Interdecadal Climate Fluctuations That Depend on Exchanges Between the Tropics and Extratropics

Daifang Gu* and S. G. H. Philander

The unexpected and prolonged persistence of warm conditions over the tropical Pacific during the early 1990s can be attributed to an interdecadal climate fluctuation that involves changes in the properties of the equatorial thermocline arising as a result of an influx of water with anomalous temperatures from higher latitudes. The influx affects equatorial sea-surface temperatures and hence the tropical and extratropical winds that in turn affect the influx. A simple model demonstrates that these processes can give rise to continual interdecadal oscillations.

Interactions between the ocean and atmosphere contribute to climate fluctuations over a broad spectrum of time scales. Studies of those interactions have thus far focused on El Niño–Southern Oscillation (ENSO) phenomena that have a period of 3 to 4 years and whose principal signature is in the tropical Pacific (*1*). Superimposed on this natural mode of the coupled ocean-atmosphere system are interdecadal fluctuations that contribute to the irregularity of the Southern Oscillation (*2*), for example, the recent persistence of unusually warm conditions over the tropical Pacific during the early 1990s. Despite theories and models that explain and simulate the Southern Oscillation (*1*) and that correctly predicted the occurrence of its warm phase, El Niño, in 1987 and 1991 (*3*), the persistence of the recent warming came as a surprise (*4*) that remains unexplained.

The Southern Oscillation 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 (*5*). Many coupled ocean-atmosphere models of ENSO exploit this feature by using an ocean that is composed of two immiscible layers, a warm upper and a cold deeper layer, that are separated by a thermocline whose mean depth is specified. However, these models have not taken into account changes in the thermal structure of the tropical oceans, arising as a result of an influx of water from higher latitudes, for example. The persistent warming during 1990s could have been associated with such an influx.

In the eastern equatorial Pacific a shoaling of subsurface isotherms signals the end of El Niño and the return of colder surface waters. That is how El Niño of 1987 terminated. During the persistent warming of the tropical Pacific in the early 1990s, there was again a shoaling of isotherms after 1992, but this time it had no surface manifestation. The thermocline was deeper in 1992 than it had been in 1987, resulting in much warmer subsurface layers in the east during the period 1990 to 1992 than during the period 1985 to 1987 (Fig. 1). There is no evidence that this warming in the east was associated with a compensatory shoaling of the thermocline in the west (*6*). Here we explore the implications of assuming that the warming was associated with an influx of warmer waters from the extratropics that arose from an earlier change in the prevailing westerly winds in the higher latitudes.

The link between the extratropical and tropical ocean is the relatively shallow, wind-driven meridional circulation that involves the subduction of water parcels in the eastern regions of the subtropical oceans. The water then flows southwestward, essentially adiabatically along surfaces of constant density, to the equatorial thermocline. Here, upwelling transfers the parcels to the surface, whereafter Ekman drift carries them poleward. This circulation, which has been inferred from the distribution of transient tracers such as ^{14}C and ^{3}H (*7*), is maintained by the easterly winds that drive the Ekman drift at the surface and equatorward geostrophic flow in the thermocline. Ocean models (*8*) provide detailed information about the pathways of water parcels (Fig. 2 shows examples) and indicate that the surface waters from large regions converge onto relatively small extratropical windows to the equatorial thermocline. Deser *et al.* (*9*) confirmed the initial part of the path by documenting the movement of unusually cold water subducted in the late 1970s. Presumably a similar (but earlier and warmer) extratropical disturbance caused the change in the structure of the equatorial thermocline during the early 1990s (Fig. 1). Because of equatorial upwelling, this change influences sea-surface temperatures and initiates positive feedbacks between the tropical ocean and atmosphere: the surface temperatures affect the winds, which in turn influence the surface temperatures. These feedbacks can amplify a modest initial perturbation, causing significant warming of the surface waters in the tropics. In these interactions, it is critical that, whereas the atmosphere responds rapidly to a change in sea-surface temperatures, the ocean, because of its greater inertia, responds to changes in the winds on a much longer time scale.

