Dynamics of Interdecadal Climate Variability: A Historical Perspective*

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ABSTRACT

The emerging interest in decadal climate prediction highlights the importance of understanding the mechanisms of decadal to interdecadal climate variability. The purpose of this paper is to provide a review of our understanding of interdecadal climate variability in the Pacific and Atlantic Oceans. In particular, the dynamics of interdecadal variability in both oceans will be discussed in a unified framework and in light of historical development. General mechanisms responsible for interdecadal variability, including the role of ocean dynamics, are reviewed first. A hierarchy of increasingly complex paradigms is used to explain variability. This hierarchy ranges from a simple red noise model to a complex stochastically driven coupled ocean–atmosphere mode. The review suggests that stochastic forcing is the major driving mechanism for almost all interdecadal variability, while ocean–atmosphere feedback plays a relatively minor role. Interdecadal variability can be generated independently in the tropics or extratropics, and in the Pacific or Atlantic. In the Pacific, decadal–interdecadal variability is associated with changes in the wind-driven upper-ocean circulation. In the North Atlantic, some of the multidecadal variability is associated with changes in the Atlantic meridional overturning circulation (AMOC). In both the Pacific and Atlantic, the time scale of interdecadal variability seems to be determined mainly by Rossby wave propagation in the extratropics; in the Atlantic, the time scale could also be determined by the advection of the returning branch of AMOC in the Atlantic. One significant advancement of the last two decades is the recognition of the stochastic forcing as the dominant generation mechanism for almost all interdecadal variability. Finally, outstanding issues regarding the cause of interdecadal climate variability are discussed. The mechanism that determines the time scale of each interdecadal mode remains one of the key issues not understood. It is suggested that much further understanding can be gained in the future by performing specifically designed sensitivity experiments in coupled ocean–atmosphere general circulation models, by further analysis of observations and cross-model comparisons, and by combining mechanistic studies with decadal prediction studies.

1. Introduction

Owing to societal need, the prediction of climate in the next few decades is emerging as one of the top priorities in climate research (Smith et al. 2007; Keenlyside et al. 2008; Latif et al. 2009; Meehl et al. 2009a; Hurrell et al. 2009). However, our ability to predict climate change over the next few decades is still in its infancy. Unlike centennial climate change, which is predominantly driven by anthropogenic climate forcing, interdecadal climate change is driven by natural interdecadal variability as well as anthropogenic forcing, especially at the regional (continental/basin) scale. (For consistency, in this paper, ''interdecadal'' variability refers broadly to climate variability on time scales of 10–100 yr, while ''decadal'' and ''multidecadal'' variability refer to the variability from $\sim$10 to 20 yr and from 20 yr to several decades, respectively.) A recent study further suggests that for climate changes over the next 10–30 yr, uncertainty in natural climate variability is greater than that in anthropogenic forcing (Hawkins and Sutton 2009). This implies that natural interdecadal variability is of critical importance to climate prediction of the next few decades.

Even after two decades of intensive study, our understanding of natural interdecadal variability remains poor.

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TABLE 1. Status of our understanding of interdecadal variability.

Variability	What we know	What we do not know
Overall	Stochastic noise important for driving, ocean dynamics important for time scale, tropical atmosphere important for global response, extratropical atmospheric response modest	Preferred time scale in the real world Role of extratropical ocean–atmosphere feedback Role of tropics vs extratropics
Pacific decadal	May originate from both extratropics and tropics; subtropical–midlatitude Rossby wave important for time scale	Role of tropical ocean dynamics
Pacific multidecadal	Originates from the extratropics; subpolar Rossby wave important for time scale	Role of salinity and temperature variability
Atlantic decadal	Tropical WES feedback important, but North Atlantic variability may be an important driving mechanism	What determines the time scale
Atlantic multidecadal	Thermohaline circulation important for time scale	Thermohaline instability vs baroclinic instability Role of subpolar gyre

Numerous mechanisms have been proposed without firm conclusions. As such, it is time for an overview of (i) our current understanding of the mechanisms responsible for interdecadal variability from a unified framework and (ii) the historical development of our understanding.

