Ocean-Atmosphere Coupling

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Abstract.

Ocean-atmosphere coupling is one of the major concepts of climate dynamics essential for a large number of climate phenomena including the El Niño-Southern Oscillation (ENSO) and decadal climate variability. Understanding ocean-atmosphere coupling is critical for predicting changes in global temperature patterns and properties of different modes of climate variability with global warming. This article reviews the main ideas relevant to the coupling and its role in climate.

1. Introduction

The focus of the present article is ocean-atmosphere coupling – one of the cornerstones of modern climate dynamics. Originating in the pioneering work of Bjerknes (1964, 1972) and Wyrtki (1973, 1974), the concept of ocean-atmosphere coupling has become essential for explaining properties of several climate phenomena ranging from the seasonal cycle in the tropics to the El Niño-Southern Oscillation (ENSO), to decadal climate variability. The idea behind active ocean-atmosphere coupling is straightforward: a large-scale anomaly of sea surface temperature (SST) induces diabatic heating or cooling of the atmosphere, which alters atmospheric circulation and hence the wind stress and heat fluxes at the ocean surface. In turn, the wind stress variations modify the ocean thermal structure and circulation, giving rise to a series of positive feedbacks that reinforce the initial SST anomaly[1]. One can no longer treat the ocean and atmospheric circulations independently of each other, and the SST serves as a link combining the two media.

The degree to which the ocean and atmosphere are coupled varies between different regions. In the tropics, for instance, the coupling is very strong because tropical wind stress is largely controlled by tropical SSTs. In mid-latitudes, however, atmospheric circulation depends on local SSTs to much weaker extent, which implies a weak dynamical coupling. The temporal and spatial scales of coupling vary a lot as well. For instance, ocean wind waves are generated by air-sea interactions at scales of O(1cm-100m). Tropical Instability Waves, propagating westward along the equator, are weakly coupled to the atmospheric flow above (Chelton *et al* 2000) and have spatial scales of O(1000km). ENSO and the seasonal cycle in the tropical Pacific involve the entire tropical basin and operate on timescales from annual to interannual. Here, we will concentrate only on such large-scale phenomena.

There are four questions that one needs to address when looking for active (also called dynamical) coupling: (a) What is the response of the atmosphere to a given SST anomaly?

[1] A broader definition of active coupling is sometimes used, which allows for the cases when only negative dynamical feedbacks from the ocean to the atmosphere are possible.

(b) What is the response of the ocean to a given wind stress anomaly? (c) Can the wind stress amplify the initial SST perturbation (e.g. are there positive feedbacks)? A correlation between the wind stress and the SST is not sufficient, since the atmosphere can simply drive the ocean. The fourth question is whether oscillatory behavior is possible in the system, which is usually related to ocean dynamics that provide slow timescales and potential mechanisms to reverse temperature tendencies.

2. Ocean-atmosphere coupling in the tropics.

In the tropics, sea surface temperatures have a profound effect on atmospheric circulation. Moist air rises into cumulus towers over the warmest regions, which therefore have heavy rainfall; aloft, the air that has been drained of its moisture diverges from these regions and subsides over the colder regions that get little precipitation. Surface winds, the easterly trades in the case of the Pacific, restore moisture to the air by means of evaporation while converging to the warmest regions. The convergence of moist air and the release of latent heat through condensation drive this direct thermal circulation. Changes in surface temperatures alter rainfall, winds, and other atmospheric variables.

A positive SST perturbation in the eastern equatorial Pacific results in anomalous diabatic heating of the atmosphere, anomalous air flow convergence to this region, and hence eastward wind anomaly that reduces the strength of the zonal westward winds. The local atmospheric response to the heating is baroclinic: for warm SST perturbations a low pressure anomaly is set in the lower troposphere and a high pressure anomaly in the upper. A simple reduced-gravity model is frequently used to describe the induced atmospheric circulation (Gill 1982, also Cane and Zebiak 1987, Battisti and Hirst 1989). Its solution consists of stationary atmospheric baroclinic Kelvin and Rossby waves surrounding the heat source. The effect of diabatic heating is much stronger in the tropics than elsewhere because the tropical atmospheric circulation is largely controlled by the release of latent heat supplied by evaporation from the warm ocean surface.

