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Interannual oscillations in an ocean general circulation model coupled to a simple atmosphere model

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A model is being developed for tropical air–sea interaction studies that is intermediate in complexity between the large coupled general circulation models (GCMs) that are coming into use, and the simple two-level models with which pioneering El Niño Southern Oscillation studies were done. The model consists of a stripped-down tropical Pacific Ocean GCM, coupled to an atmospheric model that is sufficiently simple that steady-state solutions may be found for low-level flow and surface stress, given oceanic boundary conditions. This permits examination of the nature of interannual coupled oscillations in the absence of atmospheric noise. In preliminary tests of the model the coupled system is found to undergo a Hopf bifurcation as certain parameters are varied, giving rise to sustained three to four year oscillations. For stronger coupling, a secondary bifurcation yields six month coupled oscillations during the warm phase of the El Niño-period oscillation. Such variability could potentially affect the predictability of the coupled system.

Introduction

Pioneering studies of interannual variability due to tropical air–sea interaction were done with coupled shallow water or two-level models, including those of Philander *et al.* (1984), Hirst (1986), Zebiak & Cane (1987), Schopf & Suarez (1988) and Battisti (1988). Currently a few groups are undertaking coupled general circulation model (coupled GCM or CGCM) studies, particular at the Geophysical Fluid Dynamics Laboratory (GFDL) in Princeton, at the Max Planck Institute in Hamburg, and at the British Meteorological Office.

Because CGCMs are costly to run and because their complexity can make it difficult to diagnose the modelled phenomena, simpler models such as those mentioned above will remain valuable tools for the eventual understanding of tropical air–sea interactions. The model described here lies at a level of complexity intermediate between these two classes. This 'intermediate coupled model' is being developed with the aim of expanding the heirarchy of coupled models available for coupled air–sea interaction studies.

The choice of coupling an ocean GCM to a simpler atmospheric model was made specifically to eliminate atmospheric 'noise' from the coupled system. Atmospheric GCMs produce a wide spectrum of variability that is independent of coupled processes. The inherent high-frequency variability can considerably disrupt coupled oscillations on interannual timescales. This makes it difficult to distinguish between inherently chaotic vacillations, regular oscillations disrupted by noise, or a series of events each of which is set off by an atmospheric perturbation. The underlying hypothesis of this model, consistent with the results of the simpler coupled models, is that atmospheric noise is not essential to the oscillations. Ideally, to the extent that the simple atmospheric model can mimic the time average response of an atmospheric GCM, the intermediate coupled model should represent a distillation of the underlying oscillation mechanism in the CGCMS as it would appear in absence of atmospheric noise.

Of course, simple atmospheric models are far from perfect, nor do the existing ones agree with one another. Comments on this are included in the penultimate section. The version of the intermediate model presented here is preliminary in nature and a number of approximations are made that it would be desirable to improve upon in the long run. However, even these initial experiments suggest some interesting inferences about El Niño and coupled oscillations.

The model

Ocean

The ocean model is related to the tropical ocean GCM of Philander & Pacanowski (1984), which is in turn adapted from the model of Bryan & Cox (1968). It includes Richardson-number-dependent vertical mixing, which helps to maintain a sharp and reasonably realistic thermocline during long model integrations. When forced with observed wind stress, the model has been shown to reproduce realistically the seasonal cycle in the Atlantic (Philander & Pacanowski 1984), the 1982–83 El Niño (Philander & Siegel 1985) and the seasonal cycle in the tropical Pacific (Philander, Hurlin and Siegel (1987) hereafter refered to as P.H.S.)

To make extended integrations within reasonable limits of computer resources, the resolution in this model is reduced by about a factor of three in every direction relative to the P.H.S. version. A spacing of 1° latitude is used in the band between 10° S and 10° N, with the mesh spacing increasing smoothly to 4° at the southern boundary at 30° S and 5.6° at the northern boundary at 50° N. Thermal damping toward an observed temperature profile and increased horizontal mixing are included near the north and south boundary, as in P.H.S. A longitudinal mesh spacing of 3° is used. Realistic coasts are not included in these preliminary experiments, but the domain of the box is very similar to the Pacific Ocean, reaching from 130° E to 80° W. Ten vertical levels are used, seven of which occur in the top 300 m. Salinity is neglected and a linear equation of state is used.

The surface boundary conditions applied to the model for the simulation of the climatology are taken from observations. The annual average wind stress from Hellerman & Rosenstein (1983) is multiplied by a factor of 0.7 because it is considered to be too strong by about this amount. A surface heat budget is included, as in P.H.S., with bulk aerodynamic formulas for evaporation and sensible heat, with the near surface air temperature specified from observations (Levitus 1982) and the relative humidity assumed constant at 0.8.

Two main ocean climatologies are considered in this series of experiments. In each case, the ocean was spun up from rest with constant annual average surface stress and air temperature applied in the upper-boundary condition. A good approximation to a steady state is reached at three years. The structure of the equatorial currents and upwelling is in good agreement with the higher-resolution P.H.S. simulation, as is the temperature field at depth.

Sea surface temperature (SST), the most important field as far as coupling to an atmospheric model is concerned, is shown for climatology I in figure 1. Considering the box geography, the ocean model produces a very reasonable climatology, with a warm pool of roughly the right extent in the western equatorial Pacific and cold water along the Equator and the coast of 'South America'. The Indonesian warm water is even slightly divided into two lobes; the larger lobe, south of the Equator, produces the South Pacific convergence zone when the

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atmospheric model is included, whereas the warm water north of the Equator produces the Inter Tropical Convergence Zone (ITCZ) in the east. The warmest water is about 31 °C, slightly warmer than observed but well within reason.

