A THEORY FOR INTERDECADAL CLIMATE FLUCTUATIONS

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The unexpected and prolonged persistence of warm conditions over the tropical Pacific during the early 1990's should be viewed, not as a prolonged El Niño, but as part of a decadal climate fluctuation that is governed by different physical processes. Whereas interannual fluctuations, including El Niño, amount primarily to a horizontal redistribution of warm surface waters within the tropics, interdecadal climate fluctuations involve changes in the properties of the equatorial thermocline because of an influx of water from higher latitudes. The influx affects equatorial sea surface temperatures and hence the tropical and extra-tropical winds that in turn affect the influx. Such processes can give rise to continual interdecadal oscillations.

Interactions between the ocean and atmosphere contribute to climate fluctuations over a broad spectrum of time-sales, from seasons to decades and longer. Studies of those interactions have thus far focused on the Southern Oscillation, which has its principal signature in the tropical Pacific sector, and which has a period of three to four years. Superimposed on this natural mode of the coupled ocean-atmosphere system, are lower frequency interdecadal fluctuations that contribute to the irregularity of the Southern Oscillation (Trenberth and Hurrell, 1994). The recent persistence of unusually warm conditions over the tropical Pacific during the early 1990's is an example (Latif et al 1996). In spite of theories and models that explain and simulate the Southern Oscillation (Neelin et al, 1995, Philander, 1990), and that correctly predicted the occurrence of its warm phase, El Niño, in 1987 and 1991 (Latif et al, 1994), the persistence of that recent warming came as a surprise (Ji et al, 1996). At present it has no explanation.

The Southern Oscillation, between complementary El Niño and La Nina states, involves an east-west redistribution of warm surface waters so that, during El Niño, the thermocline deepens in the eastern tropical Pacific while it shoals in the west. To a first approximation, neither the mean depth of the tropical thermocline nor the temperature difference across the thermocline changes. Many coupled ocean-atmosphere models of the Southern Oscillation exploit this result by having an ocean that is composed of two immiscible layers, a warm upper and cold deeper layer, that are separated by a thermocline whose mean depth is specified. During the persistent warming of the tropical Pacific in the early 1990's, isotherms in the equatorial thermocline deepened in the east but there is no evidence of a compensatory shoaling in the west[7]. Figure 1, which shows the measured change in the thermal structure along 110 W, just west of the Galapagos Islands, indicates that the warming was more pronounced at depth than at the surface. We next explore the implications of assuming that this warming was associated with an influx of warmer waters from the extra-tropics. because of an earlier change in the prevailing westerly winds in the higher latitudes. The influence of the extra-tropical winds extends through certain windows to the oceanic interior. The approximate location of the windows can be determined by following surfaces of constant density as they rise from the depth of the tropical thermocline to the ocean surface in the extratropics. If a change in atmospheric conditions causes the winds in the latter region to pump downward unusually warm water, then it is possible that, in due course, temperatures in the equatorial thermocline will rise. Liu and Philander (1994) and Liu et al, (1994), building on earlier results of Luyten et al, (1983), used an ocean model to locate the extratropical windows to the tropical thermocline, and to trace the routes that water parcels follow. The subtropical regions where the surface winds drive convergent Ekman flow are potential windows to the deeper ocean, but not necessarily to the tropics. Water parcels that are forced downwards in the western side of an ocean basin join the subtropical gyres that include intense western boundary currents such as the Gulf Stream and Kuroshio Current. Water parcels that are forced downward in the central and eastern parts of the subtropical ocean basins are likely to travel westward and equatorward, to join the Equatorial Undercurrent that carries them eastward along the equator. Equatorial upwelling transfers these parcels to the surface layers whereafter poleward Ekman drift returns them to regions of subduction in the extratropics. Figure 2 shows data that tentatively confirm the initial part of this journey. These results are from a study by Deser et al, (1996) of an interdecadal climate fluctuation that included a cooling of the surface waters in the western and central northern Pacific Ocean during the period 1976 to 1988. The reference thermal state is the time-average of temperatures as measured since 1900. The anomalous temperatures in °C are departures from the reference temperatures, during three different periods: 1977 to 1981, 1982 to 1986,