The response of the atmosphere to the warming in the tropics involves an intensification of the extratropical westerlies, leading to colder surface waters (because of evaporation) in extratropical regions that happen to be windows to the equatorial thermocline (*10*). The cold water pumped downward in those regions arrives in the tropical thermocline a dozen years later, halts the warming, and initiates the cold conditions in the tropics. The extratropical winds weaken in response to the cold temperature in the tropics and cause the sea-surface temperatures to increase in the region where surface waters subduct. These arguments imply a continual, interdecadal climate fluctuation with a period that depends on the time it takes for water parcels to travel from the extratropics to the equator [see Latif and Barnett (*11*) 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 intention-

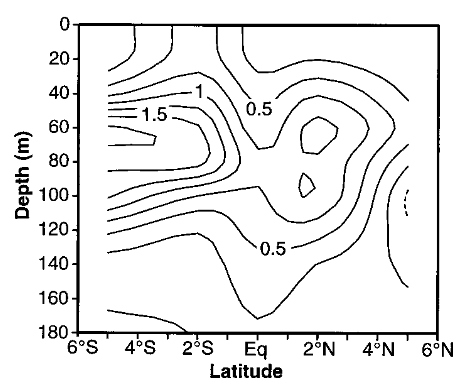

Fig. 1. The change in temperature (in degrees centigrade) as a function of depth and latitude along 110°W, obtained by subtracting the mean temperature for the period 1985 to 1987 from that for the period 1990 to 1992, using data from Tropical Atmosphere Ocean (TAO) Array (*14*). Eq, equator.

Department of Geosciences, Princeton University, Princeton, NJ 08544, USA.

*To whom correspondence should be addressed at Program in Atmospheric and Oceanic Sciences, Department of Geosciences, Post Office Box CN710, Sayre Hall, Princeton University, Princeton, NJ 08544, USA. E-mail: daifang@splash.princeton.edu

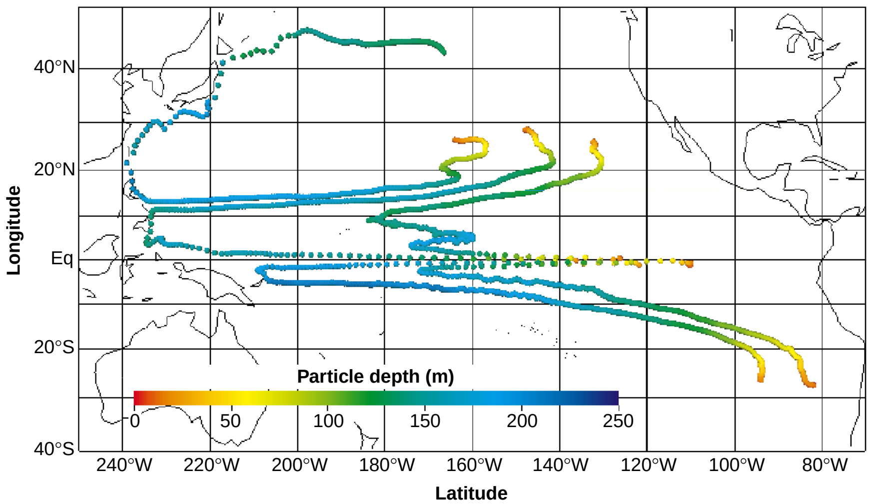

Fig. 2. The paths of water parcels over a period of 16 years after subduction off the coasts of California and Peru as simulated by means of a realistic oceanic general circulation model forced with the observed climatological winds (*8*). From the colors, which indicate the depth of parcels, it is evident that parcels move downward, westward, and equatorward unless they start too far to the west off California, in which case they join the Kuroshio Current. Along the equator they rise to the surface while being carried eastward by the swift Equatorial Undercurrent.

ally suppresses interannual variations in order to focus on the interdecadal variations.

The oceanic component of the model, shown in Fig. 3, consists of two tropical boxes, one at the surface at temperature $\tilde{T}_2$, the other immediately below it in the thermocline at temperature $\tilde{T}_3$, plus an extratropical surface box at temperature $\tilde{T}_1$. (In following discussion $\tilde{A} = \bar{A} + A$, where $\bar{A}$ is the mean or time average and A is the deviation from the mean.) The tropical box covers the approximate region 20°S to 20°N, the extratropical box the regions 25°N to 50°N (and 25°S to 50°S). The temperature $\tilde{T}_2$ is determined by heat fluxes

$$\frac{d\tilde{T}_2}{dt} = -\tilde{Q}_h - \tilde{Q}_v - \tilde{Q}_d \qquad (1)$$

