On the Decadal and Interdecadal Variability in the Pacific Ocean

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ABSTRACT

The Pacific decadal and interdecadal oscillation (PDO) has been extensively explored in recent decades because of its profound impact on global climate systems. It is a long-lived ENSO-like pattern of Pacific climate variability with a period of 10–30 years. The general picture is that the anomalously warm (cool) SSTs in the central North Pacific are always accompanied by the anomalously cool (warm) SSTs along the west coast of America and in the central east tropical Pacific with comparable amplitude. In general, there are two classes of opinions on the origin of this low-frequency climate variability, one thinking that it results from deterministically coupled modes of the Pacific ocean-atmosphere system, and the other, from stochastic atmospheric forcing. The deterministic origin emphasizes that the internal physical processes in an air-sea system can provide a positive feedback mechanism to amplify an initial perturbation, and a negative feedback mechanism to reverse the phase of oscillation. The dynamic evolution of ocean circulation determines the timescale of the oscillation. The stochastic origin, however, emphasizes that because the atmospheric activities can be thought as having no preferred timescale and are associated with an essentially white noise spectrum, tne ocean response can manifest a red peak in a certain low frequency range with a decadal to interdecadal timescale. In this paper, the authors try to systematically understand the state of the art of observational, theoretical and numerical studies on the PDO and hope to provide a useful background reference for current research.

Key words: climate system, Pacific decadal oscillation, deterministic origin, stochastic origin

1. Introduction

It is well known that the tropical Pacific atmosphere- ocean system manifests predominantly as the El Niño-Southern Oscillation (ENSO) cycle with periods of 1 to 5 years, and that the North Pacific atmosphere-ocean system fluctuates with the same periods in response to the ENSO cycle (e.g., Bjerknes, 1969; Horel and Wallace, 1981; Neelin, 1990; Philander, 1990). Recent observational and numerical studies show that the first EOF mode of SST in the North Pacific is the decadal mode, while in the tropical Pacific, although the first EOF mode is the ENSO mode, the second EOF mode has a decadal timescale in response to the North Pacific decadal mode (e.g., Trenberth, 1990; Latif and Barnett, 1994; 1996; Barnett et al., 1999a, b; Mantua et al., 1997; Zhang et al., 1997). That is, the Pacific atmosphere-ocean system has a coherent decadal scale fluctuation, which is commonly called the Pacific Decadal Oscillation (PDO)** .

The PDO is a long-lived ENSO-like pattern of Pacific climate variability (Mantua et al., 1997; Zhang

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**Actually, it is argued that the decadal variability in the Pacific has at least two components (Barnett et al., 1999a, b): a stochastic mode and a deterministic mode. The former is stochastically forced by the atmosphere, where the spectra of the atmospheric forcing are white but the forced oceanic response has a red spectrum with a peak in the decadal timescale. This mode is simply the passive ocean response to atmospheric forcing. The latter is associated with a deterministically forced coupled mode of the Pacific ocean-atmosphere system, namely, the PDO. In this paper, we generally call the decadal scale stochastic mode and the coupled mode as the Pacific Decadal Variability (PDV); and we focus more on the coupled mode.

et al., 1997). While the two climate oscillations have similar spatial climate fingerprints, they have very different behavior in time. Fisheries scientist Steven Hare coined the term "Pacific Decadal Oscillation" in 1996 while researching connections between Alaska salmon production cycles and Pacific climate. Two main characteristics distinguish PDO from ENSO: first, 20th century PDO events persist for 20 to 30 years, while typical ENSO events persist for 6 to 18 months; second, the climatic fingerprints of the PDO are most visible in the North Pacific/North American sector, while secondary signatures exists in the tropics; the opposite is true for ENSO (Giese and Carton, 1999).

Observational studies have identified two full PDO cycles in the past century from the historical record: "cool" PDO regimes prevailed from 1890–1924 and again from 1947–1976, while "warm" PDO regimes dominated from 1925–1946 and from 1977 through (at least) the mid-1990s (Mantua et al., 1997; Minobe, 1997). Minobe (1999) has shown that 20th century PDO fluctuations were most energetic in two general periodicities, one from 15 to 25 years, and the other from 50 to 70 years. The global ice SST (Jones, 1994; Parker et al., 1995) and some biological evidences (Hare and Mantua, 2000) reveal that the most dramatic climate regime shift occurs around 1976/77. This regime shift is clearly the strongest with a peak-to-peak amplitudes that exceeds 0.75°C, while the 1924/25 and 1946/47 climate shifts are more modest with peak-to-peak amplitudes on the order of 0.4°C (Chao et al., 2000; Hare and Mantua, 2000).

The general picture for climate shift of the Pacific atmosphere-ocean system can be depicted as follows: when PDO transits from its 'cool' regime to 'warm' regime, the SSTs in the central North Pacific vary from anomalously warm to anomalously cool with the opposite transition occurring along the west coast of America and in the central east tropical Pacific. Correspondingly, the PDO sea level pressure (SLP) anomalies vary from high pressure to low pressure over the North Pacific, and from low to high over western North America and the subtropical Pacific. These pressure pattern changes cause enhanced counterclockwise wind stress over the North Pacific. Figure 1 shows the cold phase and warm phase PDO surface climate anomalies as well as the PDO index. When the SSTs are anomalously cool in the central North Pacific, the SLPs are below average over the same region. The negative (positive) PDO index (Fig. 1c) indicates a cold (warm) phase, covarying with a positive (negative) SLP index (Fig. 1d). The most notable feature for these indices is their tendency for year-to-year persistence, with positive/warm or negative/cool index values tending to prevail for 20–30 year periods. However, within the 20–30 year regimes there are several short-lived sign reversals in the indices, including 3-year reversals from 1959–1961 and again from 1989–1991 (Mantua et al., 1997).

