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A diabatic mechanism for decadal variability in the tropics

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Abstract

Can atmospheric forcing of the ocean in high latitudes induce decadal variability in low latitudes? Most theoretical studies that have considered this question assign a critical role to adiabatic, advective, subsurface oceanic links between the tropics and extra-tropics. Observational evidence of such links is proving elusive. This study posits that given the constraint of a balanced heat budget for the ocean in a state of equilibrium, atmospheric forcing over a broad spectrum of frequencies in high latitudes can force decadal variability in low latitudes without any explicit evidence of oceanic links.

The oceanic response to an abrupt change in diabatic forcing, a sudden increase in heat loss in high latitudes say, is characterized by two time-scales. The one, τ_w, is relatively short and is associated with planetary and coastal waves that propagate from the disturbed region to the equator (and then back to higher latitudes.) The other, τ_d, is on the order of a few years and depends on diabatic processes responsible for increasing the oceanic heat gain in low latitudes. Through these processes the system is driven towards a new balanced heat budget in which the heat gain, mainly in the equatorial upwelling zones, equals the heat loss in high latitudes. When the forcing, rather than abrupt, is sinusoidal with period P, then the amplitude of the response depends on the ratio P/τ_d. The response is modest when that ratio is small because the period P is too short for the ocean to adjust. As P gets larger compared to τ_d, the amplitude increases, but explicit evidence of the waves that connect high and low

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latitudes is very hard to detect. The ocean acts as a low pass filter to the forcing with characteristic timescale τ_d.

Keywords: Thermocline variability; Ocean circulation; Ocean waves; Climate

1. Introduction

The tropical Pacific Ocean has a rich spectrum of variability that includes the Southern Oscillation at a distinctive period of approximately 4 years, and broad-band lower-frequency fluctuations referred to as decadal variability. The latter, which modulates the properties of the former (Fedorov and Philander, 2000), was associated with a warming of the surface waters of the eastern tropical Pacific in the late 1970s as is evident in Fig. 1. The spatial structure of this warming is similar to that of the Southern Oscillation, except for having a much wider latitudinal scale, and for having a connection to the extra-tropics (Zhang et al., 1997). The origin of this so-called Pacific Decadal Oscillation (Mantua et al., 1997) is a matter of much debate.

One proposed mechanism for decadal variability in the tropics invokes oceanic connections between the tropics and extra-tropics. In particular, Gu and Philander (1997) proposed that the ocean provides a "bridge" between the tropics and the subtropics by subducting anomalies, created in mid-latitudes by surface forcing, into the main thermocline, where they are conservatively advected to the equator. Once these anomalies reach the equator, they modify the thermocline structure, affecting sea surface temperatures and therefore feeding back on the surface climate. In this mechanism, what determines the timescale of the variability is the adiabatic advective time, which is assumed to be decadal (Gu and Philander, 1997).

Deser et al. (1996) tentatively identified subducting anomalies that traveled equatorward from the extra-tropics, but further investigations either failed to trace them to low latitudes or found their amplitudes to be greatly reduced by the time they reached the equator (Harper, 2000; Schneider et al., 1999a,b). At present, the evidence in favor of this mechanism is

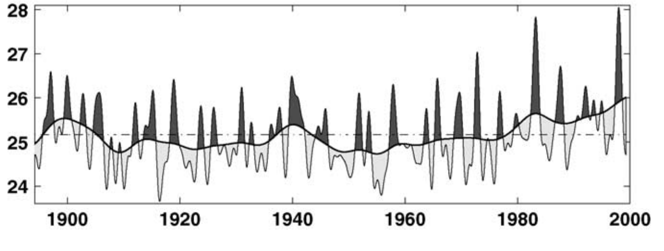

Fig. 1. Interannual fluctuations in SSTs at the equator in the Eastern Pacific. Temperatures are averaged over the area from 5°S to 5°N and from 80°W to 120°W. Superimposed are the decadal fluctuations, obtained by low-pass filtering. Annual cycle and higher frequencies are removed. The dash-dotted line is the time average for the record (from Fedorov and Philander (2000).

suggestive at best. The problem is that anomalous signals that subduct into the main thermocline disperse into waves, thus attenuating the amplitude and coherence of the anomaly (Liu, 1993, 1999; Lysne et al., 1997; Yang, 1999; Johnson and Marshall, 2002). Therefore, available measurements of anomalous propagation are insufficient to demonstrate a clear connection between extra-tropical forcing and variability in the tropics. This does not mean that there is no connection, but rather those tracking anomalies in the interior is an ill-suited method for establishing such connection.

Previous studies have focused on adiabatic oceanic processes, that is, the processes in which the tropics and extratropics are connected through conservative advection. In those cases the decadal scale of the variability is determined by the time it takes for anomalies to reach the equator. In this paper, we explore the thesis that changes in the heat budget of the ocean are of central importance to decadal variability in low latitudes. We propose that the tropics and extratropics are indeed connected through a variety of pathways, but that the decadal timescale emerges from the nature of tropical diabatic processes.

The wind-driven circulation of the upper ocean affects a poleward transport of heat, from the upwelling zones of the tropics and subtropics where the ocean gains most of its heat, to higher latitudes where that heat is lost to the atmosphere. In a state of equilibrium, the heat budget is balanced so that heat gain equals heat loss. Should there be a change in the heat loss, then the ocean has to adjust so that the heat gain changes in a corresponding manner. Boccaletti (2004) and Boccaletti et al. (2004) explore in detail how the ocean adjusts to a sudden increase in the flux of heat out of the ocean in high latitudes. Two sets of processes are involved. The first corresponds to the adiabatic adjustment to heat source and heat sink perturbations studied by Cane and Sarachik (1979). This involves coastal, equatorial and planetary waves that redistribute mass across the basin in a time τ_{w} thus altering the depth of the thermocline locally without changing its spatially averaged depth. The second set of processes depends on the thermodynamics of the ocean to alter the net volume of warm water in the ocean on a time-scale τ_{d}. A perturbation in higher latitudes affects the equatorial zone by radiating waves that quickly reach the equator where the depth of the thermocline is altered. This brings into play diabatic processes because the depth of the thermocline determines how much heat the ocean absorbs. In the eastern equatorial Pacific, for example, a shoaling of the thermocline that leads to a decrease in surface temperatures is associated with an increase in the oceanic heat gain. These two sets of processes permit an increase in the oceanic heat loss in the extra-tropics to translate into an increase in the oceanic heat gain in low latitudes, thus restoring a balanced heat budget.