where t is time, $\tilde{Q}_h$ is the atmospheric heat transport to the extratropics, $\tilde{Q}_v$ is the vertical heat transport across the thermocline, and $\tilde{Q}_d$ represents dissipation; $\tilde{Q}_h$ is assumed to be a function of the temperature difference between the tropics and extratropics: $\tilde{Q}_h = F(\tilde{T}_2 - \tilde{T}_1)$. If we linearize by retaining only the first term of a Taylor expansion, then $Q_h = \gamma(T_2 - T_1)$, where γ is a positive constant. The quantity $\tilde{Q}_v$ is proportional to the product of the temperature difference $\tilde{T}_2 - \tilde{T}_3$ and the upwelling velocity that in turn is negatively proportional to the zonal wind $\tilde{U}$: $\tilde{Q}_v \propto -\tilde{U}(\tilde{T}_2 - \tilde{T}_3)$. Therefore, the anomalous vertical heat transport to the lowest order is proportional to $|\bar{U}|(T_2 - T_3) - U(\bar{T}_2 - \bar{T}_3)$. The zonal wind is proportional to the temperature difference between the western and eastern equatorial Pacific. Temperature variations in the west are far more modest than those in the east, so that U is proportional to T_2. Therefore, $Q_v = \delta(T_2 - T_3) - \lambda_2 T_2$ (where λ_2 is a feedback parameter and δ is a positive constant), which, in Eq. 1, represents cooling from mean upwelling and a positive feedback between the surface temperature and zonal winds. The anomalous damping, a term that includes heat loss due to evaporation and that is needed to limit the amplitude of oscillations, is taken to be proportional to the cube of the anomalous temperature T_2, $Q_d = \varepsilon T_2^3$, where ε is a positive constant. Equation 1 now can be written as

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

Similar arguments for the perturbation temperature of the extratropical box yield the equation

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

where λ_1 is a negative constant and α is the fraction of the poleward atmospheric heat transport that remains in the extratropical box; the second term on the right side of Eq. 3 represents the effect of local winds that in turn depend on changes in sea-surface temperatures in the tropics (*10*); Q^* represents stochastic forcing from weather systems unrelated to tropical temperature variations.

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

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

where d is the delay time.

Functional (or delay) differential equations such as Eqs. 2, 3, and 4 have been studied extensively and are known to have unstable (growing) oscillatory solutions (*12*). The presence of damping terms in our equa-

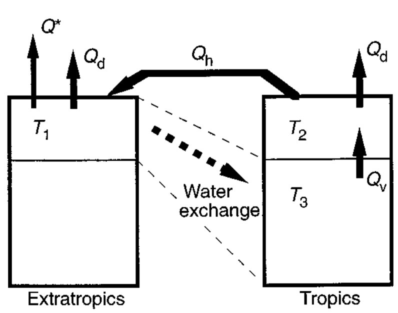

Fig. 3. Sketch of the ocean box model.

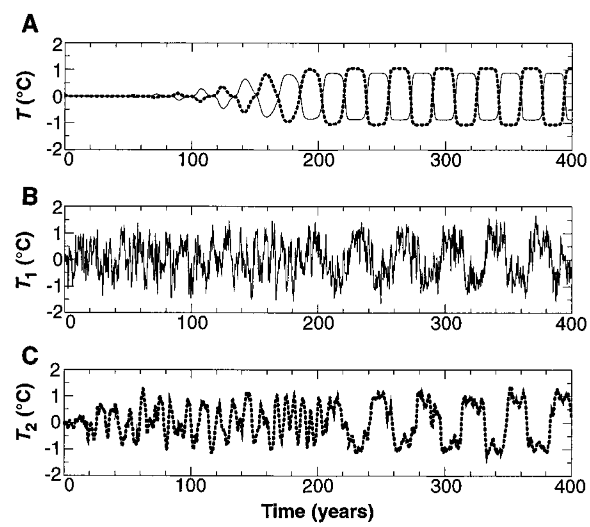

Fig. 4. (**A**) The interdecadal oscillations obtained by solving Eqs. 2 through 4 for the case of no random forcing ($Q^* = 0$). In (**B**) and (**C**), the random white-noise forcing has a normal distribution, a zero mean value, and a root-mean-square value of 2. The solid lines are for T_1 (extratropics) and the dotted lines are for T_2 (tropics). The values of the parameters are as follows: $1/\gamma = 1$ year, $1/\lambda_2 = 100$ days, $1/\delta = 60$ days, $\lambda_1 = -0.5\ \lambda_2$, $1/\varepsilon = 1$ year, and the delay time $d = 15$ years.