The equatorial cold tongue is slightly too strong and extends too far west relative to observations. This is a deficiency commonly encountered in GCM simulations of the Pacific Ocean. During the course of the coupled modelling experiments, it was hypothesized that this tends to work against interannual oscillations. It therefore seemed desirable to produce an alternative climatology with a weaker cold tongue, and the most expedient means of doing this was to reduce the climatological wind stress. The wind stress used to produce climatology I was multiplied by a factor of 0.6 and a similar three year spin up from rest was used to produce climatology II. The equatorial upwelling in this climatology is reduced roughly in proportion to the reduction in the stress while the currents are much less affected. The resulting SST pattern is rather similar to that of climatology I in figure 1, except that the minimum temperature in the equatorial cold tongue is about 2.5 °C warmer. The horizontal temperature gradient along the Equator and the vertical temperature gradient in the eastern Pacific are both somewhat reduced. Ideally, it would be desirable to modify the properties of the cold tongue by an improvement in the physics, such as ameliorating the mixing parametrization. For initial experiments, adjusting the climatological wind stress is used for simplicity.

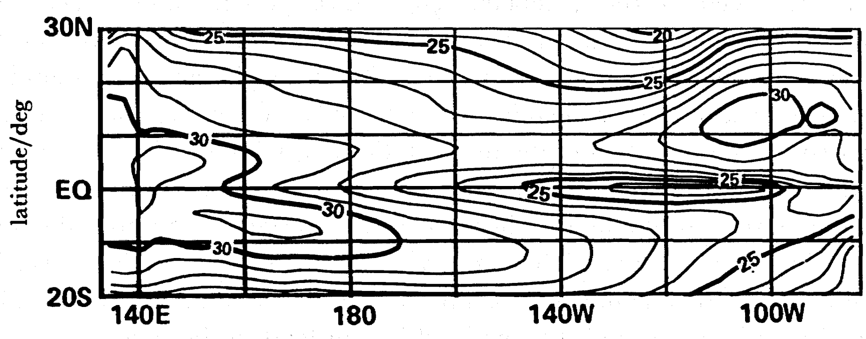

FIGURE 1. Sea surface temperature of ocean model climatology I (forced by observed wind stress). Contour interval 1 °C.

A westerly stress anomaly in the western Pacific tends to accelerate eastward currents and move warm water towards the east, as occurs in the early stages of El Niño (Rasmussen & Carpenter 1982). The response of the ocean to such an anomaly (0.3 dyn cm^{-2}† added to the climatological stress between 10° S and 10° N and west of 170° W), applied for one month and then removed again was examined as a control run for similar experiments conducted with the coupled model. SST returns rapidly to equilibrium, essentially by the third month. The phase speeds of the Kelvin and Rossby wavefronts that complete this adjustment appear reasonable compared with the P.H.S. model.

Atmosphere

The atmospheric model used in this coupled system is a combination of the models of Neelin (1988) and Neelin & Held (1987) (hereafter: N. and N.H.) The first of these gives a steady solution for surface stress when forced by a given time-averaged vertical velocity at the top of the Trade cumulus boundary layer, or, what is almost equivalent, the atmospheric latent

† 1 dyn = 10^{-5} N.

heating. The second models this vertical velocity from given oceanic boundary conditions. The first half of this approach may be regarded as fairly well understood. It has been known for some time that the linear damped model of Gill (1980) (and Matsuno 1966) gives a reasonable approximation to the low-level flow in the tropics. N. reinterprets the Gill model as a model for the atmospheric boundary layer with heavy damping due to vertical diffusion of momentum by boundary layer turbulence. Viewed in this manner, the model gives surface stress directly when forced by mid-tropospheric vertical velocity or latent heating. The model does very well when compared to the results of a GCM experiment, both for the climatology and for El Niño Southern Oscillation (ENSO) composite anomalies. A similar justification for the momentum equations of the Gill model appears in Lindzen & Nigam (1987)

In modelling the tropical atmosphere, the more difficult problem is to obtain the vertical velocity (or latent heating) from given boundary conditions. The major part of the latent heating in the tropical convergence zones is associated, not with local forcing by evaporation, but with moisture convergence by the low-level flow. The relation between the heating and the circulation must be addressed to model properly the low-level flow. Previous approaches have simply taken the heating to be a given function of sea surface temperature (Philander *et al.* 1984; Anderson & McCreary 1985) or else assumed that heating anomalies are caused by flux anomalies out of the sea surface, intensified in one manner or another by a parametrized atmospheric feedback (Webster 1981; Zebiak, 1986; Davey & Gill 1988). N.H. argue that surface fluxes are not the only process that can affect the location of the convection and that warm SSTs can directly affect the position of convection through the moist stability. N.H. propose a simple two-layer model for the mid-tropospheric latent heating and vertical velocity in the tropics, based on consideration of the moist static energy budget. The two layers of the model may be thought of as a well-mixed, moist boundary layer beneath the tropical trade inversion, and the upper troposphere, whose role is primarily to receive the mass flux convergence from the boundary layer. The temperature in the moist boundary layer is closely tied to SST and this affects the amount of moisture that this layer can hold. As a result, where SST is high the moist static stability of the two-level system tends to be small. The circulation is thermodynamically direct, forced by the net flux of energy into the tropospheric column, but the response of the model vertical velocity depends strongly on the moist stability; tropical convergence zones are associated with minima of moist stability and thus with maxima of SST.

Figure 2 shows the vertical velocity and surface stress anomalies produced by the model for the DJF(+1) phase of a composite El Niño (1969, 1972 and 1976). The patch of westerlies in the central Pacific is reproduced by the model with about the right magnitude and extent, in the right location. The absence of strong easterlies in the eastern part of the basin is also captured by this simulation. This stress pattern is driven by a strong upward vertical velocity anomaly that occurs in the central Pacific as a result of an eastward shift of the Indonesian convergence zone. Comparison to a simulation of the same composite El Niño by the GFDL climate group R15 GCM (Lau 1985) shows that the simple model properly captures the location and extent of the strong convergence. This is significant, because both the SST anomaly and the surface heat flux anomaly follow a much different pattern, extending through the eastern Pacific. The correct location of the convergence anomaly in the N.H. model simulation depends on the nonlinear relation between total SST and the boundary-layer convergence through the gross moist stability. Within the context of the model, the relative contributions of the surface flux anomaly and the moist stability can be evaluated. Although the flux