Intensive study of the mechanisms responsible for interdecadal climate variability started in the late 1980s and early 1990s, when oceanic general circulation models (OGCMs) and fully coupled ocean–atmosphere general circulation models (CGCMs) became widely available. Weaver and Sarachik (1991a,b) first studied interdecadal variability in the context of the Atlantic thermohaline circulation using an OGCM with highly simplified surface boundary conditions. Latif and Barnett (1994) proposed the first mechanism of North Pacific interdecadal variability in terms of the upper-ocean wind-driven circulation and its interaction with the atmosphere in a CGCM. Subsequently, numerous mechanisms have been proposed for interdecadal variability, ranging from self-exciting coupled modes to stochastically driven oceanic modes, from tropical origin to extratropical origin, and from variability linked to the wind-driven circulation to the thermohaline circulation.

The studies conducted in the last two decades have increased our understanding of interdecadal variability significantly in the Pacific and Atlantic, as summarized in Table 1. One significant advance is the recognition of the role of stochastic climate forcing (Hasselmann 1976), instead of the self-exciting generation mechanism, as the major driving force for almost all interdecadal variability. While the Hasselmann mechanism provides a powerful paradigm for partially understanding interdecadal variability in many locations, many fundamental questions remain. For example, where a variability mode exhibits a preferred interdecadal time scale, what is the mechanism responsible for selecting the time scale? What is the role of ocean–atmosphere feedback in interdecadal variability? In general, the validation of the detailed mechanisms has remained challenging. This difficulty is due in part to the relatively short and sparse observations available. This makes it difficult to identify robust signals of interdecadal variability and to determine if an observed interdecadal variability mode actually has a preferred time scale with a highly significant spectral peak (Wunsch 1999). Long-term proxy records, such as tree ring and coral records, provide an important source of observation for interdecadal variability complementary to instrumental records (e.g., Linsley et al. 2000; Cole et al. 2000; Evans et al. 2001; Mann et al. 1995; Peterson et al. 2000). However, proxy records are limited by their temporal and spatial distributions, as well as by the inaccuracy with which they actually represent genuine climate variability.

Here, we will review research conducted on the cases of interdecadal variability in the coupled ocean–atmosphere systems in the Pacific and Atlantic, where interdecadal variability has been studied most intensively. We will focus on the following issues:

- What are the roles of stochastic forcing and deterministic dynamics?
- What are the roles of the tropics and extratropics in driving interdecadal variability?
- What are the roles of thermohaline and wind-driven circulations?
- What is the role of ocean–atmosphere coupling?

This review builds on the earlier reviews provided by Latif (1998), Miller and Schneider (2000), Mantua and Hare (2002), Minobe (2000), and Power and Colman (2006) for the Pacific, and by Xie and Carton (2004), Latif et al. (2006), and Delworth et al. (2007) for the Atlantic. This review complements discussions on natural interdecadal variability in the Indian Ocean (e.g., Schott et al. 2009), in the Arctic region (e.g., Proshutinsky et al. 2002), over the continents (e.g., Cayan et al. 1998; Ding et al. 2009), and forced by external forcing such as solar variability and volcanic forcing (e.g., Crowley 2000; White and Liu 2008; Meehl et al. 2009b). It also complements discussions of decadal climate predictability (e.g., Power et al. 2006; Latif et al. 2007, 2009). In particular, the mechanisms responsible for various interdecadal variabilities in both oceans and from decadal to multidecadal time scales are discussed from a unified framework in light of the historical development. It is hoped that the unified framework and historical perspective will help us gain deeper insight into the mechanisms responsible for interdecadal variability. One caution for the readers is that, in spite of my best effort, the review inevitably still reflects my personal bias and limited knowledge of the subject.

The paper is organized as follows. In section 2, we briefly describe observed interdecadal variability in the Pacific and Atlantic Oceans. In section 3, we will discuss some general issues important to interdecadal variability in a unified framework and in a hierarchy of paradigms of increasing complexity. The unified perspective emphasizes the intrinsic and common dynamics applicable to different oceans. We will then review the mechanisms responsible for specific interdecadal variability. We will first review Pacific variability, in section 4, with the focus on the relative roles of the tropics and extratropics. We will then review the mechanisms responsible for Atlantic variability, in section 5, with the focus on the mechanism linked to thermohaline circulations. In section 6, we will briefly review ocean–atmosphere feedback in the extratropics, which is closely relevant to the study of interdecadal variability. Finally, a summary and discussion will be given in section 7.