This paper uses these results concerning the oceanic adjustment to an abrupt change in surface forcing to determine the response of the equatorial ocean to periodic forcing in the extra-tropics. In reality, in mid- and high latitudes, the atmosphere forces the ocean (with fluctuating momentum and heat fluxes) over a very broad spectrum of frequencies. Waves propagate this information to the equator on a time scale τ_{w}. However, the integrated effect of those waves on the oceanic heat budget is felt only on time scales that approach τ_{d}. The diabatic response of the tropics to extra-tropical forcing is modest at frequencies higher than τ_{d}^{-1}, but becomes significant at low frequencies. (The ocean acts as a low-pass filter for the extratropical forcing.) The boundary waves travel so fast that they may not be explicitly evident at low frequencies. A search for propagating signals (that are subject to dispersion and dissipation) is therefore not an effective way to establish the connections between the tropics and extra-tropics. In this paper, we examine those connections by forcing an idealized ocean basin over a range of frequencies along its northern wall.

Section 2 presents the numerical model used to investigate the issue. Section 3 illustrates the theory. Section 4 is a summary of the results and Section 5 draws conclusions.

2. The model

A small tropical basin, extending from 16°N to 16°S, 40° wide, and 5000 m deep, is forced by uniform easterly winds. The wind strength is 0.05 Pa. The model used is the GFDL MOM 4. Horizontal resolution is specified as 1/2° everywhere, with 32 levels in the vertical. The vertical resolution is 10 m in the upper 200 m. Vertical mixing is the Pacanowski and Philander (1981) scheme solved using a maximum vertical mixing coefficient of 50×10^{-4} m^2/s and a background diffusivity of 0.01×10^{-4} m^2/s, unless otherwise specified. Constant lateral mixing coefficients are $A_m = 2.0 \times 10^3$ m^2/s and $A_h = 1.0 \times 10^3$ m^2/s for momentum and heat, respectively. The basin is chosen to be small in order to allow for extensive numerical exploration at relatively low computational cost.

In steady state, the surface temperatures T are restored towards a specified profile T^*, according to a heat flux formulation given by (Haney, 1971)

$$Q = (T^* - T) \times 50\,\mathrm{W/m^{2\circ}C} \tag{1}$$

The temperature T is restored towards $T^* = 25\,^\circ$C between 12°N and 12°S, and it linearly decreases to $T^* = 10\,^\circ$C poleward of those latitudes. This temperature profile is unrealistic for the extratropics, as it confines a large temperature gradient in narrow strips at the northern and southern boundaries. However, this work focuses on the response of the tropics to extratropical forcing, and for our purposes, it does not matter how the extratropical forcing is achieved.

The boundary conditions (1) are of great importance as it distinguishes this work from previous idealized studies of adjustment such as, for instance, those of Cane and Sarachik (1979). Those authors effectively specify the surface heat budget with sources/sinks of mass. Here the heat budget is allowed to adjust as a function of surface temperature and therefore as a function of the state of the ocean.

A summary of the mean state of the model is shown in Fig. 2. The model is spun up from isothermal 10 °C. After 30 years, the upper ocean is in steady state while the abyss continues to warm at a steay slow pace, due to the small background diffusion (Boccaletti et al., 2004). At the equator, the surface flow is easterly, while off the equator the Ekman transport is directed poleward. Two approximately symmetric wind-driven circulations, one for each hemisphere, draw cold water from the mid-latitudes and transport it at depth towards the equator, where it is upwelled, warmed, and returned poleward at the surface. The equatorial thermocline is tilted to balance the wind forcing, and produces a large cold tongue in the eastern tropical portion of the domain. The heat gained in the cold-tongue region is balanced by the heat lost at the boundaries, where the warm water is restored towards colder temperatures. The overall thermocline structure, heat flux patterns, and circulation are highly simplified but reminiscent of the observed climatology.

