wave dynamics in the equatorial band playing a leading role. Interestingly, this event is apparently triggered by Rossby waves in the western Pacific within the equatorial band. These waves are reflected from the ocean western boundary and send the Kelvin signal back into the eastern Pacific, initiating the warm event. A second, nonevent, is also of interest. In this case, some Rossby activity is seen propagating into the western Pacific as in the previous case, only at a much higher latitude band (between 10° and 15°N). Not surprisingly, when this Rossby wave activity reaches the western boundary it does not trigger a subsequent warm event.

Of course, models are only tools with which we may formulate our ideas. We await a more extensive oceanic dataset, spanning both irregular and quasiperiodic epochs for ENSO, to verify the hypothesis.

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NOTES AND CORRESPONDENCE

Local and Remote Forcing of ENSO Ocean Waveguide Response—D. E. HARRISON, NOAA/PMEL, Seattle, Washington.

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Comments on "Variations of Whitecap Coverage with Wind Stress and Water Temperature"—EDWARD C. MONAHAN AND DAVID K. WOOLF, Marine Sciences Institute, University of Connecticut.

Reply—JIN WU, Air-Sea Interaction, University of Delaware.