2. Observations

Significant interdecadal climate variability has been observed over the world. Much attention so far has been paid to the Pacific (e.g., Trenberth and Hurrell 1994; Zhang et al. 1997; Minobe 1997; Folland et al. 2002; Power et al. 1999) and Atlantic (e.g., Bjerknes 1964; Deser and Blackmon 1993; Kushnir 1994) sectors. In the Pacific, there is significant bidecadal variability, as seen in the leading principal component (PC) of the (6 yr) low-pass sea surface temperature and the regression map of the Pacific SST (Figs. 1b,d, Zhang et al. 1997), sometimes called Pacific decadal variability (PDV).[1] The PDV can be identified as the leading EOF mode of interdecadal SST variability in the Pacific, which has been called the Pacific decadal oscillation (PDO) (Mantua et al. 1997), or in the global ocean, called interdecadal Pacific oscillation (IPO) (Power et al. 1999; Folland et al. 2002; Randall et al. 2007). In addition, there is multidecadal variability on a 50–70-yr time scale, as seen in the first PC of the annual mean sea level pressure (SLP) over the North Pacific (NPI index, Figs. 2a,b), which shows a coherent pattern throughout the Pacific, that we will refer to as a dominant pattern of Pacific multidecadal variability (PMV).

These two patterns, one a part of PDV and the other PMV, have a very similar structure and are both similar to the pattern associated with ENSO (Figs. 1a,c). These patterns are characterized by coherent variability over the tropical Pacific with coherent variability of the opposite sign in the midlatitude North Pacific (Figs. 1d and 2b). The PDV pattern also tends to vary symmetrically with respect to the equator, with significant variability in the South Pacific (Garreaud and Battisti 1999; Power et al. 1999). So much is the similarity to ENSO that the PDV and PMV are sometimes referred to as the ENSO-like Pacific interdecadal variability (Zhang et al. 1997; Power and Colman 2006). Nevertheless, it should be noted that Pacific interdecadal variability has a large amplitude in the midlatitudes (Figs. 1d, 2b) while ENSO is dominated by maximum variability in the tropics (Fig. 1c). In addition, along the equator, the Pacific decadal variability is centered toward the central Pacific (Fig. 1d), while ENSO variability is centered toward the eastern Pacific (Fig. 1c). The region of tropical coherence also extends farther poleward for the PDV and PMV than it does for ENSO (Power and Colman 2006). The coexistence of decadal and multidecadal variability in the Pacific can be seen, for example, through wavelet analysis (Minobe 1999).

The spatial character of PDV and PMV is robust in various analyses. For example, the first and second EOFs of annual SST over the North Pacific (>20°N) have been called the Pacific decadal oscillation (Mantua et al. 1997) and the North Pacific gyre mode (NPGO) (Di Lorenzo et al. 2008). The first EOF resembles the PDV pattern and the second EOF resembles the PMV pattern described above (Di Lorenzo et al. 2010, manuscript submitted to *Geophys. Res. Lett.*). The PDO appears to be associated with the atmospheric variability of the Aleutian low while the NPGO is associated with the North Pacific Oscillation (NPO) [as defined originally by Walker and Bliss (1932), with Oscillation referring to the oscillation in space, not in time]. The PDO can be regarded as the North Pacific expression of the basinwide IPO (e.g., Folland et al. 2002; Randall et al.

[1] Here, a type of coherent variability will be referred to simply as variability, instead of an oscillation or a mode as in many other works. This may avoid the impression of periodic evolution implied by the term oscillation; it may also avoid the impression of being a physical mode implied by the term mode, rather than a mode derived using a statistical method (such as an EOF mode) as in the observation.