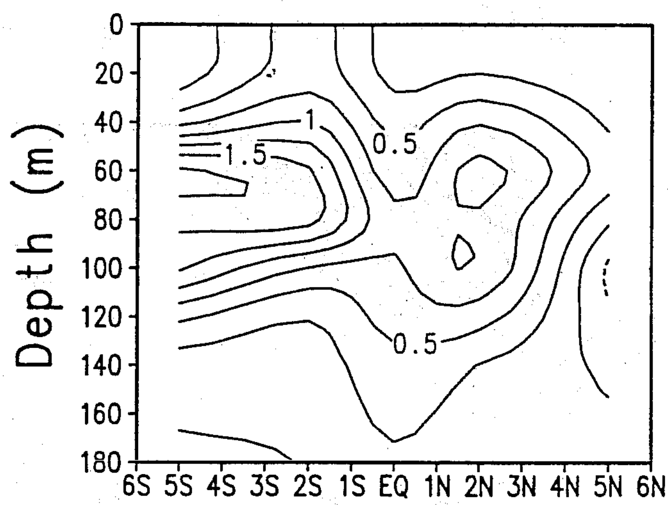

Figure 1: The change in temperature (in degrees centigrade) as a function of depth and latitude along 110°W, obtained by subtracting the mean temperature for the period 1985 to 1087 from that for the period 1990 to 1992, using data from Tropical Atmosphere Ocean (TAO) Array (14). Eq, equator.

