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"A Theory for Interdecadal Climate Fluctuations"

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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 Nino, but as part of a decadal climate fluctuation that is governed by different physical processes. Whereas interannual fluctuations, including El Nino, 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[1]. The recent persistence of unusually warm conditions over the tropical Pacific during the early 1990's is an example[2]. In spite of theories and models that explain and simulate the Southern Oscillation[3,4], and that correctly predicted the occurrence of its warm phase, El Nino, in 1987 and 1991[5], the persistence of that recent warming came as a surprise[6]. At present it has no explanation.

The Southern Oscillation, between complementary El Nino and La Nina states, involves an east-west redistribution of warm surface waters so that, during El Nino, 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. These models are unable to cope with a phenomenon that involves changes in the thermal structure of the tropical oceans, for example an influx of water from higher latitudes in the equatorial thermocline. There are indications that the persistent warmi of the 1990's was associated with such an influx.

In the eastern equatorial Pacific a shoaling of subsurface isotherms signals t end of El Nino and the return of colder surface waters. That is how El Nino 1987 terminated. During the persistent warming of the tropical Pacific in th early 1990's, there was again a shoaling of isotherms after 1992 but this time had no surface manifestation. This happened because the thermocline wa deeper in 1992 than it had been in 1987. Figure 1 shows how much warme the subsurface layers in the east were during the period 1990 to 1992, than th period 1985 to 1987. There is no evidence that this warming in the east wa associated with a compensatory shoaling of the thermocline in the west.[7] We next explore the implications of assuming that the warming wa associated with an influx of warmer waters from the extra-tropics because o 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[8] and Liu et al[9], building on earlier results of Luyten et al[10], 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 subtropcal 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[11] 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, 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 140°W -- so that it is possible that some of the water subsequently joined the subtropical gyre while some continued equatorward. Tracer (tritium) data[12] 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[11] indicate that water parcels move along surfaces of constant density even though their temperatures are altered. A decrease in temperature would have to be accompanied by a decrease in salinity for the density to remain unchanged. (This means that stronger westerly winds that cause more evaporation and lower temperatures would have to be accompanied by heavier rainfall that decreases salinity. That is usually the case.)

Changes in extra-tropical winds can result in subsequent changes in the structure of the equatorial thermocline, of the type seen in figure 1. Next, because of equatorial upwelling, sea surface temperatures are influenced and they in turn affect the surface winds. The positive feedback that now comes into play -- the winds influence the surface temperatures which in turn influence the winds -- characterizes interactions between the atmosphere and ocean. It can amplify an initial change, causing a modest warming of the surface waters near the equator to develop into El Nino, for example. After a while, the amplification comes to a halt, and is reversed, because of a key difference between the ocean and atmosphere: whereas the atmosphere responds rapidly to a change in sea surface temperatures, the ocean, because of its greater inertia, has a delayed response to changes in the winds. High sea surface temperatures in the eastern tropical Pacific during El Nino can not persist indefinitely because the delayed oceanic response, in the form of subsurface waves that travel across the Pacific, in due course leads to an elevation of isotherms in the east, a lowering of sea surface temperatures and the termination of El Nino. In addition to this particular delayed response, which, as mentioned earlier, involves an east-west redistribution of surface waters in the tropics, there is another one that involves links between the tropics and extra-tropics. Its oceanic part has already been discussed in the previous paragraphs so that we next turn to 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, and with an eastward and equatorward extension of storm tracks across the Pacific[1,13]. The more intense extra-tropical westerlies result in colder surface waters (because of evaporation) in an extra-tropical region that happens to be a window to the equatorial thermocline[13]. The cold water pumped downwards in that region in due course arrives in the tropical thermocline, halts the warming, and sets the stage for cold conditions in the tropics. The cool surface waters in low latitudes now affect the extratropical winds in such a way as to increase sea surface temperatures in the region where surface waters subduct[13]. 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[14] 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 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 Q_d in equation (1) represents damping that we take to be proportional to the cube of the perturbation temperature whose equation can therefore be written

$$\frac{dT_2}{dt} = -\gamma(T_2 - T_1) - \delta[T_2 - T_3] + \lambda_2 T_2 - \varepsilon T_2^3 \qquad (2)$$

where all the temperatures now refer to perturbation temperatures and γ, δ, 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

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

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[15]. 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 constant λ_1 is negative because of the negative correlation between the anomalous heat flux into the extratropical ocean and tropical temperature[13]; it 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 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. The period of the oscillation is determined mainly by the delay time d , which depends on the time it takes parcels to travel from the surface in the extra-tropics to the equator. In reality no single value can be assigned to d , because surface water subduct over a wide range of latitudes, and because there are various routes parcel can follow to reach the equator. For example, those that approach from the south can have a direct routes but those from north, because North Equatorial Countercurrent, may have to travel far to the west before they proceed equatorward. (Tracer data indicate that they can reach the equator from the North Pacific[12].)