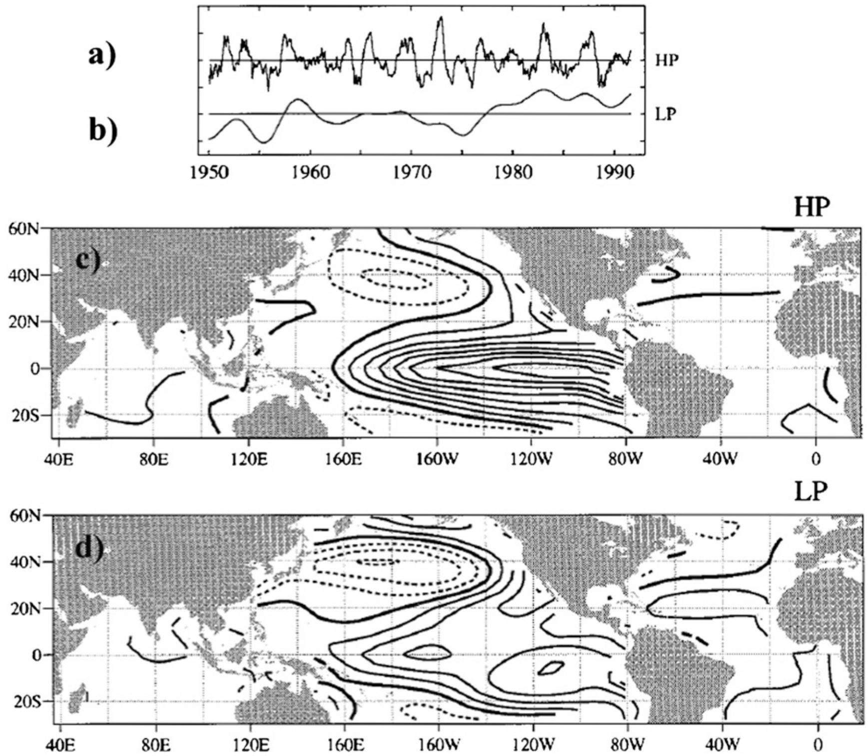

FIG. 1. The leading (normalized) principal components (PCs) of 6-yr (a) high-pass and (b) low-pass filtered SST over the Pacific domain shown together with (b),(d) the associated regression patterns for global SST. The interval between tick marks on the vertical axis of the top panel corresponds to one standard deviation; the spacing between the curves is arbitrary. Contour interval is 0.1 K per standard deviation of the expansion coefficient time series. Negative contours are dashed; the zero contour is thickened. (Adapted from Zhang et al. 1997.)

2007). The second and third rotated EOFs of annual SST over the Pacific resemble the PDV and PMV, respectively (Barlow et al. 2001). The first and third rotated EOFs of the low-pass annual SST over the Pacific resemble the PDV and PMV patterns, respectively (Wu et al. 2003). In addition to the basinwide variability, there is evidence of quasi-decadal (~10 yr) variability localized in the Pacific (e.g., Brassington 1997; Tourre et al. 2001; Luo and Yamagata 2001; Qiu 2003). The mechanism for the local decadal variability is not the focus here.

The Atlantic interdecadal climate variability is dominated by multidecadal variability (50–70 yr) and quasi-decadal (~10 yr) variability (Deser and Blackmon 1993; Latif et al. 2007; Alvarez-Garcia et al. 2008), which we will refer to as the Atlantic multidecadal variability (AMV) and Atlantic decadal variability (ADV). The AMV can be seen in the first EOF of annual SST over the North Atlantic, characterized by anomalies of one sign across the North Atlantic with maximum amplitude in the subpolar North Atlantic (Figs. 3a,b) (Delworth et al. 2007). This multidecadal variability corresponds to the Atlantic multidecadal oscillation (AMO) identified in previous work and has a significant impact on climate variability in and around the Atlantic (e.g., Folland et al. 1986; Kushnir 1994). The ADV can be seen in the second EOF of annual SST in the North Atlantic, which exhibits a tripole anomaly over the North Atlantic (e.g., Delworth et al. 2007). This tripole SST variability appears to be associated with atmospheric variability linked to the North Atlantic Oscillation (NAO) (Hurrell 1995). There is also evidence of decadal variability in a Pan-Atlantic pattern (Fig. 4). This pattern seems to be significantly correlated with sea ice variability in the subpolar North Atlantic (Deser and Blackmon 1993) and with tropical Atlantic variability (TAV), especially the interhemispheric SST gradient (Fig. 4; Tanimoto and Xie 1999, 2002).

Summary

The coexistence of decadal and multidecadal variability appears to be a robust feature in both the Pacific and Atlantic. The difference in time scales between the decadal and multidecadal variability raises the