The next question is what determines tropical temperatures. Over much of the globe it is the radiation balance at the ocean surface. In the tropics, however, the dynamical response of the oceans to the winds becomes a major factor. This is because the warm surface waters of the tropics constitute only a shallow layer floating on the cold water below. The winds, by causing variations in the depth of the thermocline, can expose cold water to the surface. For example, intense trade winds along the equator normally drive the warm waters westward along the equator while bringing cold water to the surface in the east, thus contributing to the colder surface temperature in the eastern part of the basin (Fig. 1c). A relaxation of these winds allows water to flow eastward, reducing the zonal temperature gradient along the equator and creating positive feedbacks that give rise to El Niño - the warm phase of a natural oscillation driven by tropical ocean-atmosphere interactions and perhaps the most striking example of active coupling (Fig. 1b).

The explanation for El Niño and its cold counterpart La Niña, and for the mean zonal SST gradient along the equator (Dijkstra and Neelin 1995), involves a circular argument: changes in sea surface temperature are both the cause and consequence of wind fluctuations. Consider, for example, conditions during La Niña, when intense trade winds keep the warm surface waters along the equator in the far west, thus maintaining a zonal temperature gradient that contributes to the intense winds. A modest disturbance in the form of a westerly wind anomaly near the dateline can generate currents that transport some of the warm water eastward, thus decreasing the zonal temperature gradient. The resultant weakening of the trades will cause even more warm water to flow eastward, reducing the upwelling of cold water to the surface and setting a strong SST anomaly in the eastern equatorial Pacific.

One of the main positive feedbacks involved in this process, the wind-thermocline-SST feedback, is often referred to as Bjerknes feedback. Other relevant positive feedbacks include the wind-evaporation-SST feedback and the cloud-SST feedback (decks of low stratus clouds tend to form over coldest waters shielding them from excessive solar radiation and hence maintaining colder temperatures).

To explain the cyclicity of El Niño which occurs roughly every 4-5 years (Fig. 1a) one needs to take into account negative feedbacks associated with the ocean dynamics that involve a delayed ocean response to wind forcing. The delay in ocean adjustment is related to the propagation along the equator of oceanic Kelvin and Rossby waves of the opposite signs (downwelling and upwelling) that can reverse SST anomalies[2].

The positive feedbacks involved in ENSO are also crucial for maintaining the meridional asymmetry in the structure of tropical SSTs (Philander *et al* 1996, Xie and Saito 2001) and the seasonal cycle in the tropical Pacific (Chang and Philander 1994). The radiation forcing in the equatorial region is semi-annual since the sun crosses the equator twice in the course of one year; however, the seasonal cycle in the eastern and central equatorial Pacific is strictly annual. This is because temperature there is controlled mainly by dynamical factors, especially by the strength of upwelling of cold water to the surface. As a result, the SSTs are usually colder south of the equator, while the Inter-Tropical Convergence Zone (ITCZ) develops over the warmest waters that stay predominantly north of the equator (Fig. 1c).

Consider the winds crossing the equator from south to north over the eastern equatorial Pacific. These winds bring cold water to the surface in the vicinity of the equator in the Southern hemisphere, but depress the thermocline in the Northern hemisphere making surface warmer. Thus, these winds maintain the cross-equatorial temperature gradient which, in turn, maintains the strength of the meridional winds. The effect is reinforced by the wind-evaporation-SST and cloud-SST feedbacks (Chiang and Vimont 2001). Consequently, seasonal variations in incoming solar radiation lead to moderate meridional displacement of the Inter-Tropical Convergence Zone (ITCZ) and to annual variations in the meridional winds and equatorial upwelling which can counteract the effect of solar radiation.

The role of positive feedbacks and active ocean-atmosphere coupling for tropical basins other than the Pacific are still being debated. For instance, it has been suggested that decadal

[2] For further information on El Niño, we refer the reader to the relevant article in this volume, and to Philander (1990).