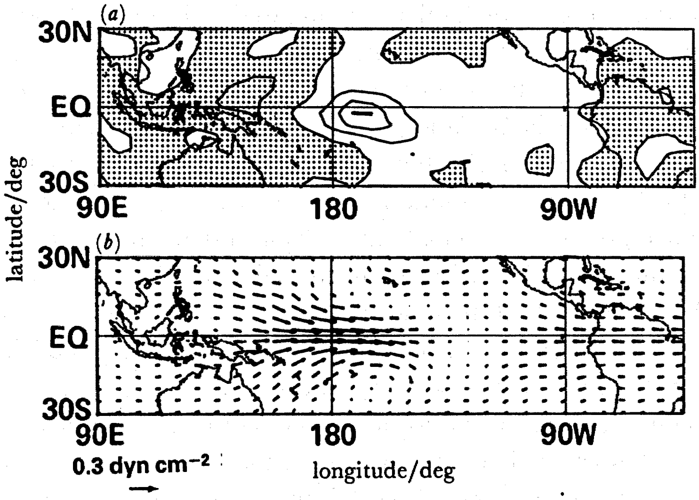

FIGURE 2. Atmospheric model stress anomalies for the DJF(+1) mature phase of a composite El Niño; (*a*) vertical velocity, (*b*) surface stress.

anomaly does contribute somewhat to the convergence anomaly, its contribution is secondary and is strongly modulated by the effects of the moist stability.

Simulation of other phases of the cycle shows that the westerly stress patch moves eastward as the event progresses from onset to mature phase, and then is replaced by an easterly anomaly associated with the cold phase of the cycle. The model has also been used to simulate the total climatology. Qualitatively, it does reasonably well at capturing the positions and extent of the major convergence zones. The strong convergence in these regions is almost entirely a result of the dependence of the moist stability on SST. The net flux forcing tends to have much slower spatial variation than the convergence field. More importantly, the maxima of the flux term do not correspond to the observed convergence zones: a model that depended only on the flux term to produce the convergence zones would do poorly at simulating the climatology.

However, the N.H. model fails to capture the climatological subsidence in the subtropics because of the neglect of transient eddy flux divergences that are important in these regions. As a result, the simulation of the climatological stress field tends to have more broadly spread regions of tropical trades in each season than are observed. For instance, in the January simulation, the trades tend to straddle the Equator, rather than being confined to the Southern Hemisphere. For purposes of coupled model studies, the ocean circulation that would be forced by this climatology would be somewhat crude. The atmospheric model is therefore used as a nonlinear anomaly model in the coupled model, as described below. The ability of the atmospheric model to simulate at least qualitatively the total climatology, however, does lend some hope that the nonlinear aspects of the simulation are being roughly captured by the model. The El Niño anomalies themselves seem to be sufficiently well simulated, including the nonlinear aspects, that the model provides a useful simple tropical atmospheric model for ENSO studies.

Coupling

Another advantage of setting up the coupled model as a 'demi-anomaly model', with the atmospheric half of the model used in anomaly mode, is that by taking care to define the atmospheric climatology in a manner that is internally consistent with the model, a self-consistent, known, stationary point of the coupled system is constructed. The atmospheric climatology is defined as the atmospheric model response to the ocean climatology. Because the ocean climatology is in equilibrium with the specified observed wind stress and air temperature, this state will also be a stationary solution of the coupled system. This approach to anomaly modelling differs from that of Zebiak & Cane (1987), or from the 'flux correction' method of Latif *et al.* (1988), in that the atmospheric climatology is defined via the total response of the atmospheric model to the ocean *model* climatology. Constructing the system such that one stationary solution is known presents certain advantages for understanding the properties of the phase space. Beginning an integration with a substantial perturbation from the stationary point, if the system evolves to a steady state other than the known one, it is clear that there exist multiple equilibria; if the system oscillates, it may be possible to understand its behaviour in terms of the instability of the known steady state, and so on.

An open question in coupling a simple atmospheric model to an ocean model is how to treat the surface sensible and latent heat flux interfacial condition. When the ocean model is run alone, the surface flux boundary condition of P.H.S. is used, which requires an air temperature specified from observations. This may be reasonable for the climatology but in the coupled system the near-surface air temperature is certainly influenced by the SST beneath it. As the SST evolves during the course of a coupled run, it is not clear how best to model the changes in the air temperature. Extrapolating a mid-tropospheric temperature from the atmospheric model to the surface is a poor solution to this problem, for two reasons. First, the two-level models give a poor representation of mid-tropospheric temperature (it is pointed out in N. that this is unimportant to the simulation of surface stress) and, second, atmospheric boundary layer temperatures depend little on the temperatures in the free atmosphere.

A sensitivity study was performed to determine the importance of the air temperature to the SST field. A climatological run was repeated with a different air temperature field: specifically, the climatological air temperature was replaced by its zonal average across the basin at each latitude. Under these boundary conditions, zonal asymmetries in the SST field could only be caused by dynamics, and not by the specified air temperature. The resulting SST pattern was very similar in most respects to the climatology in figure 1. It thus appears that the effects of the specified air temperature in the boundary condition are secondary to the effects of dynamics. This suggests that it may be reasonable, for preliminary experiments, to leave the air temperature fixed at its observed climatological value. Flux anomalies associated with SST perturbations will be over-estimated because normally the near-surface air temperature would compensate to some extent for changes in SST. This will damp SST perturbations more strongly than otherwise would be the case, but the above sensitivity study suggests this effect will be secondary. The effect on the atmospheric model will depend on the extent to which flux anomalies contribute to the heating anomalies. This effect will be small in the current atmospheric model but might be considerable in, for instance, the Zebiak (1986) model. Surface temperature outside of the ocean basin was specified in an idealized manner during the coupled runs.

From a given configuration of SST and ocean surface heat flux, the atmospheric model determines a mid-tropospheric vertical velocity field and from this surface stress. Because only anomalies are used for coupling, a reference climatology velocity is subtracted. The model climatology of vertical velocity is calculated by using averages of SST and surface flux over the last six months of the ocean climatology run as boundary conditions. The regions of strongest rising motion correspond surprisingly well to the Western Pacific (Indonesian) Convergence

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Zone, the South Pacific Convergence Zone, the ITCZ, and even the 'Central American' Convergence Zone. Over the eastern Pacific, there is weak upward motion rather than weak downward motion but this does not severely affect ENSO convergence and stress anomalies. It is worth pointing out that the strongest convergence occurs over the ocean, despite the fact that the net flux of energy into the atmosphere is considerably weaker over the tropical oceans than over land. Over the ocean, the net surface flux is downward, into the ocean, and this reduces the net flux available to force the atmosphere relative to land surfaces. However, the low gross moist stability over the warm SSTs more than compensates for this effect.