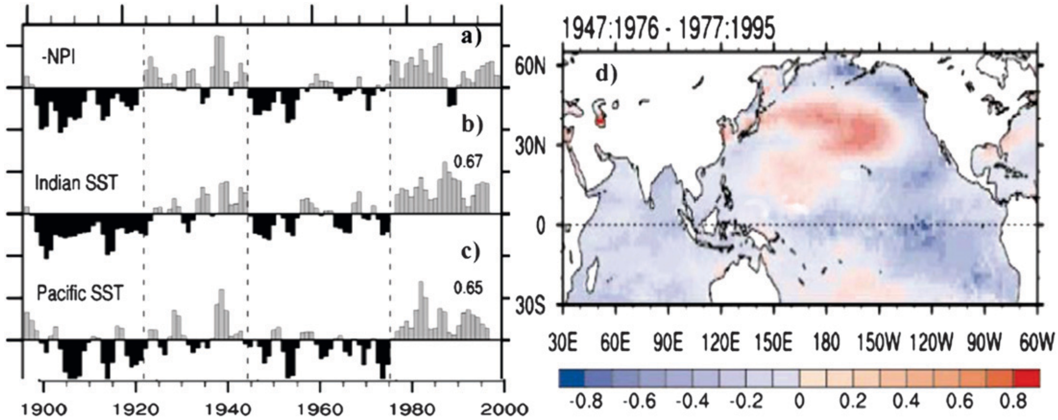

FIG. 2. Selected climate anomaly records for boreal winter during 1900–97: (a) the inverted NPI (area-averaged sea level pressure for the region 30°–65°N, 160°–140°W), (b) tropical Indian Ocean SST, and (c) southeastern tropical Pacific SST. All records are normalized and smoothed with a three-point binomial filter. Each tick mark on the ordinate represents one standard deviation. The numeral to the right of each tropical index represents its correlation coefficient with the inverted NPI. (d) Epoch difference map of boreal winter SST (from 1977–95 to 1947–76) from the Kaplan et al. (1998) dataset. (Adapted from Deser et al. 2004.)

possibility of multiple mechanisms for interdecadal variability.

3. General mechanisms: Stochastic versus deterministic dynamics

Two key questions arise for all interdecadal climate variability: is the variability self-excited or stochastically driven and what determines the time scale of the variability? With recent studies in more realistic models, especially in CGCMs, it appears that almost all interdecadal variability is driven by stochastic climate noise that is associated with internal atmospheric variability (rather than self-exciting). As such, a red noise model, which has no preferred oscillation time scale, is often used as a null hypothesis to test against whether an interdecadal variability mode is oscillatory. If an interdecadal mode is found to have spectral peaks significantly different from a red noise, this mode can be assumed to be of certain oscillatory nature. In this case, the default paradigm will be a stochastic climate model of a damped oscillator and the critical issue is then to understand the mechanisms that determine the oscillation time scale of the interdecadal mode. Here, we provide a brief overview of various mechanisms. More details on these mechanisms will be provided later.

a. Generation mechanism

The stochastic climate theory, although proposed long ago for general climate variability (Hasselmann 1976), was not appreciated in the study of interdecadal (and interannual) climate variability until the mid-1990s. While some studies examined variability in ocean models forced with stochastic forcing (e.g., Power et al. 1995), most early work on interdecadal variability tended to focus on deterministic dynamics and associated self-excited oscillation mechanisms. These latter studies attribute the origin of interdecadal variability to instabilities either in the ocean (e.g., Weaver and Sarachik 1991a,b) or in the coupled ocean–atmosphere system (e.g., Latif and Barnett 1994; Gu and Philander 1997). This deterministic thinking is based on several historical reasons: the apparent success of studies in explaining ENSO as a self-exciting mode at that time (Philander et al. 1984; Cane and Zebiak 1985; Suarez and Schopf 1988); a lack of understanding of ocean–atmosphere interaction outside the tropics (Frankignoul 1985; Bretherton and Battisti 2000; Kushnir et al. 2002); inadequate analyses of the variability from a coupled perspective in CGCMs and observations; and limitations in most models employed for the study of interdecadal variability at that time. Most of the models used at the time were simplified models with highly simplified atmospheres, which underrepresented or neglected entirely stochastic atmospheric variability. Nevertheless, some studies in ocean GCMs suggested the potential importance of stochastic forcing on interdecadal variability (e.g., Mikolajewicz and Maier-Reimer 1990). In their hybrid coupled model, Power et al. (1995) found that decadal variability in SST was primarily driven by stochastic heat flux forcing, with stochastic wind stress driving variability approximately half as large and with stochastic freshwater forcing relatively unimportant except at polar latitudes.