Tropical Atlantic Variability (e.g. Chang *et al* 1997, Barreiro *et al* 2005) and the Indian Ocean Dipole Mode (Saji *et al* 1999, Webster *et al* 1999) may have a role for active coupling. However, it is unclear whether these modes of climate variability actually correspond to oscillatory (perhaps damped) dynamical modes.

3. Ocean-atmosphere coupling in mid-latitudes

Ocean-atmosphere coupling in mid-latitudes remains a subject of intense research and even heated debates at times. It is believed that climate variability in mid-latitudes is determined, to a large extent, by stochastic atmospheric variability and passive oceanic response to the atmospheric forcing. Hasselman (1977) and Frankignoul and Hasselmann (1977) put forward a simple stochastic model that accounts for such physics:

$$\frac{dT}{dt} = -\frac{T}{t_o} + f(t) \qquad (1)$$

Eq. (1) is derived from a one-dimensional energy balance at the ocean surface between short- and long-wave radiation, and turbulent heat fluxes. T is the anomalous temperature of the ocean mixed layer (the upper layer of the ocean with vertically uniform temperature), t_o is the effective timescale for damping thermal anomalies by net heat fluxes in the absence of forcing (t_o~3 months), and the temporal derivative of T reflects ocean thermal inertia. The forcing term f(t) describes the atmospheric "noise" with an essentially white power spectrum. The ocean responds to this forcing by reddening the spectrum of temperature variations.

This model can be easily modified to include the effect of thermal (passive) coupling between the ocean and the atmosphere, which result in smaller energy surface fluxes at low frequencies because the ocean would have time to adjust to variations in air temperature and because heat fluxes are proportional to the air-sea temperature difference. The net result is a larger variance of both sea surface and air temperatures (Barsugli and Battisti 1998). The inclusion of passive coupling in Eq. (1) is equivalent to a reduction in the damping coefficient $1/t_o$.

While useful for explaining gross properties of climate variability in mid-latitudes, this stochastic model does not describe specific climate modes, such as the North Atlantic Oscillation (NAO) which dominates the climate of North Atlantic and Western Europe and is related to the position and strength of the storm track (e.g. Hurrell 1995, Deser and Blackmon 1993), and the Pacific Decadal Oscillation (PDO) which affects a large region extending from the North Pacific to the equatorial region to the Southern Hemisphere (Zhang *et al* 1997, Mantua 1999, Bond and Harrison 2000). The observational records for these modes show either weak spectral peaks (not always statistically significant) or an enhancement of the low-frequency part of the power spectrum that cannot be fully explained by the reddening of the spectrum due to ocean thermal inertia. Various mechanisms based on active ocean-atmosphere coupling in mid-latitudes have been proposed to describe the physics of these modes, even though the observational evidence for such coupling remains not as strong as in the tropics.

It is clear that atmospheric variability can induce significant large-scale SST anomalies in the ocean. In fact, ocean response to varying winds in mid-latitudes includes (a) changes in wind-induced upwelling, (b) changes in the depth of the thermocline and in ocean circulation, e.g. shifts in the paths of the Gulf Stream and Kuroshio caused by wind stress curl anomalies, and (c) variations in heat fluxes across the ocean surface caused mostly by changes in evaporation. The latter factor is essential - the patterns of wintertime anomalous SST in mid-latitudes resemble those of anomalous turbulent heat fluxes at the air-sea interface (Fig. 2). Because of ocean thermal inertia, the induced temperature anomalies can persist for several months or even years. Wintertime temperature anomalies in the ocean mixed layer can be shielded from the surface by the seasonal thermocline in summer and then reemerge on the surface during next winters (Alexander and Deser 1995).

It is also clear that the atmosphere can feel meso-scale temperature variability and temperature fronts in the ocean (e.g. Samelson *et al* 2006). For instance, SST anomalies associated with Tropical Instability Waves and even Gulf Steam rings modify the air flow above affecting the stratification of the atmospheric boundary layer. A change in stratification alters the height of the boundary layer and hence the surface wind stress. Neverthelss, the key unresolved question is whether a mid-latitude SST anomaly can induce a sufficiently strong large-scale atmospheric response that can amplify the initial temperature anomaly via positive feedbacks (observations suggest that on many occasions positive SST anomalies are quickly damped by negative heat fluxes).