It has taken decades for us to merely recognize the 1976/77 regime shift, despite the strength and scope of the changes initiated by this regime shift. We are still far away from fully understanding the mechanism of PDO because of its far reaching consequences not only for the large marine ecosystem of the North Pacific (Hare and Mantua, 2000), but more importantly, for the global climate variability. If the PDO can be viewed as a background state for ENSO, the interaction between PDO and ENSO may modify the strength and period of ENSO and in turn, cause PDO to be modulated. In addition, PDO may have a substantial influence on the highly disputed issue of global warming that has a longer time scale. That is why the PDO attracts so much attention and has been extensively explored. Moreover, because of its complexity that involves it in almost every aspect of the climate system, it would take decades to achieve the final understanding of its mechanism and to predict its occurrence.

2. Observations

The well-known 1976/77 regime shift in the Pacific climate makes the PDO transit from its 'cool' regime to 'warm' regime. The annual SST difference between 1977–88 and 1966–76 (Fig. 2a) is more than 0.6° cooler in the central North Pacific, and about same magnitude warmer in the Northeastern Pacific and central eastern tropical Pacific, covarying with a deeper Aleutian low pressure system and stronger extratropical westerlies across the central North Pacific, enhanced southward flow along the west coast of North America, and stronger subtropic and subarctic gyre circulation in North Pacific (Deser et al., 1996; White and Cayan, 1998). This pattern is approximately reflection symmetric about the equator, slightly distorted by the shape of the basin (Chao et al., 2000; White and Cayan, 1998). The SST in the central South Pacific exhibits the same sign and comparable magnitude with that in central North Pacific. Figure 2 also illustrates that the central North Pacific SST (Fig. 2b) has a simultaneously negative correlation with the tropical SST (Fig. 2c).

The corresponding variation in the upper ocean thermal structure from the surface to 400-m depth in the central North Pacific shows a series of cold pulses beginning in the fall of 1976 that appear to originate at the surface and descend with time into the main

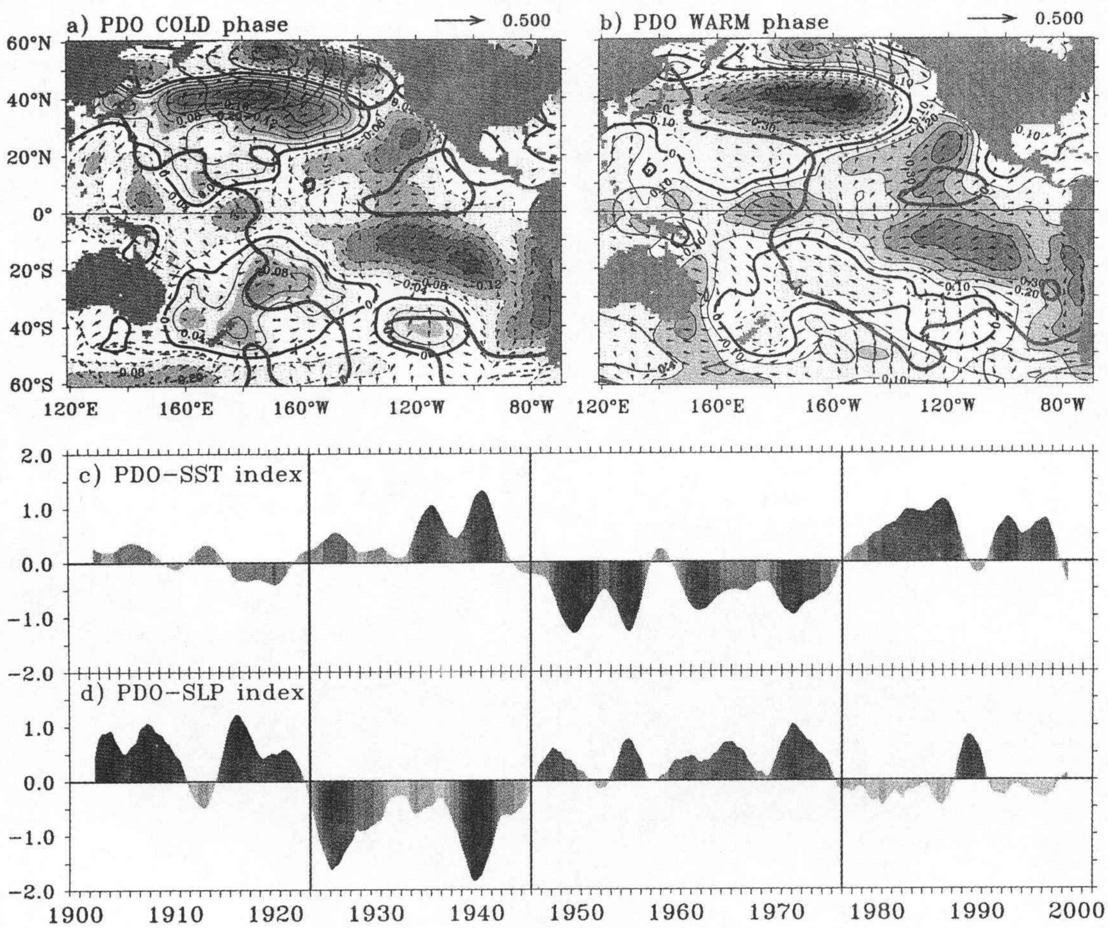

Fig. 1. (a) Cold phase and (b) warm phase PDO surface climate anomalies, (c) PDO index, and (d) North Pacific sea level pressure (SLP) index. Shaded contours represent SST anomaly (°C), contour lines represent SLP (hPa) and vectors represent surface wind stress (dyn cm^{-2}). The PDO index is defined as the leading principal component of the North Pacific monthly SST anomaly (poleward of 20°N). The SLP index is an area- averaged monthly SLP anomaly of the North Pacific (poleward of 20°N). A 5-year running mean is applied to the PDO index and the SLP index. Data used in (a) and (b) are from COADS. The PDO index and SLP index are obtained from Mantua (1997).

thermocline to at least 400-m depth (Fig. 3). Individual cold events descend rapidly ($\sim$100 m yr^{-1}), superimposed upon a slower cooling ($\sim$ 15 m yr^{-1}) by the 'envelope' of these cold pulses. Each of the cold pulses corresponds to a period of enhanced westerlies (Deser et al., 1996). The interdecadal climate change, while evident at the surface, is most prominent below $\sim$ 150 m where interannual variations are small. Unlike the central North Pacific, the temperature changes along the west coast of North America appear to be confined to approximately the upper 200–250 m (Deser et al., 1996).