More recent studies with advanced statistical analyses and sensitivity experiments using CGCMs tend to point to stochastic forcing as the major driving mechanism for

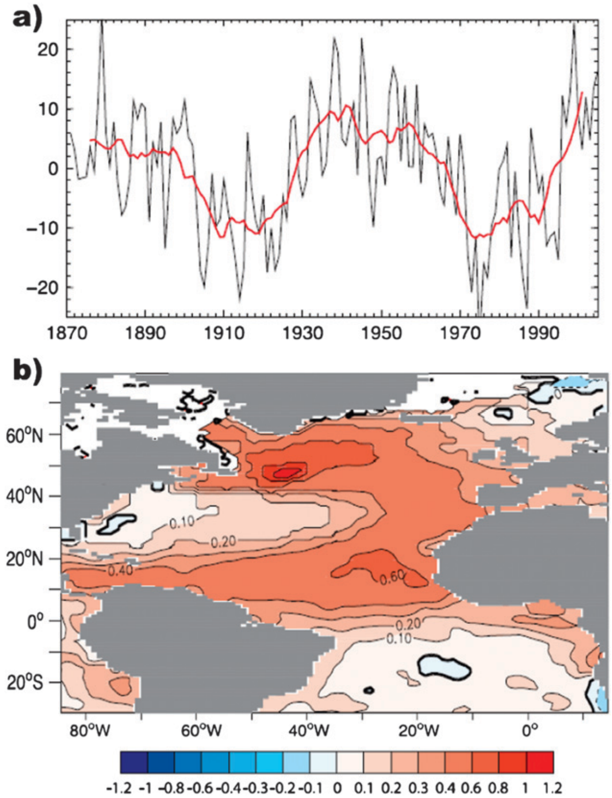

FIG. 3. (a) The first PC (dimensionless) and (b) the regression map of the observed annual mean SST (K per two standard deviations of the PC). Contour interval is 0.1 between −0.4 and 0.4 and 0.2 otherwise. The EOF analysis is conducted over the period 1870–2005 over the domain from 0° to 60°N in the Atlantic using the SST data of Rayner et al. (2003). (Adapted from Delworth et al. 2007.)

interdecadal variability. For example, in a series of systematic studies on Atlantic multidecadal variability in the Geophysical Fluid Dynamics Laboratory (GFDL) CGCM, Delworth and his colleagues (Delworth et al. 1993; Delworth and Greatbatch 2000) found that the multidecadal variability evident in their model could be generated largely in the ocean model by stochastic heat flux forcing, although its variance can be enhanced modestly by ocean–atmosphere coupling. Similar conclusions are derived from CGCM experiments using the so-called ensemble coupling technique (Yeh and Kirtman 2004, 2006) in which interdecadal variability is found to decrease as atmospheric internal variability is decreased. In ensemble coupling, an ensemble of atmospheric general circulation models (AGCMs) is coupled with a single ocean model with the ensemble mean of the AGCM output to force the ocean, such that stochastic atmospheric variability is suppressed (Kirtman and Shukla 2002). There are also CGCM studies that highlight the roles of positive ocean–atmosphere feedback and the self-exciting mechanism (e.g., Latif and Barnett 1994; Timmermann et al. 1998). However, these studies are based on the analysis of a single control simulation. In general, it is difficult to identify the role of ocean–atmosphere coupling unambiguously in a single control simulation, a point to be pursued later. It should be pointed out that, even if the interdecadal variability is driven by stochastic forcing, it is still important to understand the mechanism of positive ocean–atmosphere feedback, because these positive feedbacks can enhance the variance of interdecadal variability under stochastic forcing.

b. *Preferred time scales*

Regardless of the generation mechanism, it is important to understand the mechanisms responsible for the preferred time scales of interdecadal variability, where preferred time scales actually exist. Given our paradigm of a damped oscillator, two time scales are relevant: the damping time scale (persistence) and the oscillation time scale (period).

The damping time scale determines how long an initial anomaly can stay above the noise level and is therefore the key time scale for predictability, at least in a linear framework. The damping time scale also puts an upper bound on a meaningful oscillation time scale because interdecadal variability can exhibit oscillatory behavior only if its oscillation time scale is shorter than the damping time scale (Fig. 5). Due to the relatively fast atmospheric processes, it is natural to attribute the memory of interdecadal variability to the subsurface ocean. As such, the damping time may be estimated crudely as a thermal damping time of the upper-ocean heat content. If we assume temporal anomalies in the basin scale, the upper-ocean heat budget anomaly roughly satisfies the linearized local heat balance equation:

$$\rho C_p H \partial_t T' \sim \partial_T Q T', \qquad (1)$$

where ρ, C_p, and H are the density, heat capacity, and depth of the perturbed water, respectively; Q is the total surface heat flux, and $\partial_T Q$ is the heat flux sensitivity to SST perturbation. The variable T' is the temperature anomaly, which, for simplicity, has been assumed comparable with the SST anomaly as in a bulk mixed layer. The thermal damping time scale can be estimated as $\tau \sim \rho C_p H / |\partial_T Q|$, so the damping time scale increases with water depth. Given the observed total heat flux sensitivity at the basin scale as $\partial_T Q \sim -10$ W m^{-2} K^{-1} (Frankignoul et al. 1998), the thermal damping time scale can be estimated as ~1 yr for the top 50 m, and 10 yr for the top 500 m of the ocean. Therefore, one may expect the upper ocean in the main thermocline to be involved in interdecadal variability. Indeed, the deep

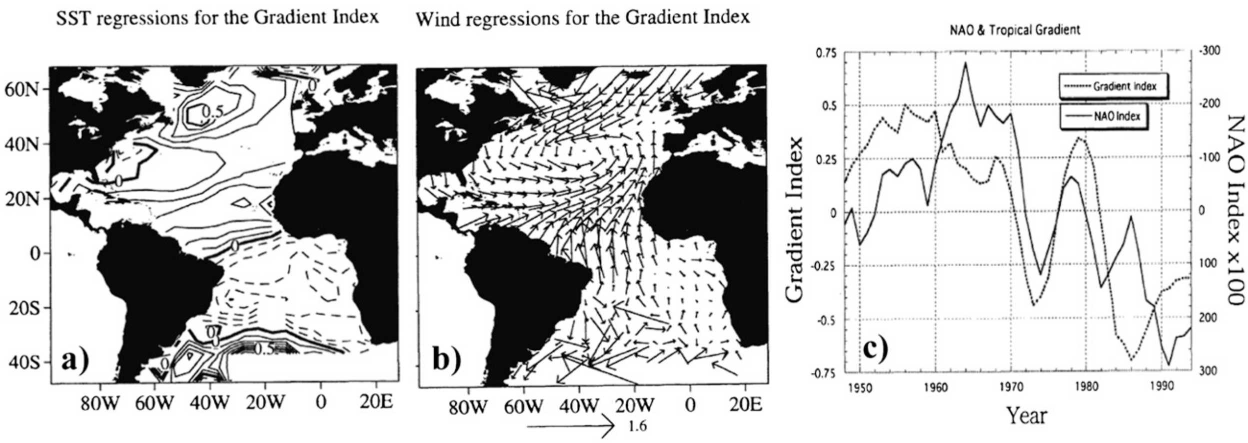

FIG. 4. The simultaneous regression patterns of (a) SST (°C) and (b) surface wind vector fields (m s^{-1}) based on the cross-equatorial SST gradient index. The index is defined as the SST difference between the northern (10°–20°N) and southern (10°–20°S) tropical Atlantic. Contour interval is 0.1°C for SST. Positive (negative) values are represented by the solid (dashed) lines. The reference wind speed vector (1.6 m s^{-1}) is indicated below (b). (c) The 5-yr running mean cross-equatorial SST gradient (dashed line) and NAO (solid line) indices. (Adapted from Tanimoto and Xie 1999.)

ocean provides a key memory for interdecadal variability and contributes greatly to the spectrum of low frequency variability (Fraedrich et al. 2004).

Most studies on the preferred time scale of interdecadal variability have focused on the oscillation time, which is usually identified as a distinctive spectral peak (Fig. 5). While the damping time scale does not necessarily need to involve ocean dynamics as presented in (1), the oscillatory time scale usually does involve ocean dynamics. From the predictability perspective, it is important to understand the oscillation mechanism because a distinctive oscillation period points to the potential for enhanced predictability (Griffies and Bryan 1997; DeSole and Tippet 2009).

The oscillation time scale is determined by the dynamics of the system and the spatial structure of the forcing. For convenience, we will categorize various stochastic climate models as a hierarchy determined by the degree of complexity of the model (Table 2). At the lowest level (tier 1) is the original stochastic climate model of Hasselmann (1976), which can be thought of as involving local ocean–atmosphere interaction only and will therefore be called the local stochastic climate model. This model of climate variability exhibits red noise with no preferred oscillation time scale (this model can also be thought of as a damped oscillator model with the frequency of zero). The low frequency variability is generated simply by the persistence of the anomaly competing against accumulated random variability. It should be pointed out that, in this model, seemingly periodic evolution can occur during a finite time window owing to a few accidental "swings" of a certain "period" instead of a systematic oscillating mechanism with a preferred period in a long time. At other times, the seemingly oscillatory behavior will disappear or appear with a different apparent oscillatory period, again arising from sampling error.