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 as is evident in figure 5.

The sensitivity of the results to changes in the values of various parameters is displayed in figure 5. 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 as shown in figure 5(c).

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 2 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[8] 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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Figure Captions

Figure 1 : The change in temperatures as a function of depth and latitude along 110°W, obtained by subtracting the mean temperature for the period 1985 to 1987 from that for the period 1990 to 1992. (TAO data provided by M. McPhaden.)

Figure 2 : The equatorward and downward propagation of anomalous temperatures, averaged over longitudes 170W to 145W, during three periods: 1977 to 1981 (a), 1982 to 1986 (b) , and 1987 to 1991 (c). (Courtesy of C. Deser)

Figure 3 : Sketch of the ocean box-model.

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. The solid lines are for T_1 (extratropics) and the dotted lines are for T_2 (tropics).

Figure 5 : Sensitivity of the solutions to the changes in the values of poleward atmospheric heat transport γ and in the time d it takes the extratropical winds to affect the tropical thermocline. In (a) there is no detectable change in the linear dependence of the period on the parameter d as the value of γ changes from 0.5 to 1.0 to 2.0 (solid line, heavy dashed line, and light dashed line respectively). In (b) these changes are seen to affect the amplitude of the oscillation. In (c) it is evident that changes in the values of the feedback parameters λ_1 and λ_2 (which have the values 3 in the case of the solid line, 4 in the case of the heavy dashed line, and 5 in the case of the light dashed line) can change the period of the oscillation.