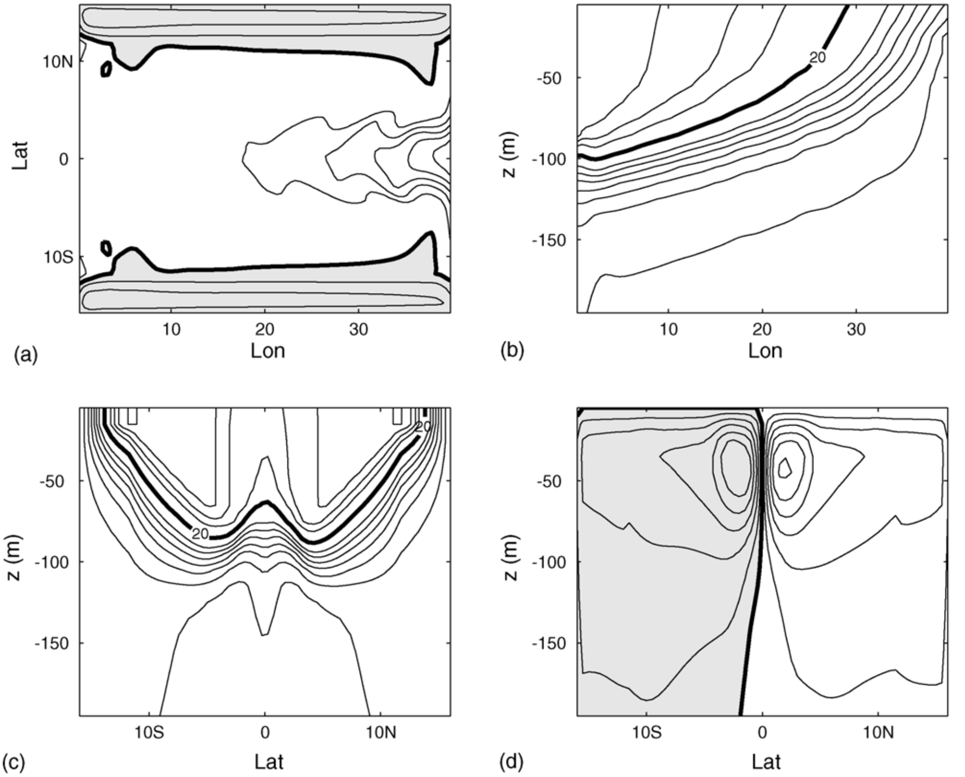

Fig. 2. Mean structure of the steady state solution for the reference case: (a) surface heat fluxes; shaded areas are negative; contour interval 50 W m^{-2}; (b) equatorial temperature section; contour interval 2 °C; (c) meridional temperature section along 20°E; contour interval 2 °C; (d) seridional streamfunction; shaded areas are negative; contour interval 2.5 Sv.

Once the upper ocean has reached steady state, that state is taken as initial condition for a series of 20-year runs in which we impose a periodic perturbation to the high latitude restoring temperature. The restoring temperature at the poleward boundaries is denned by $T^* - \bar{T}^* = 5.0 \times \sin(2\pi t/P)$ where P, the period of the perturbation, is 1, 2, 5, 10, and 20 years. Notice that while the extratropical forcing is perturbed, the tropical forcing is maintained steady, so that any time-dependent response in the tropical state is remotely controlled.

3. Theory

This work completes a series of papers (Boccaletti et al., 2004; Boccaletti, 2004) investigating the role of the heat budget in constraining the thermal structure of the upper ocean and its response to perturbation.

The problem of what determines the depth of the observed thermocline is a central question in physical oceanography (Pedlosky, 1998). The thermocline is the most prominent feature of the oceanic thermal structure and its depth and width determine many properties of the oceanic response to perturbation. Ventilated thermocline theories, which are explicitly adiabatic, give a reasonable picture of the wind-driven circulation (Welander, 1959; Robinson and Stommel, 1959; Luyten et al., 1983; Huang, 1991), but they are fundamentally incomplete as they require the specification of background stratification. Boccaletti et al. (2004) showed that in steady state, it is possible to estimate that background stratification by imposing a balanced heat budget. In order to do so; however, the tropics have to be explicitly considered. Regions of upwelling, specifically in the tropics, have the unique property that the underlying vertical thermal structure is directly coupled to the heat budget: when the thermocline is deep, surface temperatures are warm, and the heat gain is small; when the thermocline is shallow, surface temperatures are cold, and the heat gain is large. Furthermore, the tropical ocean, which is neglected in ordinary ventilated thermocline theories, is of paramount importance for closing the heat budget as it is there that most of the heat enters the ocean.

Boccaletti et al. (2004) expressed the heat budget of the equatorial region as:

$$\alpha(T^* - T) = (\kappa + w_{\mathrm{Ek}} h)\frac{\Delta T}{D} \tag{2}$$

where the thermocline is of scale depth D; T is the mean temperature of the warm water sphere bounded by the thermocline, while $\Delta T = T - T_{\mathrm{c}}$ is the temperature difference across the thermocline between warm waters at temperature T and the abyssal waters at temperature T_{c}. The term on the left hand side of Eq. (2) represents the heating due to air–sea exchanges. On average, the temperature of the warm water sphere will be driven towards an effective restoring temperature T^* on a timescale represented by α (Haney, 1971). α is the restoring timescale in units of m/s. The term on the right hand side of Eq. (2) represents the cooling of upper ocean temperature due to both diffusion to the abyss and to cold upwelling. The diffusive term is estimated as $\Delta T\kappa/D$, where κ is the diffusivity. The upwelling term is represented as $w_{\mathrm{Ek}}h/D\Delta T$, where w_{Ek} is an entrainment velocity at the base of the mixed layer and h the mixed layer depth. In the limit of a very deep thermocline, the term goes to zero as expected, as no cold water can be entrained into the mixed layer; on the other hand, in the limit of $D = h$, all the entrained water is at a temperature T_{c} and is then exported at a temperature T, leading to a cooling $w_{\mathrm{Ek}}\Delta T$. Obviously, the heat budget responds also to the strength of the wind through the magnitude of w_{Ek}, but in this paper, we shall consider the atmosphere as fixed. We shall return to this issue in the conclusions. (Other choices can be made for the dependence of upwelling on thermocline depth, provided they obey the same limits for deep and shallow thermocline. Another choice could be a hyperbolic tangent (Zebiak and Cane, 1987). The choice of Eq. (2) is, however, advantageous, particularly in the context of steady state theory, as it allows recovery of a number of classical results as special limits (Boccaletti et al., 2004).) The effects of diffusion can be significant when κ is large, but in this paper, we shall consider only the limit in which κ is small, as suggested by observations (Ledwell et al., 1993).