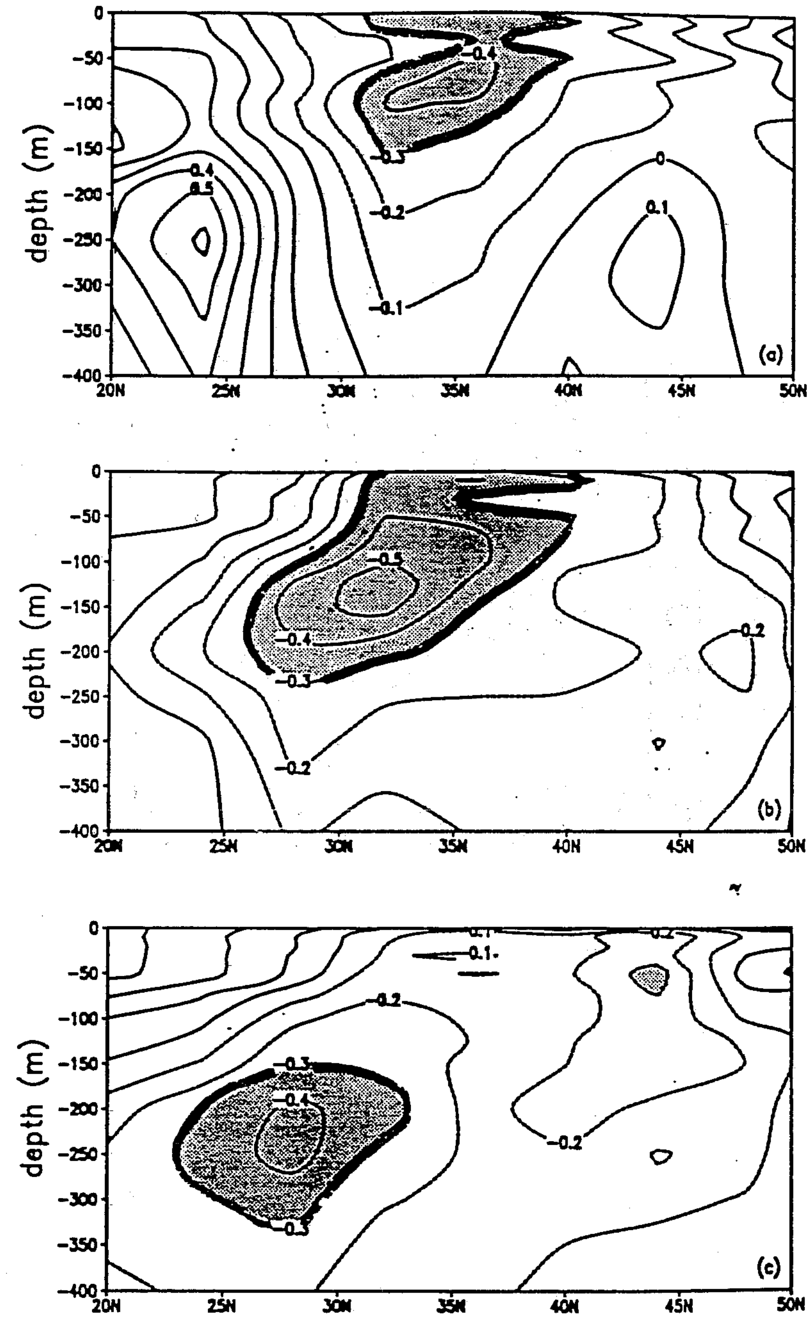

Figure 2: The equatorward and downward propagation of anomalous temperatures, averaged over longitudes 170W to 145W, during three periods: 1977 to 1981, 1982 to 1986, and 1987 to 1991. (After Deser et al. 1996.)

and 1987 to 1991. Unusually cold water is seen to move downward and equatorward along the surface of constant density. The data are from the central Pacific -- they are averages for the longitudes 170°W to 145°W -- so that it is possible that some of the water subsequently joined the subtropical gyre while some continued equatorward. Tracer (tritium) data[8] indicate that surface waters from the extra-tropics do indeed reach the Equatorial Undercurrent but further data analyses are necessary to determine the paths followed by water parcels. The results of Deser et al indicate that, and theoretical studies assume that the water parcels move along surfaces of constant density even though their temperatures are altered. This is most likely if a cooling is accompanied by a decrease in salinities, a warming by an increase. (Stronger westerly winds that cause more evaporation and lower temperatures are usually accompanied by heavier rainfall that decreases salinity.)

Changes in extra-tropical winds can result in a subsequent changes in the structure of the equatorial thermocline, of the type seen in figure 1. Equatorial upwelling can translate a warming of the equatorial thermocline into a warming of the surface waters whereafter the unstable ocean-atmosphere interactions mentioned earlier amplify the warming. Tropical processes that establish the Southern Oscillation can reverse this trend but let us for the moment disregard those processes and instead turn to the extra-tropics for a mechanism that can also halt the warming trend in low latitudes. This mechanism depends on the link between atmospheric variability in the tropics and extra-tropics. On both interannual and interdecadal time-scales, the appearance of westerly winds (or relaxed easterly winds) over the western tropical Pacific, usually during periods when unusually warm surface waters cover large parts of the central and eastern tropical Pacific, are associated with an intensification and equatorward shift of the Jet Stream, an eastward and equatorward extension of storm tracks across the Pacific, and with an intensification of the Aleutian low pressure system (Trenberth and Hurrell, 1994, Lau and Nath, 1994). A warming in the tropics is usually associated with more intense westerlies and colder surface waters (because of evaporation) in an extra-tropical region that happens to be a window to the equatorial thermocline. The cold water pumped downwards there in due course arrives in the tropical thermocline, halts the warming and sets the stage for cold conditions in the tropics. These arguments imply a continual, interdecadal climate fluctuation with a period that depends on the time it takes water parcels to travel from the extra-tropics to the equator. (See Latif and Barnett, (1994) for a discussion of decadal variability that does not involve the tropics.) The arguments presented here can be quantified by means of the following idealized model that intentionally suppresses interannual variations in order to focus on the interdecadal variations. The oceanic component of the model, shown in figure 3, consists of two tropical boxes, one at the surface at temperature T_2, the other immediately below it in the thermocline at temperature T_3, plus an extra-tropical surface box at temperature T_1. The tropical box