The time integration of the coupled model proceeds by running the ocean model with fixed wind stress for a certain period and averaging the SST and surface flux over this time. These averaged quantities are then used by the atmosphere to calculate a new stress anomaly field that is used for the next period of ocean integration. The time step for this coupling was chosen to be 15 days. Because there is little variance in the SST at short timescales in this model, the interval of the coupling could be made shorter if desired.

Initial results with the coupled model

Because of limitations imposed by computer resources, the coupled model experiments described in this section represent an exploration of a very small portion of parameter space. Many of the quantities that are relevant to the behaviour of the coupled system, for instance the strength of the climatological upwelling, are not parameters that can be directly controlled but rather are internal variables that change in a complex manner as the specified elements of the model, such as climatological surface stress, are varied. It is thus difficult to perform 'clean' experiments in which only one element of the coupled interaction is changed at a time and, as a result, a certain amount of guesswork is involved in the design of each experiment. In view of this, the experiments are presented more or less in chronological order, the motivation for each change being clear from the previous experiment.

Each experiment is begun from the end of one of the climatological runs. An initial westerly wind stress anomaly is applied for one month, just as for the ocean-only experiment described previously. The coupled system continues to evolve freely after this stress is switched off. The differences between the runs mainly involve either changes in the ocean climatology or in the atmospheric model parameters.

Run I was begun from climatology I. Figure 3 shows the SST evolution over the nine years of this run. The behaviour is very different from that of the uncoupled ocean model. The changes in SST produced by the initial stress perturbation are maintained in a warm phase for several months by wind stresses produced by the anomalous SST. This coupled interaction evolves to a phase that is colder than climatology, followed by a second, though less-intense warm phase and cold phase, before the oscillation dies out to the climatological steady state. The coupled system thus appears to be undergoing a damped oscillation about the known stationary point, having a period of just over three years.

It seemed desirable to discover whether and under what conditions a sustained oscillation could be obtained. One of the notable features of climatology I is the overly extensive cold tongue along the Equator that is too cold relative to observations (this despite damping toward observed air temperatures in the surface heat budget). Because the strongest coupling tends to occur on the margin between the warmest SSTs and the region where the thermocline is shallow

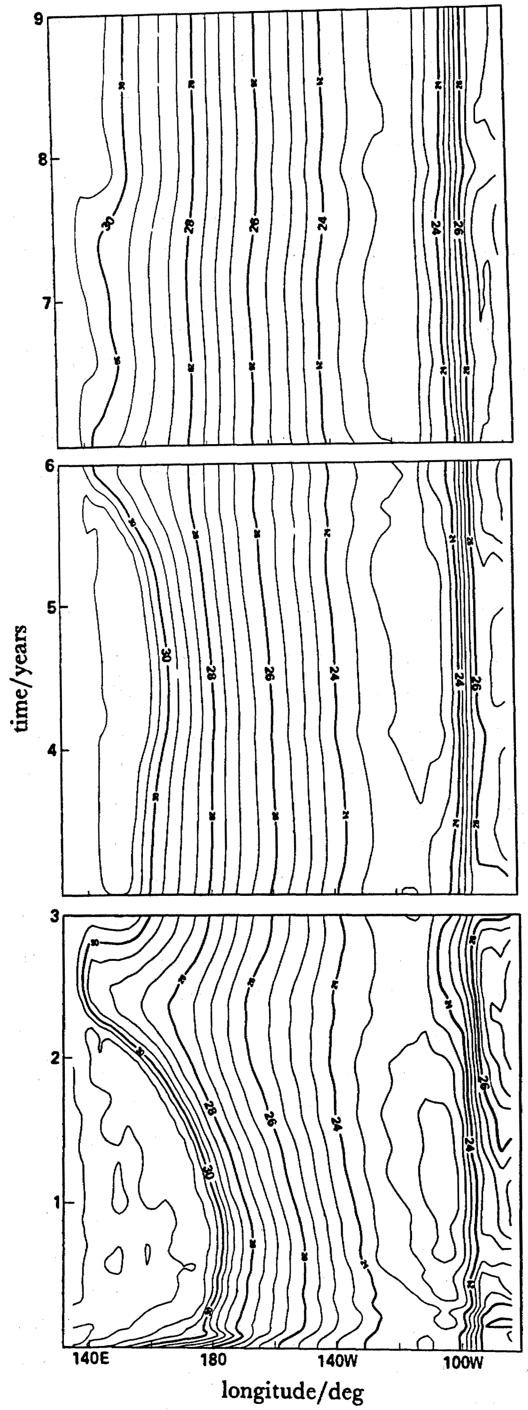

FIGURE 3. Evolution of sea surface temperature along the equator for coupled run I (see text). Contour interval 0.5 °C.

(i.e. the vertical temperature gradient is strong), a sharp horizontal temperature gradient at the western margin of the cold tongue will tend to restrict the length of the region in which coupled instabilities can grow. Furthermore, strong climatological upwelling tends to act as a damping on surface temperature perturbations. With this motivation, climatology II was produced, to test the effect on the oscillation of reducing the strength of the cold tongue.