Boccaletti (2004) considered the inital value problem of an idealized ocean, subject to an extratropical perturbation of its heat budget. A number of studies have dealt with the

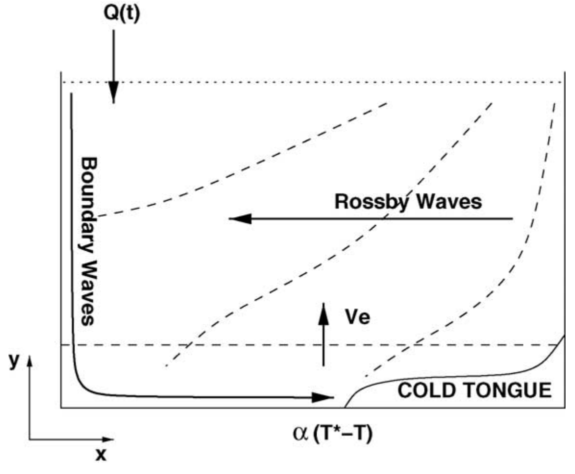

Fig. 3. Adjustment to mid-latitude forcing in a closed basin. The response to a mass sorce $Q(t)$ results in a boundary adjustment and an interior Rossby adjustment.

dynamical adjutment to perturbation, most notably those of Cane and Sarachik (1979); Kawase (1987); Liu (1999); Johnson and Marshall (2002), and they all show how the perturbation projects on a series of dispersive waves, inundating the basin. The fastest waves are those propagating along the boundary as Kelvin waves (Johnson and Marshall, 2002). Interior modes are slower, some even resembling simple advection (Liu, 1999). A cartoon of the processes involved is shown in Fig. 3. As a result of a subtropical perturbation, a signal reaches the western boundary, where it then propagates equatorward. After crossing the basin at the equator and reaching the eastern boundary, the signal travels poleward, shedding a westward Rossby front, which communicates the perturbation to the main thermocline (Anderson and Gill, 1975). For the purposes of long-term variability, the boundary waves are practically instantaneous, but the Rossby adjustment introduces a lag due to the slow propagation of long Rossby waves cross the basin (Cessi and Primeau, 2001; Johnson and Marshall, 2002).

In the real world, as well as in models, this intricate wave response makes observation of the adjustment particularly difficult. Boccaletti (2004) showed that despite the complexity of the adjustment a simple perspective could be gained by considering the heat budget. Fig. 4 shows the anomalous heating and cooling, as a function of time, for a run in which the high latitude temperature is suddenly changed from 10 to 15 °C. The dynamical adjustment is similar to that shown in Fig. 3 but from the point of view of the heat budget, the adjustment is fairly simple. Once a perturbation has occurred, the ocean reaches a new balanced heat budget by changing the thermocline depth and therefore changing the size of the cold tongue. This occurs at rate determined by the net increase or decrease of warm water, which is set by entrainment across the thermocline.

The timescale for the diabatic adjustment of the thermocline can be estimated as follows. From the representation of tropical upwelling given in Eq. (2), we expect the time-dependent

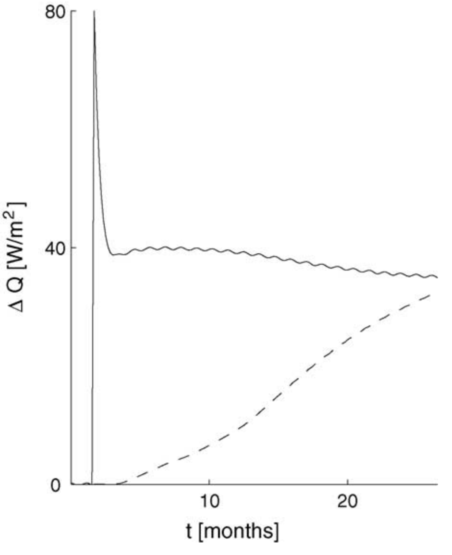

Fig. 4. Response of the heat budget to an imposed cooling anomaly, obtained by instantaneously changing the high latitude restoring temperature from 10 to 15 °C. Plot of the time evolution of the heat anomaly in the region of heating (dashed, averaged over [−4° to 4°]), and cooling (solid, averaged elsewhere). The anomalous heating is multiplied by −1 for ease of comparison.

problem to be governed by

$$D_t = -w_{\mathrm{Ek}}\frac{h}{D} \tag{3}$$

Eq. (3) just states that the thermocline can only deepen globally at a rate proportional to upwelling into the warm water sphere. Eq. (3) then implies a timescale

$$\tau_{\mathrm{d}} \sim \frac{D^2}{w_{\mathrm{Ek}}h} \tag{4}$$

which for $D \sim 200$ m, $w_{\mathrm{Ek}} \sim 10^{-5}$ m/s, $h \sim 50$ m gives a timescale of order 2 years.

In this paper, we consider the response of the tropical ocean to periodic perturbations. We shall assume that the perturbation is around a mean state defined by a mean thermochne depth $\bar{D}$ and a mean static stability ΔT so that

$$D = \bar{D} + D' \tag{5}$$

If we assume that the perturbation in high latitudes produces a forcing $F(t)$ then Eq. (3) becomes

$$D'_t = -\frac{w_{\mathrm{Ek}}h}{\bar{D}^2}D' - F(t-\tau_{\mathrm{w}}) = -\frac{1}{\tau_{\mathrm{d}}}D' + F(t-\tau_{\mathrm{w}}) \tag{6}$$

where $F(t-\tau_{\rm w})$ is the midlatitude forcing, delayed by the amount of time it takes for the signal to reach the tropics. The heat budget allows us to relate the perturbation depth of the thermocline D to the heat perturbation. The surface heating perturbation is given by $Q' = -\rho_0 C\alpha T'$ where α is the restoring coefficient, and T' the perturbed temperature of the upper ocean. The heating is equal to the upwelling across the temperature difference ΔT (Eq. (2)). Linearizing Eq. (2) (and ignoring diffusion) we get:

$$\alpha T' = w_{\rm Ek}\left(\frac{\Delta\bar{T}h}{\bar{D}^2}D' - \frac{h}{\bar{D}}T'\right) \tag{7}$$

where we have assumed that $\Delta T = \Delta\bar{T} + T'$, as the temperature below the thermocline does not change. A relation between T' and D' can then be found as:

$$T'\left(\alpha + \frac{\bar{D}}{\tau_{\rm d}}\right) = -\frac{D'}{\tau_{\rm d}}\Delta\bar{T} \tag{8}$$

so that finally, the relation between the depth of the thermocline and the heat budget is

$$\frac{1}{\rho_0 C}Q' = \frac{\alpha\Delta\bar{T}}{\tau_{\rm d}\alpha + \bar{D}}D' \tag{9}$$