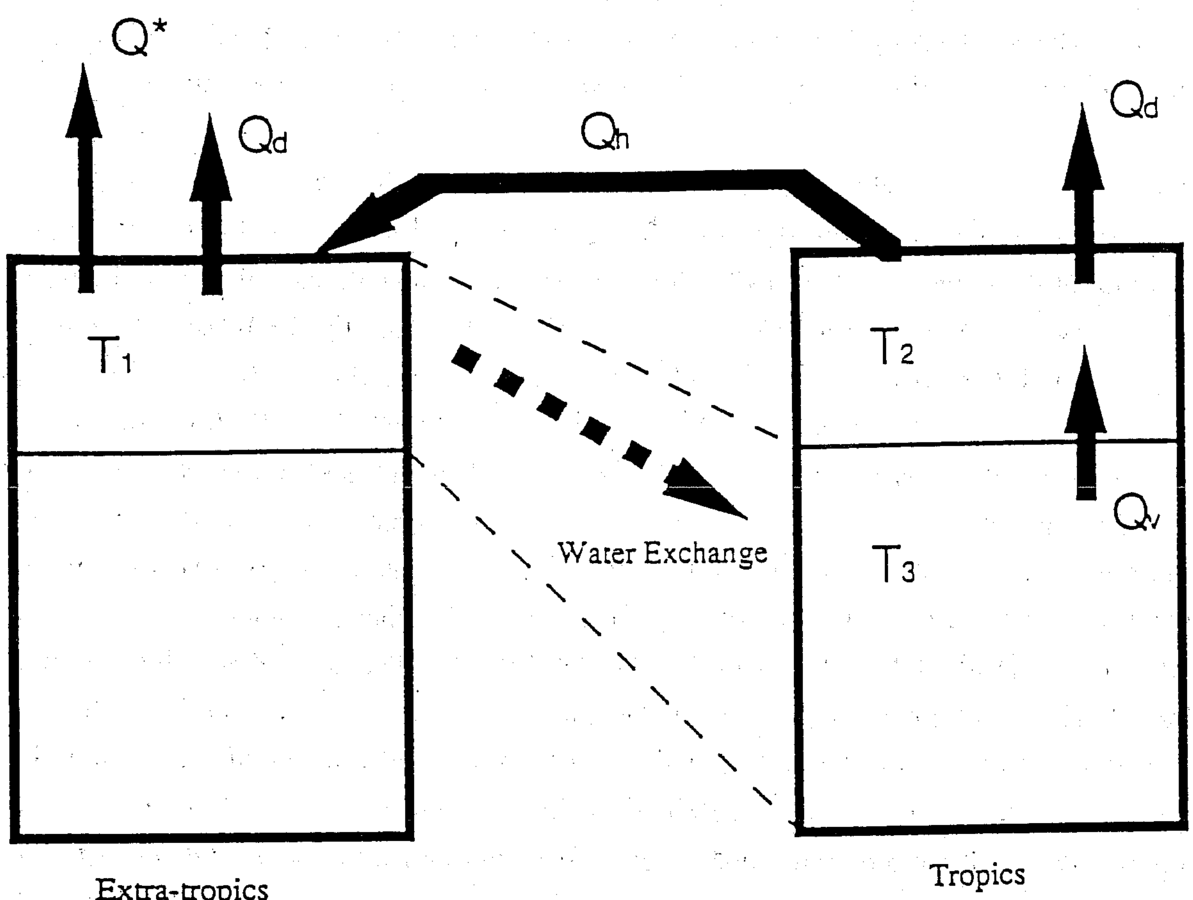

Figure 3 : Sketch of the ocean box-model.