The "slow" component of cooling in the central North Pacific is what really accounts for the decadal change in the Pacific (Deser et al., 1996). The temperature anomalies for three consecutive 5-yr periods (1977–81, 1982–86, and 1987–91) as a function of depth and latitude for longitude band 180°–145°W show a continuous southward and downward movement along isopycnals during these years (Fig. 4). In the early stage of cooling, the largest negative temperature anomalies (less than –0.2°C) are found in the upper 150 m between 30°N and 40°N. The cooling detached from the surface by the last pentad and penetrated into the upper portion of the permanent pycnocline in the subtropical gyre. This "slow" cooling moves downward and southward to at least 400 m depth at a speed of $\sim$15 m yr^{-1} and $\sim$ 0.25 cm s^{-1} respectively, which is roughly consistent with the speed derived from the mean curl of the wind stress 25–40 m yr^{-1} and 0.2–0.4 cm s^{-1} respectively (Huang and Qiu, 1994).

More intensive observational analysis of the temperature in the upper 400 m reveals that the decadal signal actually propagates in the thermocline along lines of constant potential vorticity(PV)(or isopycnals) from the ventilation zone in the central North Pacific to approximately 18°N in the western Pacific (Fig. 5)(Schneider et al.,1999a; Wang and Liu,2000). The southward speed of the thermal signal is approximately 7 mm s^{-1}, which yields a transit time of approximately eight years. The propagation path and speed

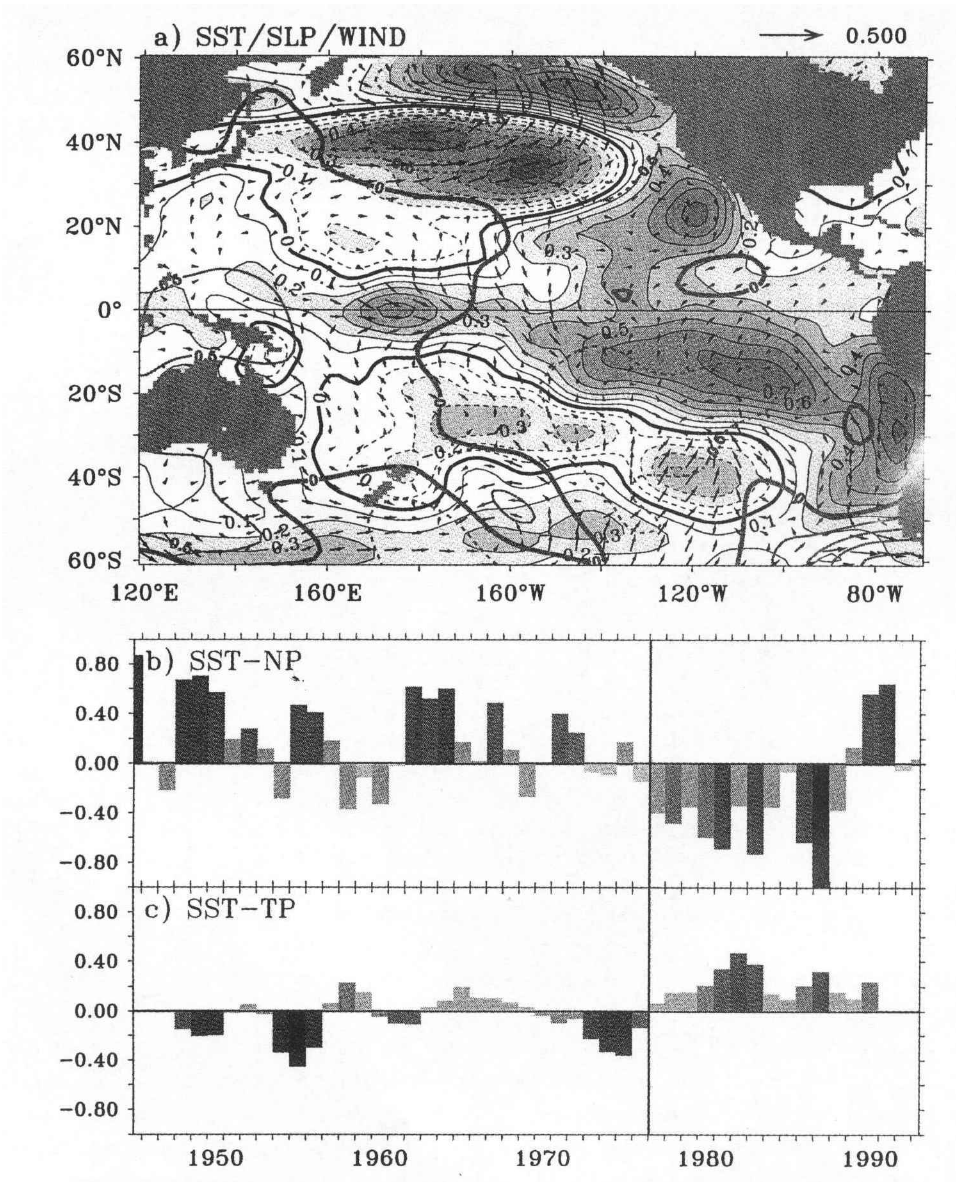

Fig. 2. (a) Annual surface climate differences between 1977–88 and 1966–76, and time series of annual SST anomalies for (b) central North Pacific (28°–44°N, 178°–146°W) and (c) tropical Pacific (6°S–6°N, 180°–90°W). In (a) shaded contours represent SST difference (CI=0.1°C), light contours represents SLP difference (CI=0.5 hPa) and vectors represent surface wind stress difference (dyn cm^{-2}). In (b) and (c) a 5-year running mean is applied. The thick line in (b) and (c) labels the phase shift year 1976/77. Data used in (a), (b), and (c) are from COADS.