A warm SST anomaly in mid-latitudes induces a diabatic heating of the atmosphere, but in contrast to the tropics, the immediate, baroclinic response of the atmosphere is weak (since the release of latent heat is small). Further atmospheric response includes an equivalent barotropic signal generated through interactions of the initial perturbation with the mean atmospheric flow and mid-latitudes storms (Peng and Robinson 2001, Peng *et al* 1997). In the Pacific ocean this process can eventually generate a pattern in sea level pressure similar to the Pacific-North American pattern (PNA), see Fig. 3. In the Atlantic ocean SST anomalies can generate sea level pressure anomalies resembling those associated with the North Atlantic Oscillation (Czaja and Frankignoul 1999, Sutton and Hodson 2003).

Changes in sea level pressure imply changes in the wind stress at the ocean surface. Could these wind stress anomalies feedback onto the initial SST anomaly and perhaps lead to decadal climate oscillations? Here we describe several conceptual models that answer positive to this question.

A broad class of such coupled models involve interactions between the ocean subtropical gyre circulation and the atmosphere (e.g. Weng and Neelin 1998, Cecci 2000, Primeau and Cessi 2001). Consider, for example, the model proposed by Latif and Barnett (1994, 1996) who argued that decadal climate fluctuations in the North Pacific are sustained by unstable interactions between the subtropical gyre and the Aleutian low-pressure system. When the subtropical ocean gyre is anomalously strong, more warm tropical waters are transported poleward by the Kuroshio and its extension, leading to a positive SST anomaly in the North Pacific (Fig. 4b). The atmospheric response to this anomaly involves a weakening of the Aleutian low-pressure system. At the same time, the associated heat fluxes at the air-sea interface, reduced ocean vertical mixing, and anomalous northward Ekman heat transport reinforce the initial SST anomaly (a positive feedback). The atmospheric response also consists of a wind stress curl anomaly that spins down the ocean subtropical gyre, thereby reducing the poleward heat transport and the initial SST perturbation (a negative feedback). The ocean adjusts with a lag to changes in the wind stress, which allows continuous oscillations. The time lag is related to the propagation across the basin of oceanic baroclinic Rossby waves, which are an important element of ocean adjustment to wind forcing.

Observations indicate, however, that the initial SST anomaly is usually dampened by the air-sea heat fluxes, rather than amplified. As an alternative, other authors suggested that the North Pacific decadal variability could be forced by stochastic Ekman pumping from the atmosphere and no active coupling was necessary (Seager *et al* 2001, Schneider *et al* 2002). The dynamical response of the ocean to stochastic forcing includes a slow adjustment via Rossby waves, so that the ocean acts as a low-pass filter that effectively integrates the high-frequency wind stress forcing, reddening the spectrum of temperature variations.

A number of models for decadal climate variability invoke a connection between mid-latitudes and the tropics. This connection can be established through the ocean shallow meridional overturning, usually referred to as the Subtropical Cell or STC (e.g. Liu *et al* 1994, McCreary and Lu 1994, McPhaden and Zhang 2002, Nonaka *et al* 2002). For example, Gu and Philander (1997) proposed that after subduction into the STC extra-tropical temperature anomalies are advected towards the equatorial region. Several decades later they reemerge in the eastern equatorial Pacific, modify the atmospheric circulation, and induce temperature anomalies of the opposite sign remotely in the extra-tropics (Fig. 4a), which leads to decadal climate oscillations. However, subsequent studies showed that the magnitude of the temperature anomalies, when they reach the equatorial region, is too weak (Schneider 1999, Hazeleger *et al* 2001), and that changes in the strength of the overturning caused by wind stress variations are more important than the advection of temperature anomalies (Solomon *et al* 2003).