Multiplying by $\rho_0 C\alpha\Delta\bar{T}/(\alpha\tau_{\rm d} + \bar{D})$, Eq. (6) can be written as

$$Q'_t = -\frac{Q'}{\tau_{\rm d}} - Q^F(t-\tau_{\rm w}) \tag{10}$$

where $Q^F(t-\tau_{\rm w})$ is the extratropical heat budget perturbation. The lag introduced $\tau_{\rm w}$ is introduced by the propagation of boundary and Rossby fronts around and across the basin (Cessi and Primeau, 2001; Johnson and Marshall, 2002). In spectral domain, the response of system (10) is:

$$\tilde{Q}' = \frac{\tilde{Q}^F}{(\omega^2 + \tau_{\rm d}^{-2})^{1/2}}{\rm e}^{i\,{\rm atan}}(\omega\tau_{\rm d}) - i\omega\tau_{\rm w} \tag{11}$$

For high frequencies $\omega \gg \tau_{\rm d}^{-1}$, the amplitude of the response is heavily damped. For lower frequencies, the amplitude of the response is smaller and for $\omega \ll \tau_{\rm d}^{-1}$ the damping is proportional to $\tau_{\rm d}$. The timing of the response is also phase shifted: for small frequency $\omega \ll \tau_{\rm d}^{-1}$ the phase shift is approximately $\omega(\tau_{\rm d} + \tau_{\rm w})$, whereas for high frequencies, the phase shift is closer to $\pi/2 + \tau_{\rm w}\omega$. Fig. 5 shows the response function given by Eq. (11). There are two kinds of delay mechanisms for this simple setup; for an observer measuring the response of the Tropics to a periodic forcing in the extratropics, the phase lag can be due to the actual time of propagation of the signal or to the phase shift introduced by the diabatic damping. In the case of low frequency forcing, when $\tau_{\rm d} > \tau_{\rm w}$ the phase shift is mainly due to the diabatic timescale $\tau_{\rm d}$. In this case, trying to detect the propagation of anomalous signals in order to predict the phase lag between the forcing and the response would be misleading, as the phase lag is due to the thermodynamics of the system, not to the dynamics. Assuming that the response of the tropical ocean is a forced problem, Eq.

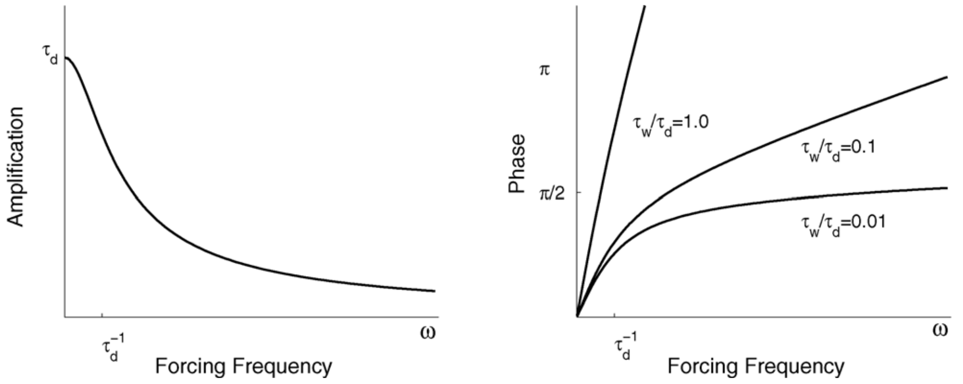

Fig. 5. Amplitude and phase change due to the linear filter represented by Eq. (11).

(11) says that anomalous heat fluxes in the subtropics produce a response in the low latitude heat fluxes which is governed by a linear relationship, whether the dynamics responsible for the communication are linear or not.

Notice that the surface damping α and the actual diabatic timescale τ_{d} are different. The low-pass filtering in this case is not mere reddening due to the relationship between atmospheric forcing and surface fluxes, but rather it is due to the relationship between those fluxes and the subsurface thermal structure. It is also important to note that the proposed mechanism is different from that illustrated by Johnson and Marshall (2002), who rely on a dynamical mechanism through which the equator constitutes a buffer zone for signals propagating around the basin. The response timescale of their model is unrelated to the adjustment timescale τ_{d} invoked here, as it depends on geostrophic outflow at the southern end of their domain.

We now turn to the numerical model described in Section 2 to illustrate these results.

4. The response to high latitude forcing

Fig. 6 shows a time–latitude diagram for the depth of the 20 °C isotherm. This is a plot similar to that shown by Schneider et al. (1999) (their Fig. 6), and, as in that case, detection of the connection between the tropics and the extratropics is difficult. The anomaly created at the very northern boundary appears to propagate southward, but is swiftly dispersed a few degrees latitude below. At the same time, the tropics exhibit a vigorous variability that mirrors the one induced in high latitudes, but out of phase by about a year and a half.

Fig. 6 shows one cycle in a run with a 5-year period. At the beginning of a period, anomalously warm water in high latitudes is pumped down by the Ekman flow as it encounters the boundaries; after half a cycle, when anomalous cooling occurs, warm isotherms move equatorward and colder water is exposed to the surface. In the tropics, the response manifests itself in the size of the cold tongue, which controls the heating, and is out of phase with respect to the high latitude forcing. The maximum heating in the tropics lags behind the maximum cooling in high latitudes. In the case of the 5-year period run shown