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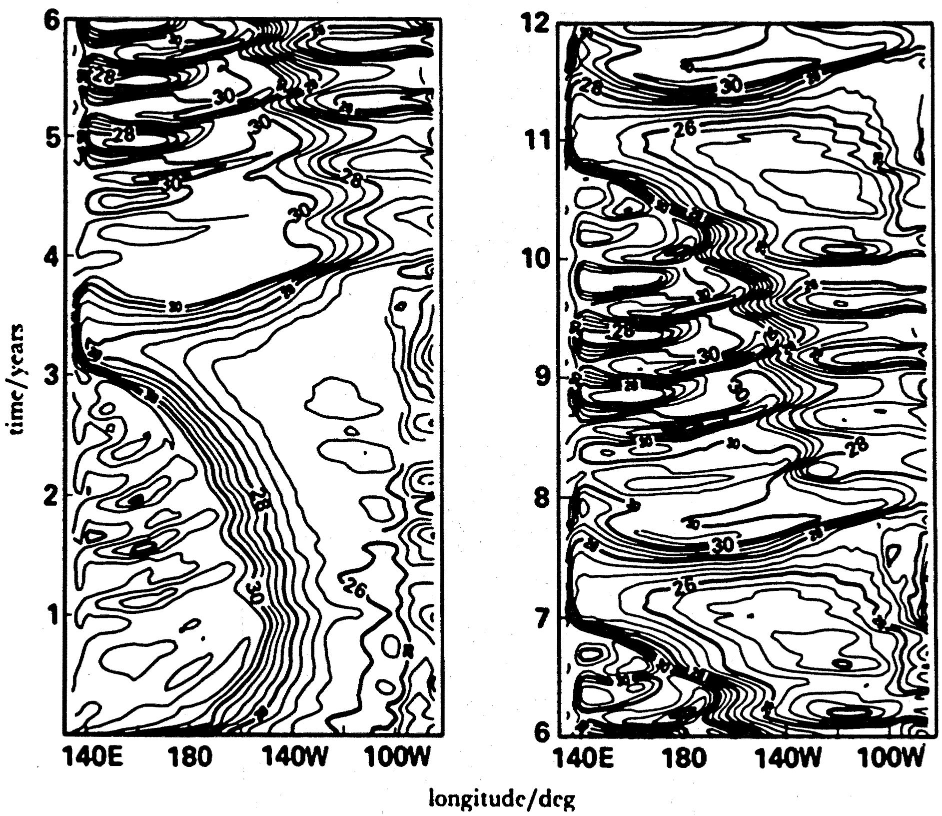

FIGURE 4. As for figure 3, but for coupled run II.

Run II was started with a similar initial stress perturbation from climatology II. The SST evolution along the Equator for the twelve years of this run is shown in figure 4. Sustained and very complex oscillations occur throughout the run and there is every reason to believe that these would continue if the model were run longer. Oscillations at two dominant periods may be distinguished. One has a period of just under four years and the pattern of its evolution strongly resembles that of the damped oscillation found in run I. The other has a period of about five or six months and appears most strongly during the warm phase of the long period oscillation. It seems reasonable to hypothesize that the system has undergone a Hopf bifurcation leading from a stable stationary point with the three to four year oscillation weakly damped, to a state where the stationary point is unstable to this oscillation. One may then anticipate that for values of the parameters such that the stationary point is only weakly unstable, a less complex cycle may occur.

The parameter that was varied between runs I and II was the factor multiplying the mean wind stress. Changes in this are computationally expensive, since a new climatology must be spun up each time. It is also difficult to interpret exactly how the 'strength of the coupling' is being affected. A much more convenient parameter may be obtained by multiplying the wind stress anomaly that results from a given departure of the oceanic state from climatology by a 'relative coupling coefficient'. Run III was performed with this parameter set to 0.8 (defining run II to have a value of unity) and was started just as for run II, from climatology II with an initial applied wind stress anomaly. The SST along the Equator for the nine years of this run is shown in figure 5. The five to six month oscillation has disappeared, and the long-period

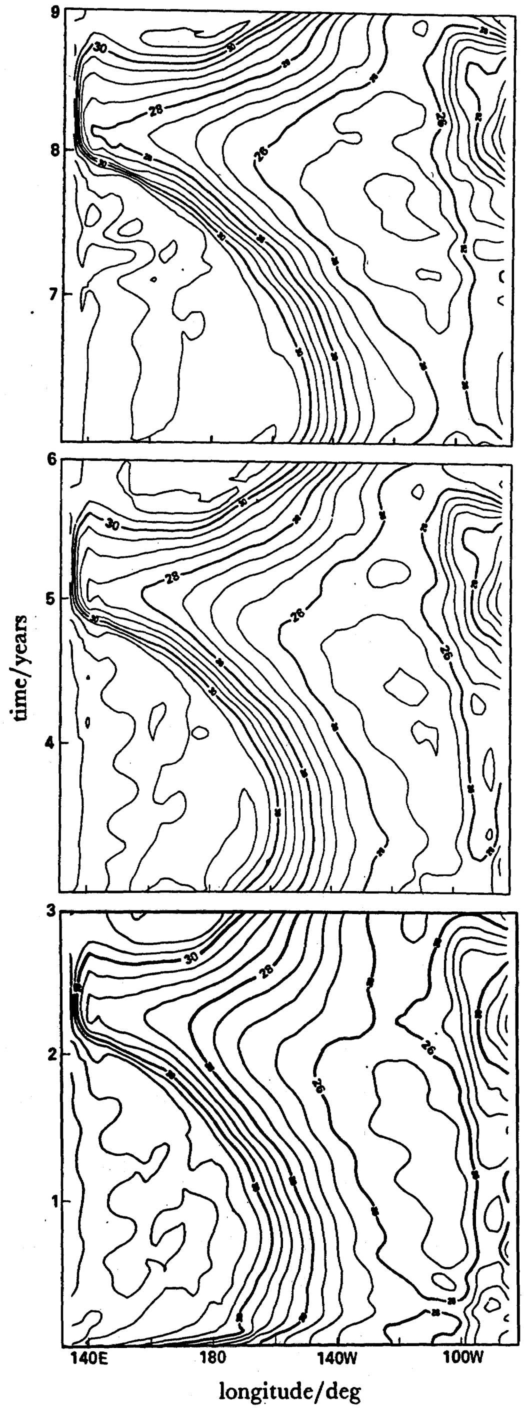

FIGURE 5. As for figure 3, but for coupled run III.

oscillation is very regular, with a period close to three years. An additional run was performed with a relative coupling coefficient of 0.6, and the long-period oscillations become damped, as in run I, so the stationary point is again stable.

To verify that the oscillations do indeed arise from an instability of the stationary point, an experiment was performed that repeated the first three years of run III but with the magnitude

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of the initial stress perturbation reduced to 0.01 dyn cm^{-2}, which for numerical purposes may be regarded as an 'infinitesimal' perturbation. The initial perturbation grew roughly exponentially, as expected, with an *e*-folding time of about three months.