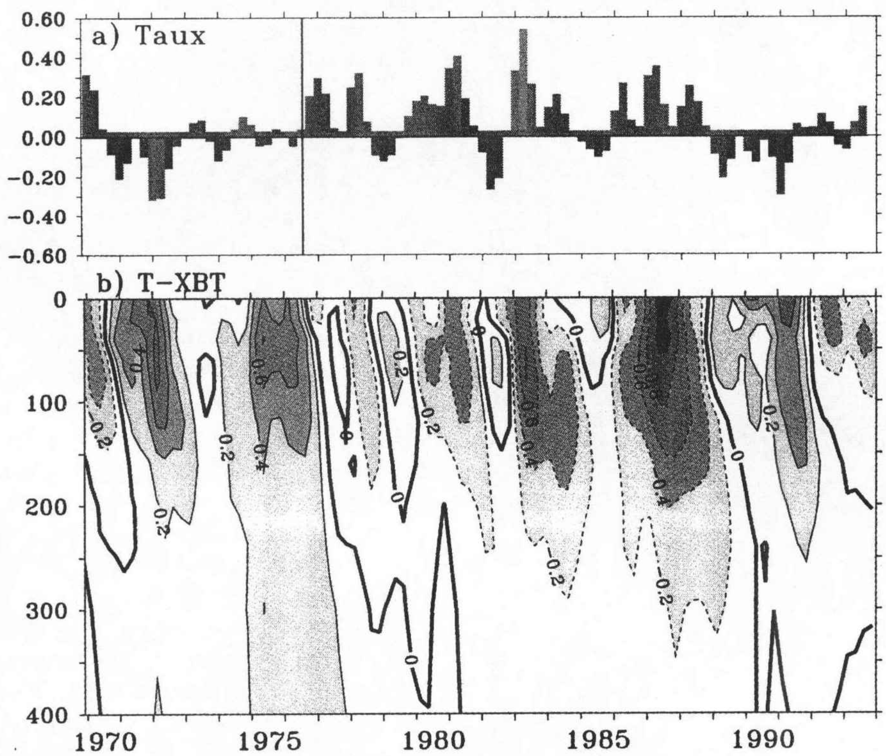

Fig. 3. (a) Seasonal zonal wind stress anomalies and (b) time-depth plot of seasonal anomalous temperature in the central North Pacific. The contour interval is 0.2°C in (b) and data is from XBT. In (a) positive (negative) values denote stronger (weaker) than normal westerlies (after Figure 7b in Deser et al., 1996).

can be well described by the geostrophic mean circulation and by ventilated thermocline theory. The thermal anomalies appear to be forced by perturbations of the mixed layer heat budget in the subduction region of the central North Pacific east of the date line. Warm anomalies (positive depth anomalies) were generated in the central North Pacific during the early 1970s and subducted southward to 18°N around 1982 (Fig. 5b). After 1978, a succession of colder winters initiated a cold anomaly (negative depth anomaly) in the central North Pacific that propagated along a similar path with a similar speed as the warm anomaly and then arrived in the western tropical Pacific at 18°N around 1991.

Besides the temperature, the salinity in the North Pacific also appears to have decadal changes, though there are fewer observations. Overland et al. (1999) identify a decadal variation of the salinity in the North Pacific by analyzing station observations. The decrease in density and increase in stratification in the Gulf of Alaska after 1977 corresponds primarily to a decrease in salinity in the upper 150 m. They found that while the PDO has an east/west character in temperature, the salinity signature would have a north-northwest/south-southeast character, similar to the pattern of interannual variability in precipitation. Since the salinity records do not indicate a similar spatial pattern of temperature decadal change, therefore, they hypothesize that North Pacific precipitation, acting in concert with other factors such as coastal runoff and wind-driven Ekman currents, influences the salinity decadal variability in the North Pacific. Besides these station observations on salinity in the northeastern Pacific, there is another observed salinity decadal change from the Hawaii Ocean Time-series (HOT) program at station ALOHA, which is located 100 km north of Oahu, Hawaii at 22°45'N, 158°W. Based on this observation, Lukas (2001) discloses a pronounced freshening (~0.015%) and cooling (~0.5°C) of the upper pycnocline (100–300 m depth) starting in 1991 and continuing through 1997. The freshening appears progressively later on deeper isopy-

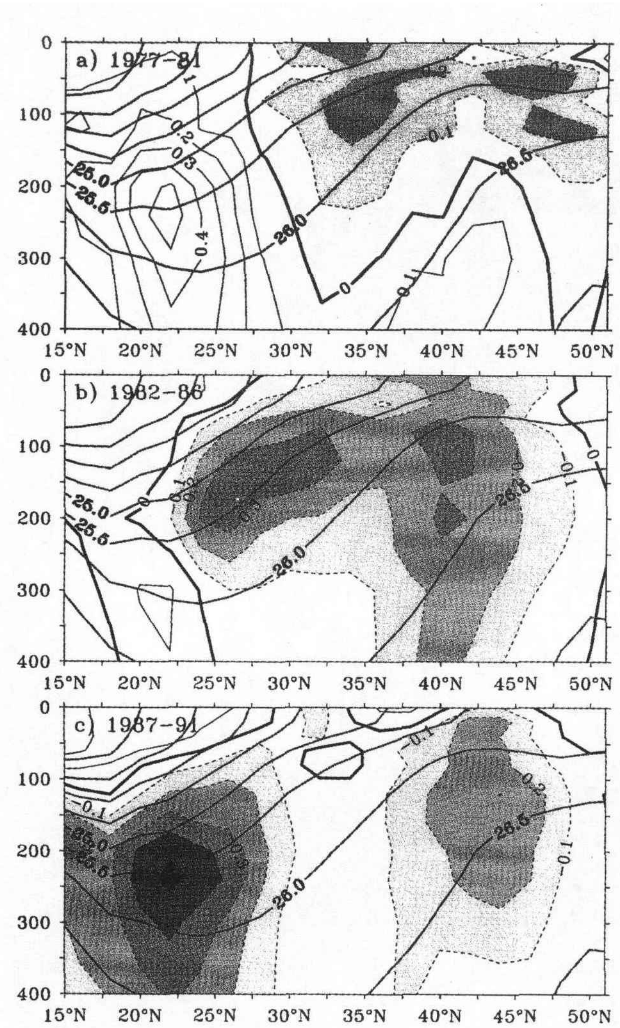

Fig. 4. Latitude-depth plot of annual temperature anomalies for longitude band 180°–145°W for (a) 1977–81, (b) 1982–86, and (c) 1987–91, superimposed upon the climatological isopycnals (σ_t, kg m^{-3}, thick light solid lines). The contour interval is 0.1°C for temperature and 0.5σ_t for density. One may notice that the temperature anomalies tend to move downward and southward along 25.5σ_t (after Figure 10 in Deser et al., 1996).

cnals as expected from thermocline ventilation theory. However, it is not clear whether and how this salinity decadal change is related to the PDO.