The importance of ocean-atmosphere coupling in the North Atlantic is also a subject of ongoing debate. One of the key questions is whether variations in the position and strength of the Gulf Stream can affect the North Atlantic Oscillation. Marshall *et al* (2001), for example, proposed a conceptual model that coupled surface Ekman layers, ocean gyres, and ocean thermohaline circulation (THC) to the atmospheric jet stream. Meridional shifts in the zero wind stress curl line drive anomalies in gyre circulation, and north-south dipoles in air-sea fluxes drive anomalous thermohaline circulation. Both gyre and thermohaline circulations modulate SSTs and hence the overlying atmospheric jet stream.

The model identifies nondimensional parameters that control whether the ocean responds passively to NAO forcing or actively couples to the NAO. Yet, which regime better describes the actual ocean-atmosphere system in the North Atlantic is unclear, especially in light of the arguments that the NAO is not a dynamical oscillation but, for the most part, a stochastically driven phenomenon (e.g. Vallis *et al* 2004). It is also unclear whether multidecadal changes in the THC, predicted by general circulation models (Delworth and Greatbatch 2000, Dong and Sutton 2005), are indeed coupled to the NAO.

4. Ocean-atmosphere coupling and climate change.

The past two decades show a dramatic progress in our understanding of and ability to model Earth’s climate. Studies of ocean-atmosphere interactions have been instrumental in this progress, and yet we still do not fully understand the role of these interactions in climate change. In recent years, we have experienced the most intense and devastating El Niño episodes (1982 and 1997) in more than a century. An extraordinary heat wave occurred in

France in 2003. The year 2005 showed the most active Atlantic hurricane season on record. Since the 1970s, the NAO index characterizing climate of mid-latitude has stayed predominantly positive, which occurred on the background of persistent global warming trends over the second half of the 20^{th} and the beginning of the 21^{st} century. These factors raise a number of important questions regarding the characteristics of ocean-atmosphere interactions in a changing climate.

One question, for instance, concerns the connection between the NAO and global warming. Hoerling *et al* (2001) suggested that the North Atlantic climate change since 1950 is linked to a gradual warming of tropical SSTs over the Indian and Pacific oceans. Changes in ocean temperatures modified rainfall patterns and diabatic heating of the atmosphere in low latitudes, which affected the North Atlantic Oscillation remotely and forced it into the positive phase during the past half-century. This result, however, contradicts the study of Rodwell *et al* (1999) who showed that much of the decadal variability of the winter North Atlantic Oscillation over the same time interval can be reconstructed from the knowledge of North Atlantic sea surface temperatures. In contrast, other authors (e.g. Wunsch 1999) argue that to the zeroth-order approximation the observed NAO index exhibits a typical behavior of a red noise process.

Another question concerns the severity and frequency of hurricanes in a warmer climate, especially in the context of feedbacks that the hurricanes could provide for the climate system by affecting ocean vertical mixing and poleward heat transport (Emanuel 2001, Sriver and Huber 2007). Although there have been suggestions that the potential intensity and impacts of hurricanes have been increasing over the past half-centuary (e.g. Emanuel 2005), the entire subject remains highly controversial.

Possible changes in the properties of El Niño are another important issue (e.g. Fedorov and Philander 2000, 2001; An and Jin 2000, Wang and An 2001). Does the occurrence of strong El Niño events in the last thirty years provide an omen of what to expect as global temperatures continue to rise? Can the climate system shift towards continuous, or permanent, rather than intermittent El Niño events (imagine conditions similar to El Niño of 1997 lasting for many, many decades)? Proxy temperature records show that the climate system was locked into a permanent El Niño state during the early Pliocene, between approximately 3 and 5 million years ago, when the Earth was experiencing greenhouse conditions similar to today's (e.g. Fedorov *et al* 2006). A permanent El Niño would imply very different characteristics of ocean-atmosphere coupling.

Unfortunately, the modeling results on changes in the properties of ENSO and other climate modes with global warming are not yet very reliable – differences between climate models are much larger than changes predicted by each particular model in greenhouse warming simulations (e.g. Collins *et al* 2005, Guilyardi 2006, Yeh and Kirtman 2007). Similarly, projections for changes in the mean state of the climate system vary greatly from one model to the next. While many such problems remain unresolved, the continuing progress in our understanding of ocean-atmosphere interactions, and greater computational power available for researchers, are very encouraging.