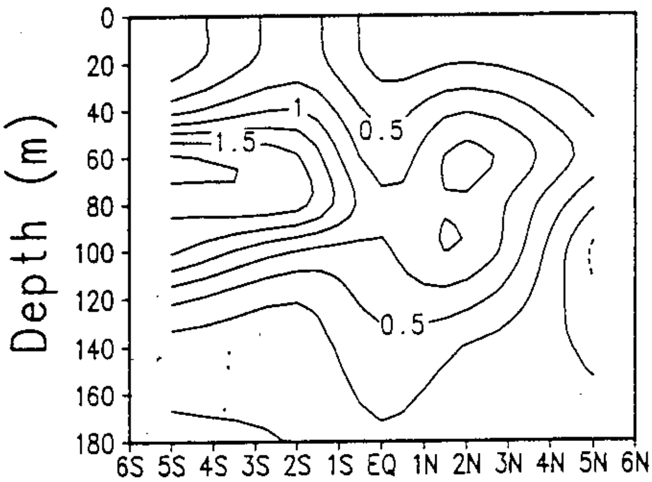

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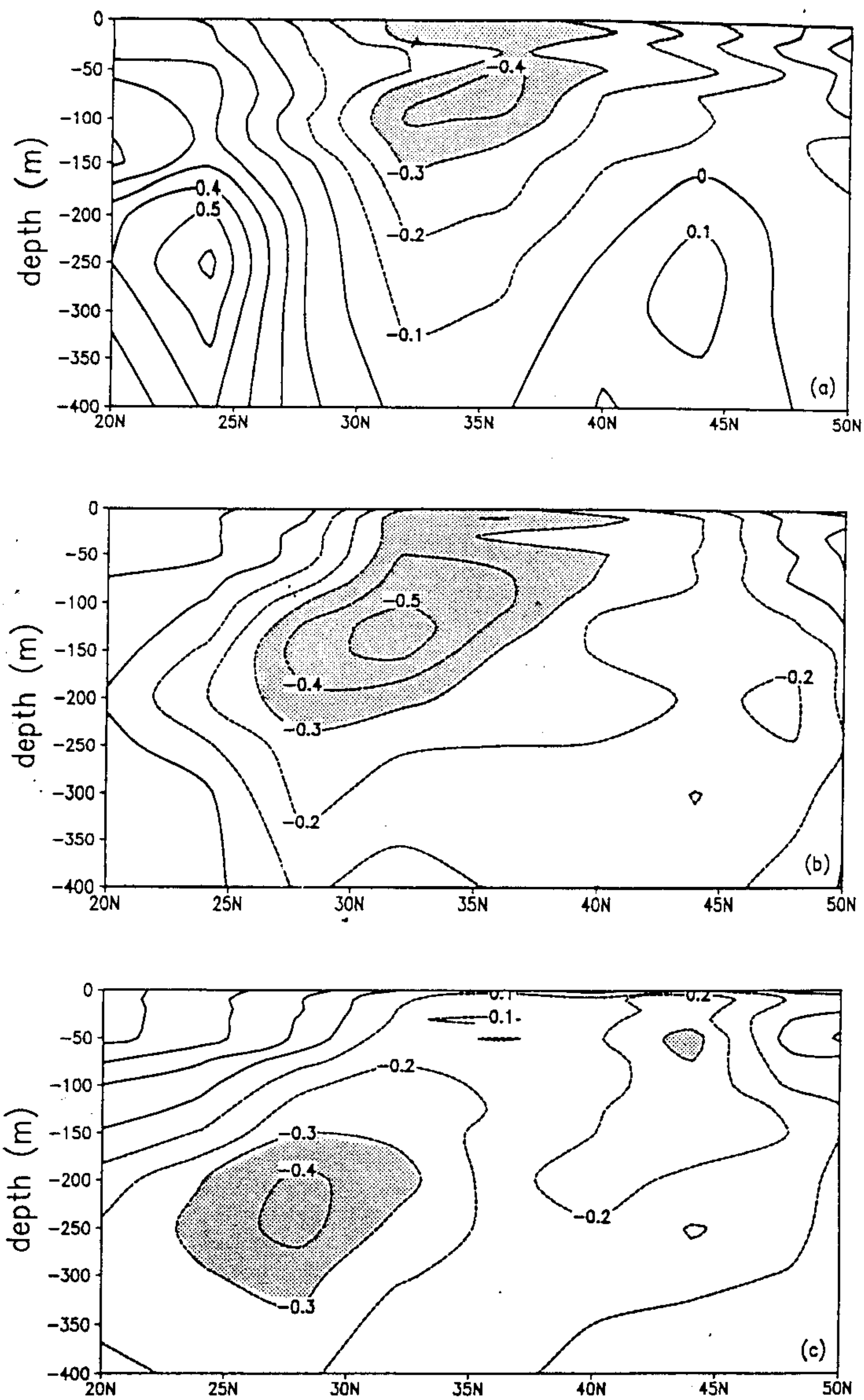
Figure 1



depth (m)
(a)
(b)
(c)
-0.4
-0.3
-0.2
-0.1
0
0.1
0.4
0.5
-0.5
-0.4
-0.3
-0.2
-0.2
-0.1
-0.2
-0.3
-0.4
-0.3
-0.2
-0.2
20N
25N
30N
35N
40N
45N
50N