In general, the observations on PDO are very poor in temporal span and spatial coverage as well as resolution. For example, the global ice SST (Jones, 1994; Parker et al., 1995) has more than 150 years of data but its resolution is coarse (5°× 2°). The XBT data (White and Cayan, 1998) only includes monthly temperature above 400 m with a time span from 1955 to 1994, and is not globally covered. Therefore, more observations are extremely needed for studying the PDO detailed structure and mechanism. People are looking for longer historical records from geological and biological evidences (e.g., tree rings) on the one hand, and using numerical models on the other hand, for better understanding of the PDO mechanisms.

3. Mechanisms

The PDO is a highly nonlinear physical phenomenon that is involved in very complicated physical processes of ocean-atmosphere interaction and extratropical-tropical teleconnection. Most of the present studies on PDO or PDV eventually focus on the following three questions (Saravanan and McWilliams, 1997): (1) What and where does the variability originate from: atmosphere processes, oceanic processes, or coupled ocean-atmosphere interactions, extratropics or tropics? (2) What determines the timescale of such variability? and (3) What determines the spatial structure of such variability? Unfortunately, these questions are still not fully answered for the time being, although, to some extent, we can explain the PDV by linear, simple conceptual models (e.g., Gu and Philander, 1997) or nonlinear, complex primitive equation coupling models (e.g., Barnett et al., 1999b). In general, there are two classes of opinions; one that thinks this low-frequency climate variability results from deterministically coupled modes of the Pacific ocean–atmosphere system (Timmermann et al., 1998), the other from stochastic atmospheric forcing.

3.1 *Deterministic origins*

3.1.1 *Latif and Barnett paradigm*

In the earlier studies on PDO, Latif and Barnett (1994) proposed an atmosphere-ocean interaction paradigm (L-B paradigm, hereafter) trying to explain the low frequency oscillation in the North Pacific (north of 25°N), based on their coupled ocean-atmosphere model and observations. Their paradigm includes three indispensable processes which setup an oscillation in the ocean-atmosphere coupling system: a positive feedback to stir up an initial perturbation, a time delay to setup the timescale of the oscillation, and a negative feedback to make the system return from one phase to the other. These processes are depicted as follows.

In the central North Pacific, the characteristic SST anomaly pattern exhibits either a reduced or enhanced meridional SST gradient. Suppose the coupled system is in its reduced meridional SST gradient state. Such a distribution of SST would result in a northward shift of the baroclinic eddy activity in the atmosphere, leading to a weakened Aleutian low and subsequently reduced

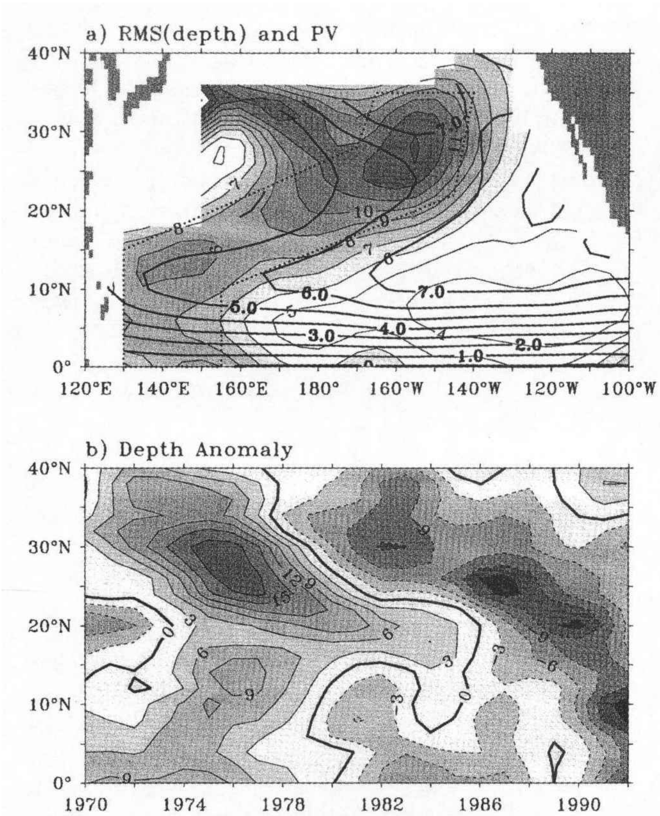

Fig. 5. (a) Rms values of depth anomalies averaged for 25–26σ_t isopycnals, superimposed upon the mean potential vorticity (PV). The contour interval is 1 m for rms of depth anomalies (shaded contours) and $1\times10^{-10}\mathrm{m}^{-1}\ \mathrm{s}^{-1}$ for PV (thick solid lines). The dotted lines in (a) outline the subduction pathway for depth anomalies, where the rms has a maximum and goes along constant PV. Panel (b) is the time-latitude plot of depth anomalies along the subduction pathway with a contour interval of 3 m.

westerly winds over the mid-latitude ocean. The associated changes in the net surface heat flux are such that they tend to reinforce the SST anomalies over most of the North Pacific. Furthermore, because the westerlies are weakened over a warm SST anomaly, surface mixing in the ocean is also reduced, which tends to strengthen the initial SST anomaly. Thus, the ocean and the atmosphere form a positive feedback system capable of amplifying an initial disturbance, giving rise to instability of the coupled system.