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Figures:

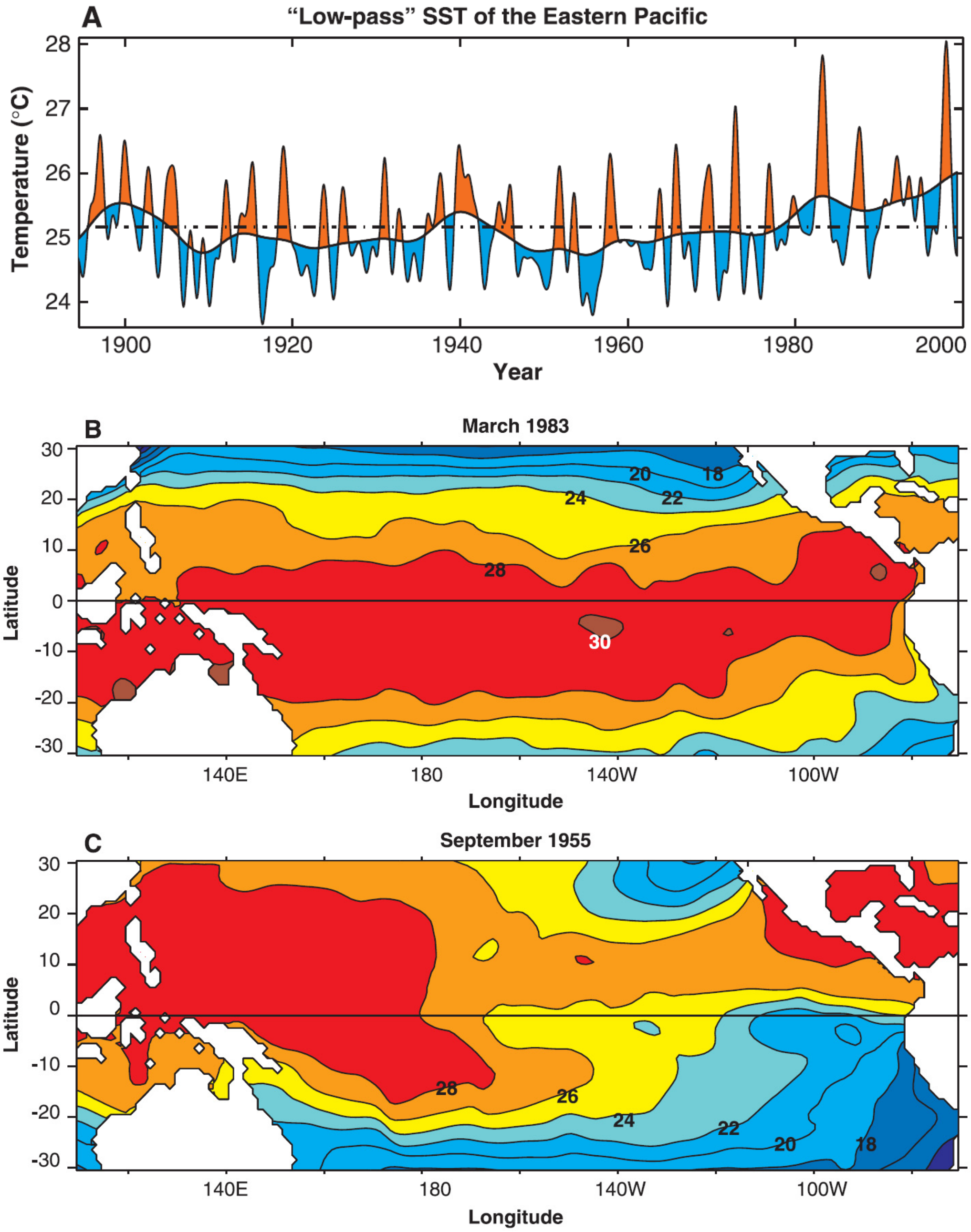

Fig. 1: (A) Interannual oscillations in sea surface temperature (SST) at the equator in the eastern Pacific shown on the background of the decadal fluctuation after removal of the annual cycle and higher frequency variability. The horizontal dot-dashed line is the time average for the record. Sea surface temperatures (^{o}C) at the peaks of El Niño (B) and La Niña (C). Some of the differences between (B) and (C), such as high temperatures that extend far north in September and far south in March, are attributable to the seasonal cycle. The mean state of the tropical Pacific resembles a weak La Niña. After Fedorov and Philander 2000.