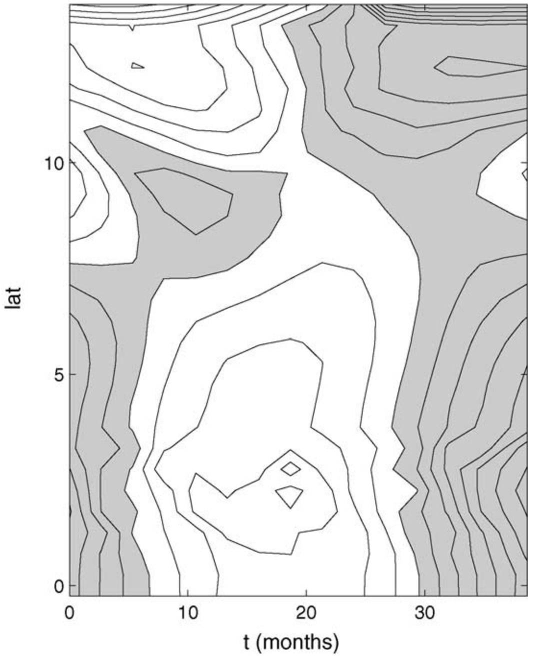

Fig. 6. Time–latitude plot of the zonally averaged depth of the 20 °C isotherm. Shaded regions are negative; contour interval is 1 m. The timescale is relative to the beginning of the plotted time-series for convenience. The case shown is for a run with a period *P* of 5 years.

in Fig. 6, if the cooling is minimum at the beginning of a cycle, the tropical heating is minimum almost a quarter of a cycle later, when the cold tongue is anomalously small and warm; the same lag is observed in the second half of the cycle, when the extratropical heat loss is at a maximum. The change in size of the cold tongue is only due to a change in the subsurface thermal structure, as the tropical surface forcing is fixed: when the thermocline shoals, warm surface layers are peeled off, revealing the underlying colder water to the surface; when the thermocline deepens the opposite occurs (Boccaletti et al., 2004).

By construction, the tropics and the extratropics are related in this set of experiments; the only time-varying forcing is at the poleward boundaries, so that the periodic response in the tropics must be remotely forced. The seemingly consistent lag between the extratropical forcing and the tropical response would lead us to look for a connection between the two areas. In particular, we might expect to be able to detect the propagation of anomalies from the extratropics to the tropics. After all, the warm water responsible for deepening the thermocline and reducing the cold tongue appears at the equator some time after anomalously warm water is produced in high latitudes, so it is natural to assume that it is the same water that has been advected to the tropics (Gu and Philander, 1997). However, Fig. 6 suggests that the actual communication is not as straightforward.

Fig. 7 shows a series of snapshots of anomalous thickness for the layer bounded by the 19 °C isotherm below and the 20 °C above. The propagating disturbances in Fig. 7

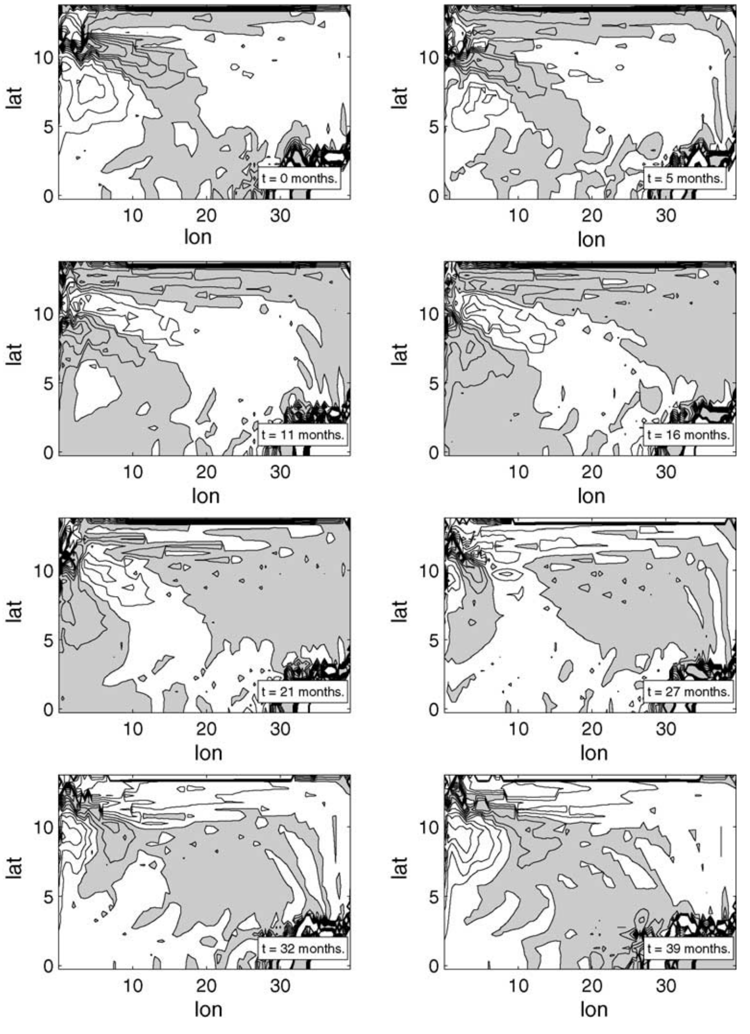

Fig. 7. Sequence of snapshots of the thickness anomalies of the layer bounded by the 19 and 20 °C isotherms. Shaded values are negative, and the contour interval is 1 m.

mirror those described in the cartoon of Fig. 3. The response projects on both boundary and interior modes. When the anomaly is created in high latitudes, first a significant stream is established along the western boundary. In just a few months the equator "feels" the high latitude pertubration through this pathway. Not all of the anomaly is redistributed along the boundary however. A perturbation also travels in the interior propagating from East to West and from the extratropics to the tropics. This is the slower planetary wave response. When integrated zonally the meridional propagation of these slower modes is not strong enough to dominate, as most of the perturbation is transferred swiftly along the boundary, from where it reaches the equator in just a few months.