The corresponding change in the ocean is that the gyre circulation is reduced due to reduced westerlies or Aleutian low. Both the southward Sverdrup transport and the northward Kuroshio current are reduced, which eventually enhance meridional SST gradients. This is negative feedback. The ocean responds to the changing wind stress forcing by means of a westward propagating planetary wave, and this determines the timescale of the oscillation.

The L-B paradigm suggests a self-maintaining mechanism inherent to the central North Pacific climate system for the decadal variability, which appears to be supported by observations (Nakamura et al., 1997). In particular, they found that during the coolest period between 30°–40°N in the mid-1980s, the cool SST anomalies are reinforced by the enhanced surface westerlies associated with the intensified Aleutian low, while the warm SST anomalies around 1970 are strengthened by the reduced westerlies. However, in their later studies, they further point out this self-maintaining coupled mode (i.e., PDO) may only account for one third of the low frequency climate variability in the North Pacific, where most of the variability may arise from stochastic forcing by atmosphere (discussed in a later section) (Barnett et al., 1999b). Based on their modeling, they also conclude that the PDO does not directly rely on and impact ENSO, in other words, there is no manifest connection between PDO in the midlatitudes and ENSO in the tropics, which excludes the ENSO-caused PDO and PDO-affected ENSO. As for the highly *simultaneous* correlation between the SST time series in the central North Pacific and NINO12 (−0.63 for their model and −0.68 for observations) (Barnett et al., 1999a; Pierce and Barnett, 2000), they argue that it is the *stochastic* component of midlatitude Pacific decadal climate variability that imposes itself on the tropical Pacific via the atmospheric process, and not oceanographic, because the oceanic process is too slow to cause simultaneous response in two well-separated geometry regions.

In the L-B paradigm, the origin of PDO-type variability is fundamentally atmospheric. The functions of ocean in their coupled system are, first of all, to act as negative feedback media to reverse the phase of oscillation; second, to set up the timescale of oscillation, which is actually determined by the advection time of oceanic circulation (about 20 years) and/or, by planetary wave traveling time in the North Pacific. However, they focus only on the surface oceanic process and the first baroclinic planetary wave, and do not take into account oceanic thermocline dynamics. In fact, the decadal variability caused by buoyancy forcing and wind forcing in the central midlatitude ventilation region can propagate downward and equatorward by subduction flow along isopycnals, and eventually affect tropical ocean variability (e.g., Zhang et al., 1998). Therefore, the connection between the midlatitudes and tropics in the decadal timescale through the oceanic bridge gives rise to another PDO paradigm to be discussed below.

3.1.2 *Gu and Philander paradigm*

Gu and Philander (1997) proposed a simple conceptual model (G-P paradigm, hereafter) that can give rise to continual interdecadal oscillations in the whole Pacific. This paradigm involves teleconnection in the atmosphere and ocean between the extratropics and tropics. Initially assuming that a warming in the North Pacific occurs due to reduced overlying westerlies in the higher latitudes, it can be carried as an influx of warm waters from the extratropics to the tropics by subduction flow and cause a warming tropics. 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 (negative feedback). The cold water pumped downward in those regions arrives in the tropical thermocline a dozen years later, halts the warming and initiates a cooling in the tropics. The extratropical wind is weakened in response to the cold temperature in the tropics and causes the SST to increase in the region where surface waters subduct. Hence, an oscillation with a full cycle, involving both atmospheric and oceanic feedback, is completed.

The unexpected and prolonged persistence of warm conditions over the tropical Pacific during the early 1990s can be explained well by the G-P paradigm. The oceanic connection between the midlatitudes and tropics, which can be well understood by classical ventilation thermocline theory (Luyten et al., 1983), is also clearly identified from other observations (e.g., Zhang et al., 1998). The Pacific upper-ocean warming and decadal changes in the ENSO after 1976 may originate from decadal midlatitude variability because this oceanic bridge can make midlatitude surface warm water, formed in the early 1970s, subduct and penetrate through the subtropics and into the tropics, thus perturbing the tropical thermocline and driving the formation of warm surface water affecting ENSO in the 1980s (Zhang et al., 1998).

In the G-P paradigm, the atmosphere variation in the tropics is mostly a passive response to the ocean, while the atmosphere in the midlatitude acts as an active driven forcing to the ocean. Since the tropical atmospheric decadal variability is maintained by extratropical processes, the G-P paradigm implies there are no potential mechanisms or processes that could maintain a self-sustained tropical decadal oscillation. The ocean component in this paradigm is critical because it not only connects variability in the high latitude atmosphere-ocean to that in tropics, but it also sets up the oscillation period that depends on the traveling time of water parcels from the extratropics to the equator by subduction flow. The latter is quite different from the L-B paradigm, in which the timescale is determined by the surface oceanic process and the first baroclinic planetary wave. Here the subduction flow in the G-P paradigm can be understood as the second baroclinic Rossby wave within thermocline (Stephens et al., 2001). Although the first baroclinic wave travels much faster than the second baroclinic wave, the time it takes for the former from east to west and for the latter from midlatitudes to the tropics are comparable (around 10 years) because the width of the Pacific is much larger than the subduction pathway. Another difference between the L-B paradigm and the G-P paradigm is, of course, that the former focuses only on the North Pacific while the latter is established for the whole Pacific.

One debate on the G-P paradigm is whether the midlatitude thermocline variability can penetrate all the way into the tropics. Some data analyses show that the anomalous SST in the equatorial central Pacific is correlated with tropical Ekman pumping while the correlation with thermal coupling in the North Pacific eight years earlier is not significant (Schneider et al., 1999a, b). Using ocean general circulation model hindcasts and a coupled ocean atmosphere model, Schneider et al. (1999a, b) further conclude that the anomalous subduction signal originating from the midlatitude outcropping region can only propagate southwestward to 18°N (about 8 years later). These results imply no significant coupling in the North Pacific and the equatorial region via advection of thermal anomalies along the oceanic thermocline. Therefore, equatorward of 18°N it is tropical Ekman pumping, rather than further propagation of the midlatitude signal, that dominates the generation of thermal anomalies on decadal timescales in the equatorial Pacific Ocean. However, they also do not exclude that the midlatitude signal may significantly affect the tropical lower thermocline and deeper water. This debate is still open to study.