This series of experiments constitutes reasonable numerical evidence of a Hopf bifurcation of the stationary climatology as parameters affecting the strength of the coupling are varied. For slightly supercritical values of, for instance, the 'relative coupling coefficient', the attractor appears to be a periodic limit cycle. As this parameter is increased further, the higher-frequency oscillation found in run II is very likely a secondary Hopf bifurcation, i.e. the limit cycle becomes unstable to a higher-frequency eigenmode. Whether subsequent bifurcations would lead to chaos as the coupling was increased is an open question. In any case, these preliminary results suggest that the internal variability of the coupled system can be complex.

It is worth examining the nature of the oscillation in both runs II and III in greater detail. A full analysis would include not only synoptic diagnostics of the space-time evolution of the oscillations and budget studies of the terms contributing, for instance, to the SST evolution, but also further modelling. Analysis of the mechanisms contributing to the linear instability of the stationary point should give more insight into the origin of the three to four year oscillation. Unfortunately, such detailed analysis requires considerable effort even in simpler models. Understanding of the oscillation mechanism in the Cane & Zebiak (1985) model, for instance, came three years after the model's first publication (Battisti & Hirst 1988). For the moment, all that may be offered is a description of the evolution and some hypotheses that will be further tested.

As run III is simpler, it will be described first. Figure 6 focuses on the last three years of the run, which represents almost exactly one cycle. The anomalies of SST, zonal current in the surface layer and vertical velocity just below the surface layer are shown. The anomalies are relative to the nine year mean of the run, rather than to the climatology. For some fields the evolution is easier to see in this format, because the mean of the cycle can be significantly different than climatology. The warm phase, which lasts through most of the sixth year, has a 2 °C warm anomaly located in the east central Pacific. This warm pool is confined to within about 5° of the equator. The warm water creates a westerly wind stress anomaly above and to the west of the warm pool. This stress drives anomalous downwelling and eastward surface currents, both of which act to maintain the warm pool. As the event progresses, the very weak easterly wind stress anomaly that occurs to the east of the warm anomaly begins to produce weak anomalous downwelling and westward surface currents to the east of the warm anomaly. Some combination of these produces anomalously cold water by horizontal and vertical temperature advection. The cold water, in turn, increases the easterly wind anomalies above and to the west of it. Both mechanisms can be shown potentially to contribute to westward-propagating unstable disturbances in a very simple shallow water-like model. The anomalies propagate westward, essentially extending the equatorial cold tongue across the basin until it begins to cut into the Indonesian warm pool, forcing the warm water off the Equator.

While the cold phase progresses westward along the Equator, the weak westerly stress anomaly to the east of the cold anomaly produces anomalous down-welling and westward currents. These anomalies, and the warming of SST that they produce, also progress westward. As they reach the dateline, however, the warming suddenly intensifies, and the onset of the warmest part of the warm phase propagates back eastward along the Equator.

This form of this coupled oscillation is not in perfect agreement with observed aspects of the

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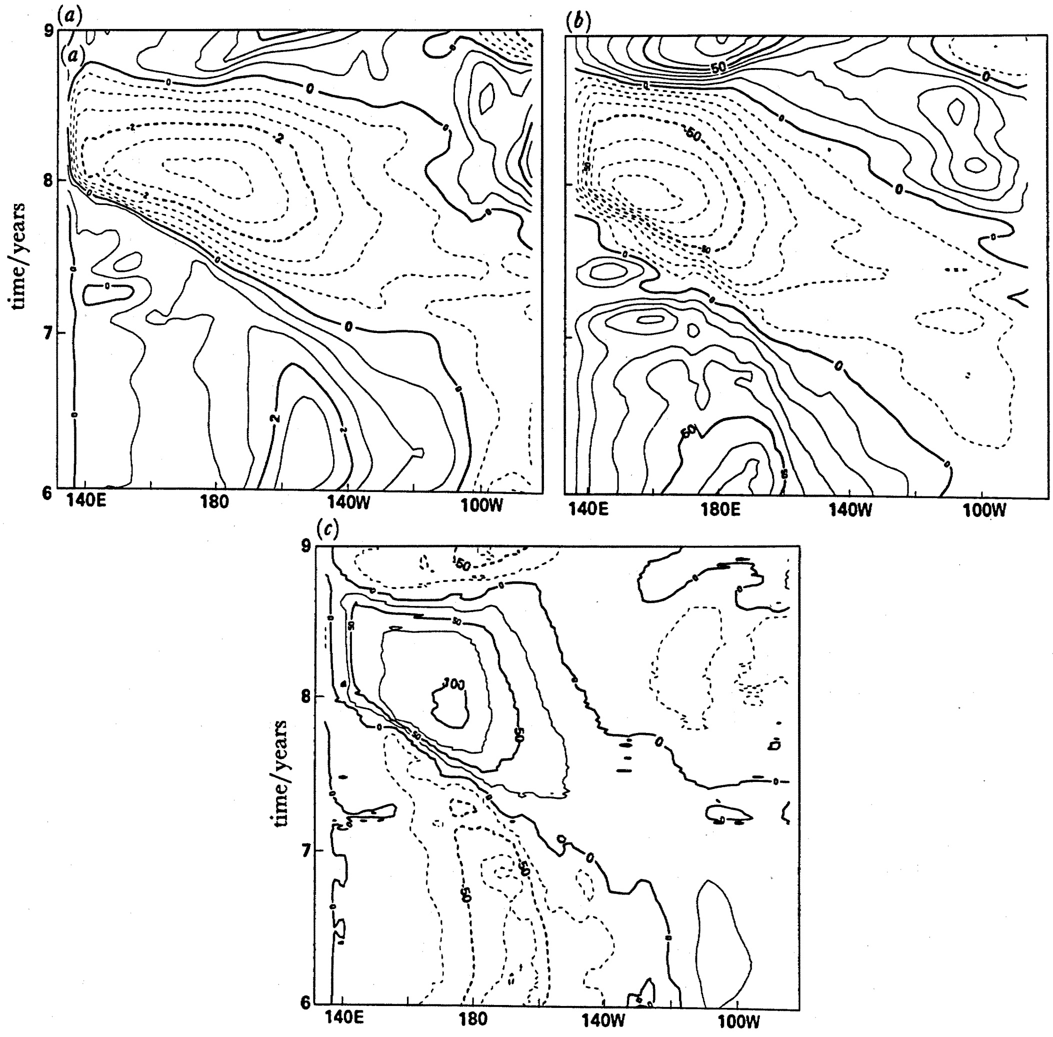

FIGURE 6. Evolution of anomalies along the equator during years 7–9 of coupled run III. (*a*) SST anomaly; Contour interval 0.5 °C. (*b*) Zonal surface current anomaly; contour interval 10 cm s^{-1}. (*c*) Oceanic vertical velocity anomaly at depth 20 m, contour interval 25 cm day^{-1}.