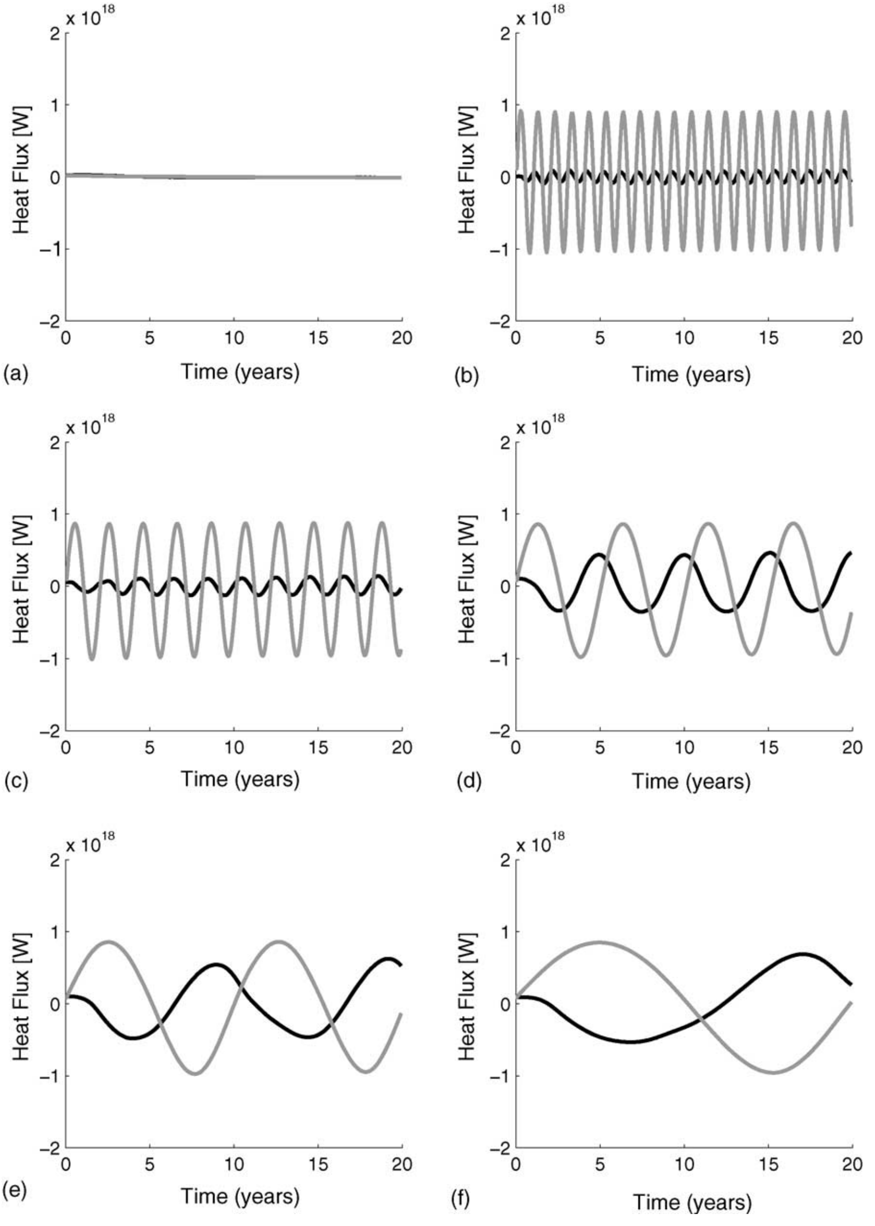

Fig. 8. Response of the model to different periodic forcing. The grey line is the anomalous heat loss in high latitudes, the black thick line is the anomalous heat gain in the tropics. The different periods are: (a) steady state; (b) 1 year; (c) 2 years; (d) 5 years; (e) 10 years; (f) 20 years.

The problem, as pointed out explicitly or implicitly by many authors before us, is that anomalies are propagated through a number of dispersive waves that actually effect the mass transfer without preserving the coherence of the anomaly (Liu, 1993, 1999; Lysne et al., 1997; Yang, 1999; Cessi and Primeau, 2001; Johnson and Marshall, 2002; Boccaletti, 2004). However, waves in particular boundary signals with high propagation speed and small amplitude, are hard to detect, so that following the dispersion of an anomalous signal is very hard. The heat budget however provides a possible alternative to detecting that connection. The diabatic response of a basin is not only governed by wave propagation. Waves are just the way in which mass is transferred around the basin, but the final response, especially at low frequencies when explicit evidence of the wave propagation may be difficult to detect, is governed by diabatic processes.

Fig. 8 shows the response of the basin to subtropical periodic forcing at different frequencies. Depicted are the high latitude heat loss Q^- (gray line) and the tropical heat gain Q^+ (black line), integrated over the areas of heat loss and heat gain, respectively. When the thermocline is shallow, the cold tongue is large, and therefore, the tropical heat flux is large. Vice versa, when the thermocline is deep, the cold tongue is small and the tropical heat flux small. The response is clearly dependent on the period of the forcing: for fast periods the equatorial response is heavily damped; as the period increases, the amplitude of the response also increases. Also the phase is sensitive to the period. Higher frequencies are more out of phase than lower frequencies when the maximum cooling and maximum heating (almost) coincide (cooling is expressed as a negative quantity, so a maximum in the anomaly curves of Fig. 8 corresponds to a lower cooling than average).

A spectral analysis of the response (Fig. 9) gives a quantitative picture of how the basin responds to the forcing. Shown are the amplitude and the phase lag as functions of forcing frequency. The theory of Section 3 is superimposed on the data, with an estimated diabatic timescale $\tau_d = 1.2$ years and a lag $\tau_w = 2$ months, corresponding to the lag introduced by