Another debate on the G-P paradigm is whether the subduction is the only process to carry midlatitude anomalies to the equatorial region. By the advection process, the warm anomalies take approximately 10 years to reach the equator, whereas the time lag between the central North Pacific and the tropical warm transition is only about 3–5 years (Lysne et al., 1997). Therefore, a wave mechanism may exist as a faster carrier of the anomalous signal (Lysne et al., 1997). This mechanism is depicted as follows: the warm anomaly propagates from the central North Pacific to the western boundary as long Rossby waves, southward along the coast as coastal Kelvin waves, and eastward along the equator as equatorial Kelvin waves, it then moves up to surface and eventually affects ENSO (Lysne et al., 1997). This wave mechanism, though giving a shorter response time of tropics-to-midlatitude signals, is only applicable to active tracers (e.g., temperature).

For passive tracers (e.g. salinity which is more like a passive tracer than an active tracer), passive tracers affect the tropics by water exchange and not by wave propagation. There is also a *western boundary pathway* for passive tracers: water subducts in the subtropics and is transported southwestward to the western boundary by the subtropical gyre, then to the equator by subsurface low latitude western boundary currents, then eastward by the equatorial undercurrent with water rising to the surface by a recirculating tropical cell (Liu et al., 1994; McCreary and Lu, 1994). The transit time for passive tracers along the western boundary pathway is about 10 years or longer.

The least disputed issue on PDO is that researchers tend to agree the timescale of PDO is determined by ocean, no matter how the ocean adjust to the changing climate system, by means of the first baroclinic planetary wave, the second baroclinic Rossby wave (advection process), or both the Rossby wave and the Kelvin wave. This view is held because only ocean dynamics can provide the mechanism to keep the climate variability alive for decades or longer.

3.1.3 *Tropical Pacific Decadal Oscillation*

Although many researchers accept that PDO originates primarily within the midlatitudes and that the tropical decadal variability is a passive response to PDO, there are also many studies showing that the PDO is intrinsically tropically driven (e.g., Knutson and Manabe, 1998; Schneider et al., 2000) and may teleconnect with the extratropics throughthe atmosphere (Trenberth, 1990; Graham, 1994). Very recently, researchers have come to realize that both the tropical Pacific and North Pacific can have their own decadal-scale oscillation origin due to local coupling processes (Liu et al., 2002; Wu et al., 2002). The most interesting idea is that a tropical decadal mode exists in spite of decadal variability in higher latitudes. The Tropical Pacific Oscillation (TPO) mode has a timescale of 20–30 years and can be enhanced significantly by extratropical-tropical teleconnection. Furthermore, the TPO may involve the higher baroclinic mode in the vertical in the tropics in order to maintain decadal timescale oscillation (Wu et al., 2002). Although this is a very fascinating paradigm, there is few observational evidence to support TPO, and it would also be difficult to isolate TPO from PDO even if these observations were available, since the observed climate variability is a result of complex climate feedbacks (Liu et al., 2002). Thus it is necessary to establish a simple conceptual model to explore the dynamics of the alleged TPO.

3.2 Stochastic origins

Stochastic forcing is thought to be one of the origins of the PDV as mentioned before, and, it might have a substantial contribution to the North Pacific variability as well as the tropical decadal variability (Latif and Barnett, 1994; Barnett et al., 1999a; Pierce and Barnett, 2000). The basic concept for stochastic origin of the PDV is that, if there is a stochastic component (spatially coherent white noise, for example, atmospheric transient eddies) in an air-sea coupling system, does the ocean response redden in its frequency spectrum? It appears that the stochastic forcing does exist for the ocean because the atmospheric activities can be thought as having no preferred timescale and are associated with an essentially white noise spectrum (e.g., Feldstein and Robinson, 1994). In this concept, the decadal variability is the low-frequency part of the red oceanic spectrum. Therefore, the question is, what determines the timescale and spatial structure of a stochastically forced decadal variability?

There have been many studies revealing that the stochastic forcing may excite oceanic modes of variability on decadal and interdecadal timescales. Most of the studies couple a linear ocean model to wind forcing that includes a stochastic component (e.g., Frankignoul et al., 1997). The linear ocean model might be a reduced-gravity model plus a time-dependent Sverdrup balance (Frankignoul et al., 1997; Jin, 1997) or a potential vorticity equation (PVE) (Munnich et al., 1998). Some people use a nonlinear reduced-gravity equation and conclude that the nonlinearity appears to be important to decadal variability (Sura et al., 2000). The wind forcing usually consists of a time-dependent amplitude with a spatially fixed pattern (Munnich et al., 1998) or sum of a mean wind field, a stochastic wind field which represents atmospheric transient eddies, and a large-scale ocean-atmosphere coupling (Sura et al., 2000). The coupling is established by relating wind stress to thermocline depth anomalies (Munnich et al., 1998; Sura et al., 2000), or to anomalies in the meridional temperature gradient of the upper ocean (Jin, 1997).

The ocean response to the overlying forcing is usually completed in terms of baroclinic planetary waves as discussed before. The timescale of the stochastically forced variability is thus determined by the time for the baroclinic planetary wave to cross the entire basin (Frankignoul et al., 1997). Munnich et al. (1998) obtained an oceanic spectrum with a red peak at a decadal timescale of 17 years based on a linear PVE for baroclinic planetary waves. Sura et al. (2000) identified a 5-yr variability in oceanic circulation using a reduced-gravity shallow-water model. Saravanan and McWilliams (1997) exhibited an interdecadal variability with a timescale of 30–40 years using an idealized ocean-atmosphere model involving thermohaline circulation. The spatial patterns of the decadal scale oceanic modes, therefore, most closely match the preferred atmospheric spatial patterns as a result of "spatial resonance" of the forced baroclinic planetary wave by the ocean-atmosphere coupling (Sura et al., 2000). The spatial inhomogeneity of the stochastic forcing is emphasized as being crucial for driving oceanic eddies and for making low frequency variability possible (Sura et al., 2000).