covers the approximate region 20S to 20N, the extra-tropical box the region 25N to 50N (or 25S to 50S). The temperature of the surface box at the equator is determined by heat fluxes:

$$\frac{dT_2}{dT} = -\ Q_h + Q_v -- Q_d \qquad (1)$$

Equatorial upwelling effects a vertical transport Q_V that depends on the vertical temperature difference, and on a vertical velocity component. That vertical flow measures equatorial upwelling in response to zonal winds that drive divergent surface currents. The intensity of those wind, as argued earlier, depends on the temperature difference between the western and eastern equatorial Pacific. Temperature variations are far more modest in the west than in the east so that variations in the intensity of the wind, and hence in the intensity of upwelling, can be taken to be proportional to T_2.
Hence Q_V is proportional to $T_2(T_2 - T_3)$. If we assume all temperatures to be composed of a time-averaged value and a perturbation, and if we linearize, then this term has two components; one is proportional to the perturbation temperature difference $(T_2 - T_3)$, the other is proportional to the perturbation temperature T_2 and represents the positive feedback in which a change in temperature in the eastern equatorial Pacific intensifies the winds which in turn reinforces the change in temperature. Q_h is the poleward atmospheric transport of heat out of the box and is

assumed to depend on the temperature difference $(T_2 -- T_1)$. The first term of its Taylor expansion can be written as $\gamma(T_2 - T_1)$, where γ is a constant. The term Qd in equation (1) represents damping that we take to be proportional to the cube of the perturbation temperature whose equation can therefore be written

$$\frac{dT_2}{dt} = -\gamma\ (T_2 - T_1) - \delta\ (T_2 - T_1)\ +\ \lambda_2 T_2 - \varepsilon T_2^3 \qquad (2)$$

where all the temperatures now refer to perturbation temperatures and δ, λ_2, and ε are positive constants. The terms on the right hand side represent respectively poleward heat transport, equatorial upwelling, the positive feedback term involving the zonal winds, that also stems from upwelling, and damping.

Similar arguments for the perturbation temperature of the extra-tropical box yield the equation

$$\frac{dT_1}{dt} = \alpha\gamma(T_2 - T_1) + \lambda_1 T_2 - \varepsilon T_1 + Q^* \qquad (3)$$

The terms on the right hand side represent respectively the fraction of the poleward atmospheric transport of heat that remains in the extra-tropical box, the effect of local winds that in turn depend on changes in sea surface temperatures in the tropics, damping, and stochastic forcing from weather systems unrelated to tropical temperature variations.

To link the extra-tropical and tropical oceans, we assume that at any time t, subsurface temperatures at the equator are the same as surface temperatures in the extra-tropics at an earlier time:

$$T_3(t) = T_1(t - d) \qquad (4)$$

Functional (or delay) differential equations such as (2, 3 and 4) have been studied extensively and are known to have unstable (growing) oscillatory solutions (Gyori and Ladas, 1991). The presence of damping terms in our equations ensures bounded solutions so that we focus on oscillatory solutions and their sensitivity to changes in the parameters. Our reference case corresponds to the following numerical values for the constants. The time-scale for poleward heat transport, $1/\gamma$, is a year; the time-scale associated with the positive feedback in the tropics, $1/\lambda_2$, is approximately 100 days, and that associated with upwelling, $1/\delta$, is 60 days. The negative constant γ_1 is assigned a value equal to half of λ_2. The delay time d for the connection between the tropics and extra-tropics is taken to be 20 years.

The results in figure 4(a) show how an initial small perturbation in the surface layers at the equator slowly amplifies while generating extratropical temperature fluctuations before settling down to an oscillation with a period of 45 years and an amplitude of about 1°C. The tropical and extratropical fluctuations are out of phase. A cycle starts with a rapid increase in equatorial temperatures (because of the local positive feedback).The developments in the tropics cause an intensification of the extratropical westerly winds, and hence enhanced evaporation and a drop in surface temperatures of the extratropics. The latitudinal temperature difference created in this manner gives rise to a poleward transport of heat that halts both the warming of the tropics and the cooling of the extratropics, thus establishing an equilibrium state that persists for a considerable time before coming to an end when the cool