tions ensures bounded solutions so that we focus on oscillatory solutions and their sensitivity to changes in the parameters. The reference case shows how an initial small perturbation slowly amplifies before settling down to an oscillation with a period of 35 years and an amplitude of about 1°C (Fig. 4A). 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). It causes a drop in extratropical surface temperature as a result of an intensification of the extratropical westerly winds and enhanced evaporation. The increased latitudinal temperature difference gives rise to a poleward transport of heat that halts both the warming of the tropics and the cooling of the extratropics, establishing an equilibrium state that persists for a considerable time before coming to an end and entering into the complementary phase. 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 extratropics to the equator. In reality, no single value can be assigned to d because surface water subducts over a wide range of latitudes and because there are various routes a parcel can follow to reach the equator (Fig. 2). This suggest a broad-band spectrum for interdecadal climate fluctuations.

In Fig. 4A, the oscillation is perfectly periodic and transitions from one phase to the other are very abrupt. The introduction of stochastic forcing in the extratropics (Fig. 4, B and C) causes the transition to be more gradual. Calculations to explore the sensitivity of the results to specified parameters reveal that the period of the oscillation depends linearly on d. A change in γ, which determines the rate at which heat is transported poleward, affects the magnitude but not the period of the oscillation. Changes in the feedback parameters λ_1 and λ_2 effect bifurcations that correspond to discontinuous changes in the period of the oscillation.

The results presented here demonstrate that it is in principle possible for links between the extratropics and the tropics—rapid and poleward in the atmosphere, slow and equatorward in the ocean—to cause continual interdecadal climate fluctuations. Further studies, observational ones such as the analysis by Zhang *et al.* (*13*) and theoretical ones, are needed to determine the detailed structure of these fluctuations and of the processes, such as oceanic subduction, on which they depend.

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Numerical Simulation of the Cretaceous Tethys Circumglobal Current

Andrew B. G. Bush

Certain paleobiogeographical reconstructions of ocean currents during the Cretaceous (about 144 to 65 million years ago) suggest that a circumglobal tropical current flowed westward through the continental configuration of that time. Although some numerical climate models failed in initial attempts to simulate this current, simulations with a coupled atmosphere-ocean model with relatively high spatial resolution and a late Cretaceous continental distribution show that a circumglobal current is a robust feature even though local surface currents in the Tethys Seaway reverse during the south Eurasian monsoon months.

The continuously changing shapes and positions of continents significantly influence the evolution and migration of life on the planet. One of the most important mammal migrations occurred during the early to mid-Miocene [~23.2 to 11.8 million years ago (Ma)] when collision of the Arabian and Turkish plates formed a land bridge between Africa and Eurasia (*1*). This collision also effectively closed the Mediterranean-Indonesian seaway, thereby blocking from that time until the present a circumglobal pathway in which flowed, it is hypothesized, a warm, saline current (*2–4*).

This westward-flowing Tethys circumglobal current (TCC) is believed to have contributed to tropical climates from 30°S to 30°N from the Triassic through the Cretaceous (~245 to 65 Ma) (*5*). Its demise in the Miocene was associated with a substantial climate change to the coastal regions of the southern Tethys Ocean, including northwest Africa, northern South America, southern North America, and western Eurasia (*2*). This climate change had a fundamental impact on the flora and fauna of these regions: For example, vegetation in North America transformed from that of a tropical rain forest in the Eocene (~55 to 34 Ma) to that of a savanna in the Miocene (*6*).

The failure of a recent numerical simulation of the mid-Cretaceous ocean to reproduce such a circumglobal, westward-flowing, equatorial current led to increased scrutiny of the geological evidence and to a controversy concerning the very existence of the TCC (*7*). In the simulations, the Tethys Ocean has prominent eastward currents—part of the wind-driven gyre circulation—but no westward return flow indicative of a TCC along the continental margins of southern Eurasia, Africa, or South America (*8, 9*). This discrepancy between biogeographic evidence and modeling efforts was attributed to the absence of a poleward shift of the mid-latitude westerly winds in the simulation, a shift which had previously been an assumed feature of an ice-free planet in the reconstructions (*8, 10*).

This report describes a numerical simulation that has the TCC as a robust feature of the late Cretaceous ocean circulation. We used an atmospheric general circulation model (GCM) (*11*) that is dynamically and thermodynamically coupled to an oceanic

Atmospheric and Oceanic Sciences Program, Princeton University, Post Office Box CN710, Sayre Hall, Princeton, NJ 08544–0710, USA.