In general in this paradigm the crucial role of the atmospheric stochastic forcing is emphasized, which is one of the fundamental components for decadal oceanic modes. The spatial structure is set by the so-called spatial resonance, that is, the ocean pattern must most closely match the preferred atmospheric spatial pattern. Though it is highly idealized and the emergence of a decadal spectral peak in the oceanic spectrum strongly depends on the selected model parameters (Jin, 1997), the coupled model in this paradigm provides insights into the main physical mechanisms of the North Pacific decadal-interdecadal variability observed in nature and simulated in coupled circulation models. However, there are two big problems in the coupled ocean-atmosphere model with stochastic wind forcing. First, the stochastically forced PDV, or more precisely, the North Pacific decadal variability, cannot impose itself on the tropical Pacific in the frame of this simple model, while the stochastic component of the PDV is found to exert a remarkable influence on the ENSO via the atmospheric bridge in complex coupled GCMs (Barnett et al., 1999a; Pierce and Barnett, 2000). Second, it is highly disputed and we are far from reaching an agreement on how the oceanic and atmospheric systems couple to each other in the *midlatitude* climate system. Researchers tend to accept that Bjerknes positive feedback (Bjerknes, 1969; Cane and Zebiak, 1985) plays a critical role in a coupled equatorial air-sea system because of negligible earth rotation: anomalous warming (cooling) on equatorial SST would cause anomalous easterly (westerly) flow in the east and anomalous westerly (easterly) flow in the west, which in turn reinforce the initial warming anomaly. The corresponding changes in the ocean thermocline show an anomalous Ekman upwelling (downwelling) to the east and downwelling (upwelling) to the west, thus a shallower (deeper) thermocline depth to the east and deeper (shallower) thermocline to the west. However, it is not obvious wehther these processes also work for the midlatitude

4. Discussions

It seems undoubtedly both the oceanic and atmospheric components are indispensable in the Pacific decadal and interdecadal oscillations. The PDV is so complex that it involves almost every aspect of climate systems. The state of the art on PDV study is such that the deeper people explore, the more arguments come out. For example, it is realized that the PDV has multiple origins: a deterministic origin and a stochastic one. These two origins usually cannot separate from each other. Even for stochasticly-originated PDV, the deterministic (coupling) processes always play critical roles by enhancing initial variabilities and providing feedback mechanisms. In the frame of deterministic origin, it is still hard to reach agreement over what the primary process controlling PDV is, extratropical coupling or tropical coupling and over which climate component plays a dominant role, atmosphere or ocean. In addition, the PDV involves a complicated interaction between different timescale climate systems (e.g., ENSO), or even weather systems (e.g., atmospheric eddies). Recent studies show that the 1976/77 Pacific climate shift might have affected ENSO in the late 1970s by changing the background tropical winds and associated equatorial upwelling (Wang and An, 2002) and these studies emphasize the critical role of the atmospheric bridge that rapidly conveys the influences of extratropical decadal variations to the tropics. The studies are quite different from Gu and Philander (1997) and Zhang et al. (1998) which emphasize the role of the oceanic bridge.

So far, it is not easy to clearly understand the dynamics of the PDO in a coupled climate system. Since most of the decadal scale memories are stored in the ocean, people are switching back to simple conceptual models from coupled GCMs, to exclusively explore the eigenmodes in a closed oceanic basin because the dynamics of oceanic eigenmodes are easy to understand. Very recently, Cessi and Paparella (2001) found that low frequency (up to decadal time scale), large-scale oceanic modes exist by applying the reduced-gravity shallow water equations to a closed basin. This kind of basin modes is slightly damped because of natural friction. However, they can be easily excited and become sustained by air-sea coupling or stochastic wind forcing (Cessi and Louazel, 2001). The constraint of mass conservation is essential for the existence of low frequency, large-scale basin modes (Cessi and Primeau, 2001). The lowest frequency, that is, the longest timescale, is determined by the transit time of planetary waves along the north boundary of the oceanic basin. These studies suggest that the dynamic memory of the ocean could last for decades for the Pacific, and explicitly demonstrate that mid and high latitude ocean is the only indispensable component of the PDV. Some studies show that the decadal very low frequency modes may also exist in the tropical basin (Jin, 2001), however, they appear to be determined by extratropical planetary waves (Liu, 2002) and the tropical ocean would not have its own decadal memory without the contribution from the extratropical ocean in a linear reduced-gravity model.

Apparently, the studies on PDV will need to last for decades for finally understanding its mechanism and further predicting its occurrence due to its complexity. Theoretical conceptual model studies are exceedingly needed to disclose the primary dynamics of PDV. The coupled climate system models are also indispensable not only to reproduce the observed decadal and interdecadal variability, but more importantly, to probe the mechanisms of PDV by designing appropriate sensitivity experiments. Of course, all of the studies must be based on observations which are inadequate at present, and hopefully, more observations will become available in the near future.

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太平洋年代际振荡研究进展

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摘　要

近10年来，太平洋年代际振荡(PDO)因其对全球气候系统的深远影响而得到广泛的研究。PDO指的是在太平洋的气候变率中具有类似ENSO空间结构但周期为10–30年的一种振荡，当北太平洋中部海面温度异常增暖(冷却)时，热带太平洋中部和东部以及北美沿岸常同时伴随有同等幅度的异常冷却(增暖)。总体而言，有两类观点分别认为PDO起源于确定的海气耦合过程或起源于大气的随机强迫。确定性起源论强调，一个海气耦合系统内部的物理过程可以提供一个正反馈机制以增强一初始扰动，及一个负反馈机制以促使振荡位相发生逆转；海洋环流的动力演变过程决定了振荡的时间尺度。随机性起源论则强调，因为大气活动没有一个特定的时间尺度，其时间尺度谱实际上对应于白噪音谱，所以大气对海洋的强迫是随机的；而海洋常在低频谱段有最大的响应振幅，其对应的周期约为十几年或几十年。作者试图系统性地理解PDO在观测、理论和数值方面的研究现状，从而为当前研究提供一个有用的背景性参考。

关键词: 气候系统, 太平洋年代际振荡, 确定性起源, 随机性起源