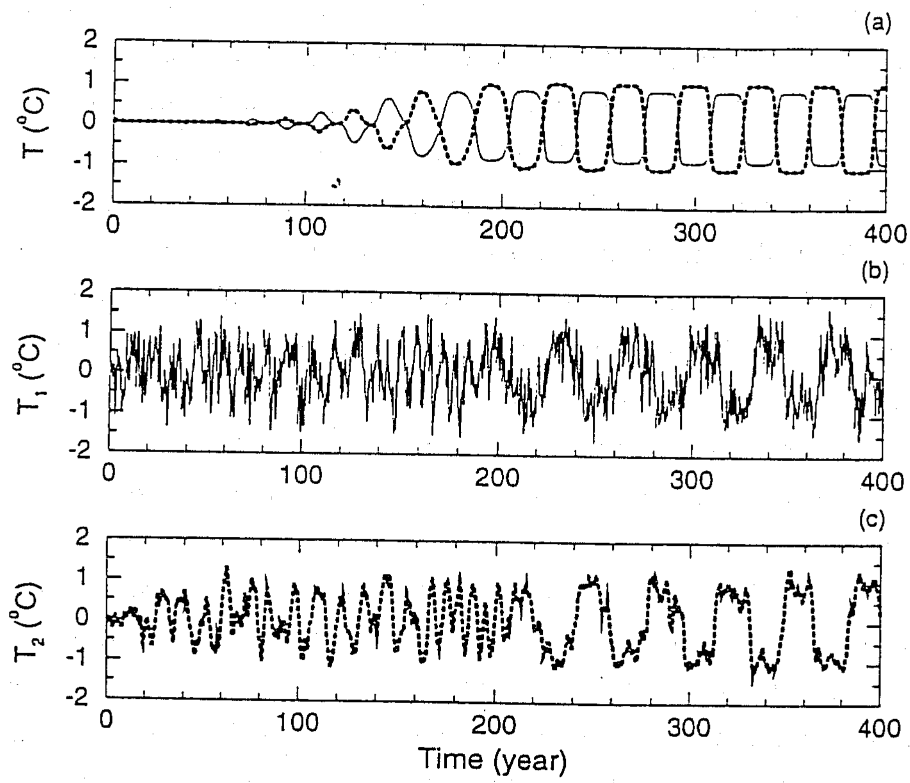

Figure 4 : (a) The interdecadal oscillations obtained by solving equations (2)-(4) for the case of no random forcing ($Q^* = 0$), and with parameters assigned their reference values given in the text. In (b) the random white noise forcing has a normal distribution, a zero mean value, and a rms of 2.

conditions in the extratropics affect first the equatorial thermocline and then the surface layer at the equator. The decrease in equatorial surface temperatures influences extratropical winds in such a manner as to increase extratropical temperatures. The resultant effect on the poleward heat transport in due course leads to the complementary phase of the interdecadal oscillation. In figure 4(a) the oscillation is perfectly periodic and transitions from one phase to the other are very abrupt. The introduction of stochastic forcing in the extratropics, in panel (b), causes the transition to be more gradual and the oscillation to be more irregular. It also leads to the presence of oscillations with a relatively short period. Such bifurcation to oscillations with different periods also occur when the values of certain parameters are altered. A change in the parameter "γ" which determines the rate at which heat is transported poleward has no effect on the period of the oscillation which is seen to depend linearly on the delay time "d". The parameter "γ" does affect the amplitude of the oscillation because the magnitude of the temperature difference between the tropics and extratropics depends on the rate of poleward heat transport. Changes in the feedback parameters "λ_1" and "λ_2" alter the period of the oscillation. The results presented here serve two purposes. The one is to demonstrate that continual, interdecadal climate fluctuations with a time-scale that depends on the time it takes extratropical atmospheric disturbances to influence the equatorial thermocline, are indeed possible. The other is to motivate observational and theoretical studies that explore the validity of the proposed mechanisms. The results in figure 1 are suggestive of equatorward propagation but it is conceivable that the path of the water parcels subsequently curved poleward, towards the Kuroshio Current. Liu and Philander (1994) calculated the trajectories of parcels that proceed equatorward but did so for idealized oceanic circulation driven by idealized winds. The calculations need to be repeated with realistic winds and special attention needs to be paid to asymmetries, relative to the equator, of windows from the extratropics to the equatorial thermocline. The interactions between the ocean and atmosphere need to be explored with models that have greater realism.

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