ENSO cycle. The observed cold phase never penetrates as far west as occurs in this experiment, and the observed warm phase propagates further east. The approach taken in these preliminary runs of the model is not to attempt to reproduce the ENSO cycle exactly, but rather to acquire some idea of the range of dynamics that can occur in even such an idealized coupled system. In the run III oscillation, both westward and eastward propagation of anomalies occurs. This has some correspondence to observations, in that the canonical El Niño of the Rasmussen & Carpenter (1982) composites has westward propagating warm anomalies, while the El Niño of 1982–83 is infamous for having propagated eastward, much as occurs during the onset of the strongest part of the warm phase in this experiment. Because this model appears to have both eastward and westward propagation mechanisms present, it may be possible to discover under what circumstances one becomes dominant over the other.

Perhaps equally important, the cycle of run III very noticeably fails to conform to the evolution that would be expected from the Schopf & Suarez (1988) delayed oscillator mechanism. The westward-propagating anomalies of SST, heat content, current and vertical velocity are all maximum at the Equator and do not correspond to the form of free equatorial Rossby waves. If anything, the westward-propagating aspects of the oscillation resemble more the infinite basin unstable modes examined by Hirst (1986) in a linear shallow water model. The oscillation of these unstable coupled modes does not depend on the existence of a western basin boundary as is the case for the delayed oscillator mechanism. Each cold phase is induced by the wind stress associated with the preceding warm phase and vice versa, as seems to occur in this model. When the infinite-basin modes are confined within a basin, they are little changed except that the zonal length scale is quantized (Hirst 1988).

Examining the SST, surface current and vertical velocity anomalies for a corresponding three year period of run II, the long-period oscillation follows much the same evolution as was found in run III (only SST is shown in figure 7). The cold phase and the onset of the warm phase propagate westward, but once the warm phase has begun the warmest anomalies propagate eastward. In fact the onset of the strongest part of the warm phase, in which 2–3 °C SST anomalies cover the eastern Pacific, nearly reaching the coast of 'South America', is associated with an event of the five to six month oscillation.

Occurrences of this oscillation are confined to the warm phase of the interannual oscillation. Strong surface jets propagate eastward from the western basin, carrying warm water across the Equator. These are coupled oscillations, accompanied by westerly wind stress and downwelling and SST anomalies and travelling slower than the oceanic first or second internal mode Kelvin wave speed. The sequence of successive short period peak warmings within the longer period warm phase seen in figure 7 is somewhat reminiscent of similar events found in the coupled GCM of Philander *et al.* (1984), although the period is shorter here. A qualitatively similar 'va-et-vient' of the warm phase at six-month intervals occurred during the 1986–87 El Niño with considerable negative impact on experimental forecasts (Barnett *et al.* 1988). Although these

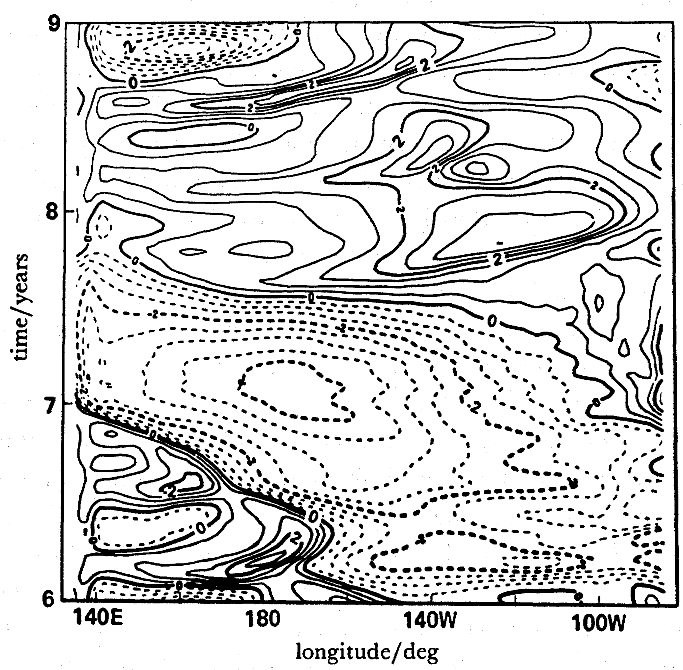

FIGURE 7. SST anomaly as for figure 6*a* but for run II.

events may have been caused by atmospheric perturbations, this figure does raise the question of whether variability of frequencies higher than that of ENSO may arise internally as coupled oscillations and affect the predictability of the system.

Comments on simple atmospheric models

Although it can be very useful for air–sea interaction studies to have a tropical atmospheric model that is free from temporal variability on purely atmospheric timescales, existing simple atmospheric models must be viewed with some caution. No consensus as yet exists even on the dominant mechanism responsible for the observed link between SST and atmospheric tropical convergence zones. The models of Zebiak (1986) and Davey & Gill (1988) follow a hypothesis that dates back to Bjerknes (1969), assuming that the convergence zones are linked to large values of the surface heat flux, increased in magnitude by a 'convergence feedback'. N.H. point out that this would lead to a disastrous simulation of the climatological convergence zones if the surface heat flux is correctly modelled and propose that the positions of the convergence zones are more strongly dependent on minima in a vertically integrated measure of the moist static stability.