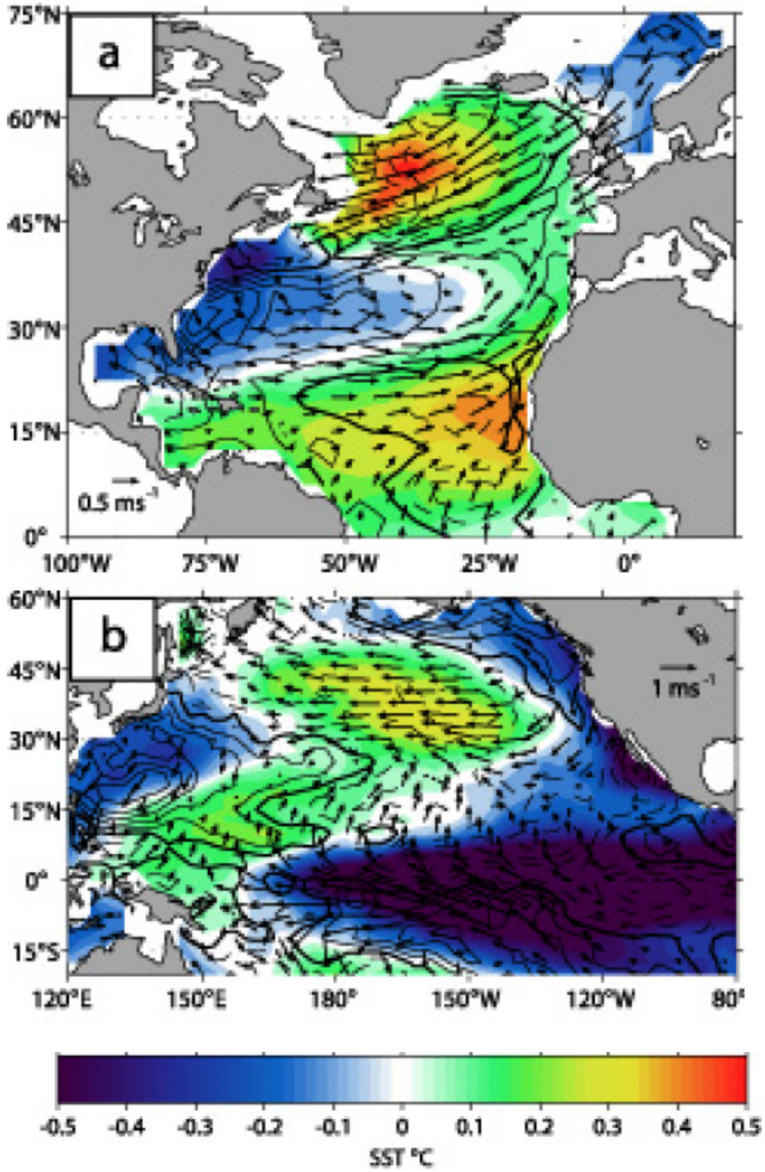

Fig. 2: The patterns of wintertime (December-March) anomalous SST, ocean-to-atmosphere turbulent heat flux (latent plus sensible), and surface wind vectors, associated (via linear regression) with the leading principal component of SST variability in the North Atlantic (a) and North Pacific (b). Data are for the period 1949 to 1999 from the NCEP/ NCAR reanalysis. Heat fluxes are in W/m^2 with positive/negative values in solid/dashed contours every 3 W/m^2 and a thick zero contour. Arrows depict the wind vectors in m/s with the scale shown in (b). The temperature values, in ^{o}C, are denoted in colors according to the scale (note that the scale is kept at the –0.5 to 0.5 °C range for clarity, however, values in eastern equatorial Pacific extend up to 1.2 °C. After Kushnir 2002.