Figure 2

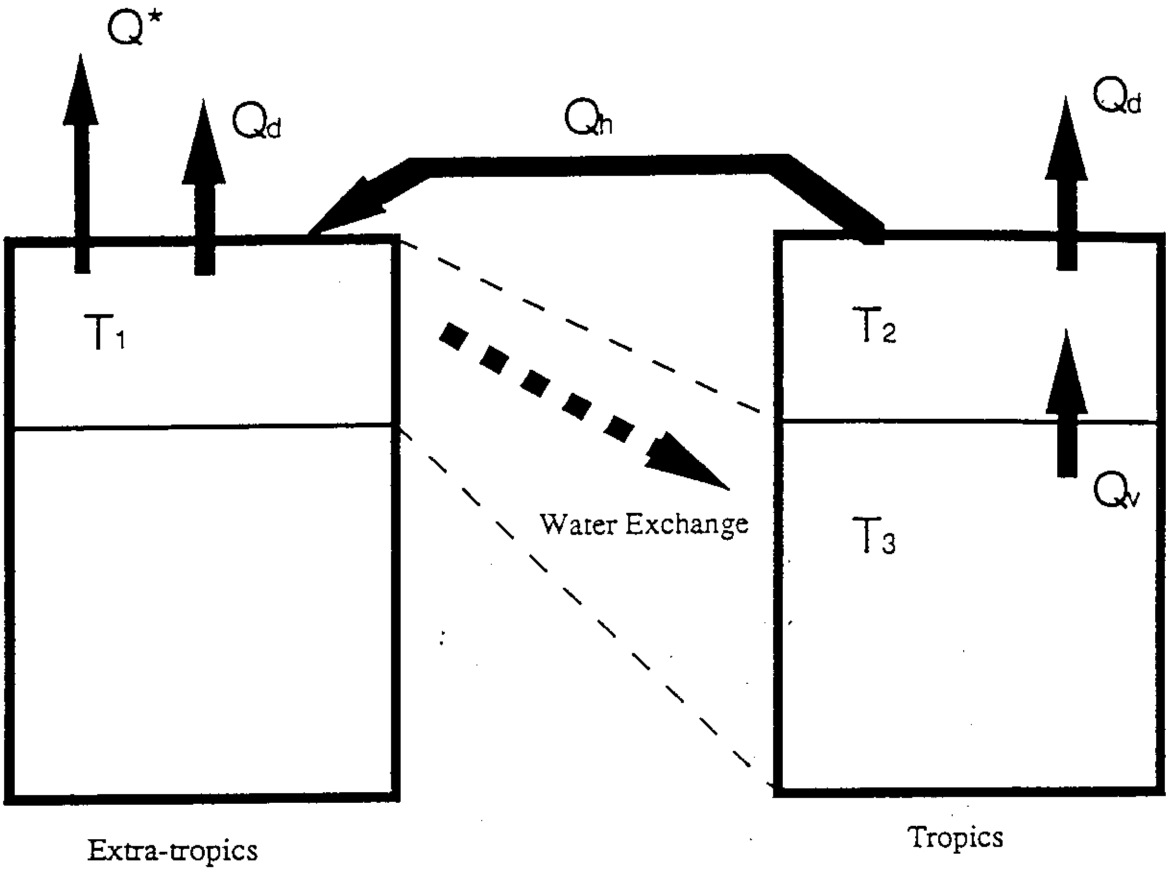


Figure 3

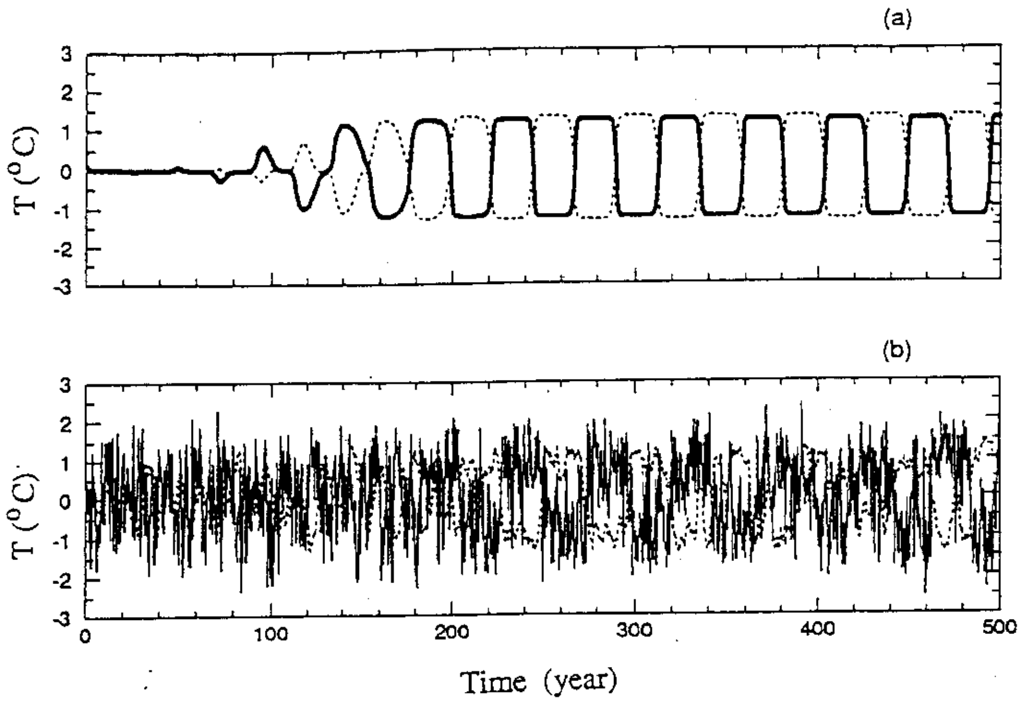


Figure 4