Still another process is considered by Lindzen & Nigam (1987; hereafter L.N.). Temperature in the moist boundary layer below the Trade Inversion is assumed to be strongly mixed toward SST, an assumption shared by N.H. L.N. point out that the resulting horizontal temperature gradients imply significant horizontal pressure gradients. In a first version of their model, they assume that all the boundary layer convergence forced by these gradients would be perfectly taken up by the cumulus mass flux, so that vertical motions would never feel the effects of stratification. This assumption leads to unrealistically strong convergence/divergence near the Equator. L.N. replaced this with the assumption that the cumulus mass flux takes up the boundary layer convergence with a specified lag time, which is assumed to result in an increase in the height of the well-mixed layer. The effect of this is to produce a negative feedback on the convergence. Although this view of the boundary layer is open to some question, especially if reduced gravity effects are included, alternative derivations of qualitatively similar negative feedbacks are possible. Basically, if the vertical motion must do work against stratification during any part of its cycle, a negative feedback on vertical motion will arise. This is implicitly related to the claim made by N.H. that if the mean circulation is thermodynamically direct then the gross moist stability must be slightly positive.

In the L.N. model, the SST forcing appears in the momentum equations through the pressure gradient term, whereas the divergence equation has the form of a damped shallow water height equation. The damping time required to obtain a circulation of reasonable magnitude is extremely short – about 30 min – and the depth scale is the depth of the Trade cumulus boundary layer, about 3000 m. It may be shown (Neelin 1989) that by a suitable transformation of variables the SST forcing may be moved from the momentum equations to the height equation, and the depth and damping time rescaled to make the L.N. model formally equivalent to the Gill (1980) model with heating proportional to SST. As a result, the response of the L.N. model to an SST anomaly will be very similar to that of the N.H. model (and to other models if the surface flux is assumed proportional to SST) to a first approximation. The differences will only appear when nonlinearity becomes important. The overall conclusion to be drawn from this is that the existing simple atmospheric models should be regarded with caution. At least three different mechanisms for relating boundary layer convergence to SST have been proposed, all of which give similar results to first order for SST anomalies. As long as a first approximation is all that is required for coupled modelling, any of the models may be satisfactory. However, if one wishes to improve upon this, for instance if the nature of the nonlinearity of the response to SST proves to be important to coupled oscillations, it will then be necessary to discover which, if any, of the models is closest to the truth.

A model that could potentially distinguish among these mechanisms would have to treat a number of fairly subtle physical processes. It would require a cumulus parametrization in which the interaction between the cumulus mass flux and the vertical profiles of temperature and moisture is well treated. Models which specify the vertical structure of these quantities *a priori* will not be able to determine for or against either the N.H. or the L.N. mechanisms. If the model is to treat the L.N. mechanism, it would require a Trade cumulus boundary layer whose height is a predicted quantity and in which the radiative, sensible and moisture transports from the adiabatic boundary layer below are consistently treated.

Conclusions

Preliminary experiments with an ocean GCM-steady atmosphere, 'intermediate' coupled model indicate that a Hopf bifurcation leads from a steady state to a limit cycle of three to four year period as certain parameters affecting the coupling are varied. Relevant quantities include the vertical gradient in the thermocline, the strength of the mean upwelling, and the magnitude of the nonlinear atmospheric response to oceanic anomalies. The El Niño-period oscillation contains both eastward and westward propagation mechanisms, which dominate at different parts of the cycle. This suggests that more than one of the many proposed mechanisms for producing coupled instabilities may be operating. This multiplicity of coupling mechanisms may account for the heterogeneity in observed El Niños.

For stronger coupling, a further bifurcation of the system yields coupled oscillations of about six-month period that coexist with the ENSO-period oscillation and considerably complicate the cycle. These oscillations appear to be related to coupled Kelvin waves and produce sudden warmings and coolings within El Niño warm phase. Although not chaotic, the system in this part of parameter space would be more difficult to predict in a practical situation where problems of initialization with sparse data would have to be faced. Whether inherently irregular vacillations will occur in any realistic parameter régime remains to be seen.

The ENSO-period oscillation in these initial experiments appears to be governed by a different mechanism than is found in the coupled two-level models of Cane & Zebiak (1985), Schopf & Suarez (1988) and Battisti (1988). It is possible that the difference might lie in the greater nonlinearity of the ocean. In this model, the vertical temperature gradients on the Equator that support the growth of coupled instabilities are themselves strongly affected by the coupling. In an anomaly model, an unstable coupled mode ceases to grow once it propagates out of the region where the specified mean conditions support its growth. In this model, a propagating anomaly can change the mean conditions. For instance, a westward-propagating cold phase extends the region of shallow thermocline with it as it moves westward, instead of entering a region of specified deep thermocline. On the other hand, there are also significant differences in the nonlinearity of the response to SST in this atmospheric model, which may change the region where the strongest coupling occurs relative to previous models. If this proves important, a considerable effort will be required to determine which, if any, of the proposed atmospheric processes is dominant. Previous atmospheric models assume convection depends on the surface heat flux, a process which is of secondary importance in both the current Neelin & Held (1987) model and in another proposed simple atmospheric model by Lindzen & Nigam (1987). Because the atmospheric boundary layer tends to come into balance with the sea surface, the steady-state values of the fluxes become unimportant compared to the horizontal gradients of temperature and moisture which this balance produces in the boundary layer. To a first approximation, these models tend to give similar predictions regarding observable quantities, suggesting that further modelling will be required to distinguish between them.

Despite these obstacles, the extreme complexity of the phenomena found in coupled GCMS, in which coupled modes of variability are continually perturbed by high-frequency atmospheric events, provides motivation for the continued development of intermediate coupled models in which the atmospheric component is steady. In the coupled GCM of Philander *et al.* (this symposium), for instance, it would be extremely difficult to isolate the coupled modes of variability from the response of the ocean to uncoupled atmospheric variability without using this hierarchical modelling approach. The apparent multiplicity of mechanisms that can give rise to self-sustaining coupled oscillations and the occurrence of coupled oscillations on multiple timescales in the intermediate model suggest that even the basic nature of the El Niño oscillation is still an open question.

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