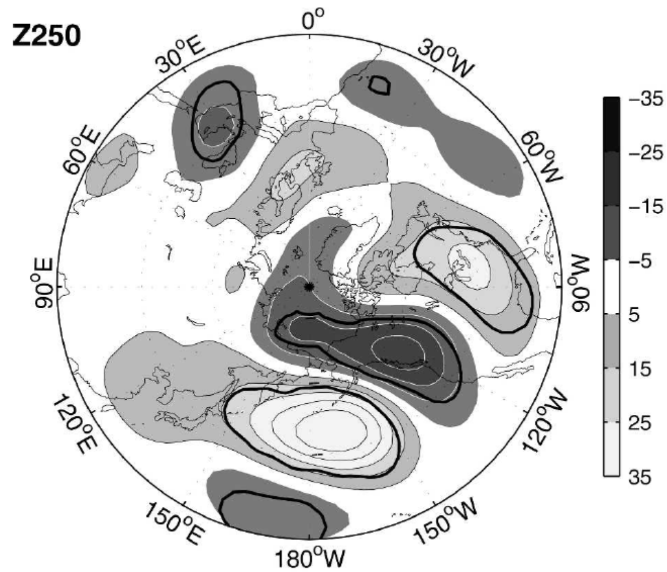

Fig. 3: A typical pattern of the atmospheric response to a large-scale extra-tropical SST perturbation. The figure shows a regression of winter anomalies of geopotential height at 250mb (in m/^{o}C) onto the timeseries of SST anomalies in the Fall. SST anomalies lead geopotential height anomalies by two months. After Frankignoul and Sennéchael 2007.

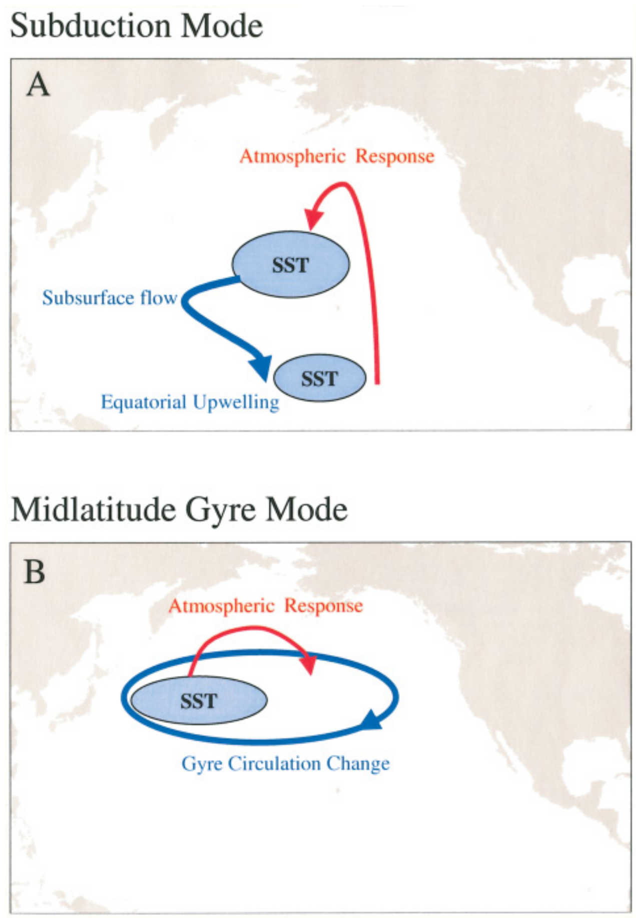

Fig. 4: A schematic description of (A) the subduction mode and (B) the mid-latitude gyre mode proposed to explain decadal climate variability in the Pacific. The subduction mode involves a mid-latitude SST anomaly that subducts into the thermocline, slowly travels towards the tropics, and then upwells in the eastern equatorial Pacific, forcing an SST anomaly of opposite polarity in mid-latitudes. The gyre mode involves the atmosphere responding to a mid-latitude SST anomaly in the western part of the basic, which is driven by a delayed response of the subtropical gyre circulation to antecedent atmospheric forcing with opposite polarity. After Miller *et al* 2003.