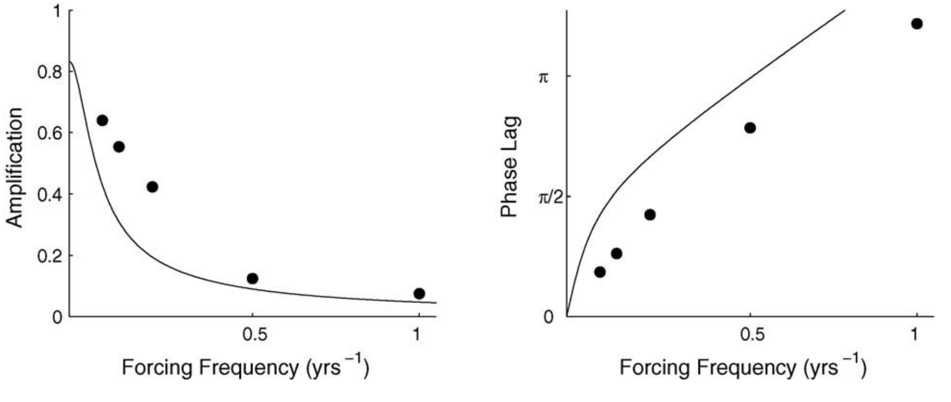

Fig. 9. Response of the basin to periodic forcing as a function of the forcing frequency. Left: amplification of the forcing. The line represents a theoretical curve of $1/(\omega^2 + \tau_d^{-1})^{1/2}$ with $\tau_d \sim 1.2$ years. Right: phase lag. The line is for a theoretical curve of atan $(\omega\tau_d) + \tau_w$, where $\tau_w \sim 2$ months and $\tau_d \sim 1.2$ years. The dots are the results of the simulations.

boundary signals. As expected from a low-pass filter, the high frequency signals are heavily damped, while the long periods (on decadal timescales) are more or less conserved to the equator. Such a clear dependence of amplitude on frequency would not result from a purely advective process. The linear filter hypothesis allows us to interpolate the data and identify the timescale τ_d, which for the graph in Fig. 9 is $\tau_d \sim 1.2$ years, and a timescale $\tau_w \sim 2$ months, corresponding to the timescale for the propagation of boundary signals.

In the phase diagram, we can identify two regimes, where different dependence on the frequency ω is observed. For high frequency, the dependence appears to be linear. Those are the strongly damped signals, dominated by the propagation of waves. For low frequency, the regime changes and a different slope is observed. That is the limit in which the diabatic processes dominate.

5. Conclusions

Decadal climate variability is often attributed to a connection between the tropics and the extra-tropics. In a model proposed by Gu and Philander (1997) this connection occurs adiabatically through the subduction and advection of anomalies from the high to the low latitudes. The time it takes for such advection to occur introduces a decadal lag in the tropical response to extra-tropical forcing, explaining the timescale of the variability (Harper, 2000). The spatial signature of the variability in the Pacific (Zhang et al., 1997) suggests a role for the ocean, but despite numerous efforts it has been very difficult to unequivocally identify the propagation of such anomalies from the subtropics to the equator (Deser and Blackmon, 1995; Schneider et al., 1999a,b).

In this paper, we have proposed an alternative perspective to the advective one explored in Gu and Philander (1997), involving tropical diabatic processes instead. Boccaletti (2004) proposed that basin wide diabatic mechanisms control the depth of the tropical thermocline over a timescale τ_d, associated with wind driven diabatic processes. This timescale, estimated in Eq. (4), is different from the adjustment timescale τ_w associated with the propagation of waves around the basin (Cane and Sarachik, 1979). It is a timescale that depends on the speed at which water masses can be converted by the surface forcing, and depends on the relationship between the tropical thermal structure, the tropical sea surface temperatures and the heat fluxes. In this paper, we have proposed that in response to periodic perturbations the ocean acts as a low-pass fitler, with characteristic timescale given by τ_d.

In order to illustrate this mechanism, we have forced a small idealized basin with a periodic forcing in the subtropics. By varying the frequency of the forcing over several experiments, we were able to show that the tropical ocean responds to this external forcing in a manner akin to a linear filter. The amplitude of the signal is heavily damped for periods of 1 or 2 years, but for decadal timescales it is roughly conserved to the equator. Despite the fact that, at each period, anomalous water masses are created and advected equatorward, the anomalies themselves are not coherently advected to the equator, but rather are dispersed in the process of adjustment. Waves, however, do propagate around the basin, carrying the information of the perturbation, and affecting diabatic processes in the region where heat is gained. In these simple experiments, a phase lag between the forcing and the response depends on waves; they introduce a time-scale τ_w of a few months and also depends on the thermodynamics which introduce a lag τ_d of a few years. According to Eq. (11) the propagation mechanism dominates at high frequencies, but at low frequencies it is the thermodynamic lag that dominates.

The diabatic timescale τ_d is the critical parameter for this model, and as shown in Eq. (4) that parameter depends on the mean state of the ocean. That in turn depends on the surface boundary conditions and is ultimately the result of a coupled process between the atmosphere and the ocean. Under different climatic conditions from the current ones, the mean state of the ocean might be substantially different and so might be the properties of the tropical decadal variability. Further experiments are necessary to determine the relationship of the diabatic timescale τ_d to the mean state.

Many factors that can contribute to the observed decadal variability have been ignored in this paper. Local variability in the Tropics has been ignored, and so has the atmospheric bridge between the Tropics and the extratropics. We have chosen to focus on a forced problem, although, ultimately, coupled process are likely to be an integral part of the story. In particular the variability of the wind in controlling upwelling (thorugh the term w_{Ek} which is here held constant), and its dependence on surface temperatures, will impact the solution. Finally, the size of the basin, which we have chosen small for compuational reasons, will have an impact on the timescale τ_w of the dispersive waves. All of these constraints need to be relaxed.

The results presented here imply that extra-tropical forcing can induce low frequency variability near the equator, even when there is no explicit coherent connection between low and high latitudes. A search for such connections may in fact be futile. A study of the oceanic heat budget should prove more profitable. Data sets that describe heat fluxes across the ocean surface, and changes in the oceanic heat content, will be of enormous value and will probably require the assimilation of measurements into a realistic model of the ocean. Recently, McPhaden and Zhang (2002) analyzed data that suggest a change in the meridional circulation of the upper tropical Pacific Ocean. That change, from the perspective of the results presented here, should be regarded as a symptom rather than the cause of a change in the oceanic heat budget. The processes responsible were presumably altered surface conditions that brought into play rapid waves that connect low and high latitudes, and diabatic processes.

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