At the next level (tier 2a), wave propagation or current advection (with a speed C) leads to a model that we will refer to as the propagative (or advective) stochastic climate model. In this class, climate variability can vary

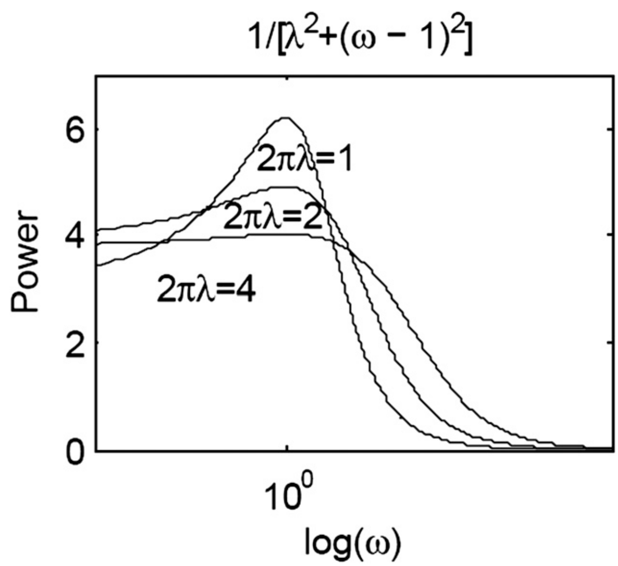

FIG. 5. Three theoretical (dimensionless) power spectra of damped oscillation with different damping rate λ: $dT/dt + iT = -\lambda T + w(t)$, where the natural frequency is taken as 1 and $w(t)$ is a white noise forcing. The spectral peak depends on the damping rate. The case of a strong damping rate $\lambda = 2/\pi$ shows little power spectrum peak; the modest damping rate $\lambda = 1/\pi$ shows a weak power spectral peak, while the weak damping case $\lambda = 1/2\pi$ shows a distinctive spectral peak.

TABLE 2. The hierarchy of stochastic climate models.

Tier	Symbolic equation	Model/mechanism	Preferred time scale
1	$\frac{dT}{dt} = -\lambda T + w(t)$	Stochastic climate model (Hasselmann 1976) local interaction	No (red noise)
2a	$\frac{\partial T}{\partial t} + C\frac{\partial T}{\partial x} = -\lambda T + w(t)$	Propagation stochastic climate model (Frankignoul et al. 1997; Jin 1997)	No
2b	$\frac{\partial T}{\partial t} + C\frac{\partial T}{\partial x} = -\lambda T + w(x,t)$	Propagation stochastic climate model with spatially variable forcing (spatial resonance; Saravanan and McWilliams 1997)	Yes
3	$T = T(h)$ $\frac{\partial h}{\partial t} + N(h) = -\lambda h + w(t)$	Stochastically driven ocean dynamic model (ocean dynamics and mode resonance; see Table 3)	Yes
4	$\psi = \psi(T)$ $T = T(h) + a\psi$ $\frac{\partial h}{\partial t} + N(h) = -\lambda h + b\psi + w(t)$	Stochastically driven coupled climate model (ocean feedback and coupled dynamics, coupled mode resonance)	Yes

from region to region, even under spatially uniform stochastic forcing, because of the spatial accumulation of wave energy by wave propagation (Frankignoul et al. 1997; Jin 1997). In most studies so far, the propagation is provided by the first baroclinic planetary wave in the extratropics, which propagates westward across an ocean basin at interdecadal time scales (Chelton and Schlax 1996) due to the non-Doppler-shift effect[2] (Anderson and Gill 1975; Liu 1999a,b). For a given period of τ, the SST variance increases downstream with the propagation to a saturation level that no longer increases; this first location of saturation is located about one propagation distance C_τ away from the region of wave generation. For very low frequency forcing, the propagation distance C_τ becomes larger than the basin size. Then the propagative model becomes the familiar Sverdrup relation. In this quasi-steady Sverdrup relation, the amplitude of the ocean variability increases toward the western boundary. Figures 6b,d show some examples in the North Atlantic. It is seen that the magnitude of the variability increases westward from the eastern boundary and then maintains its magnitude (at the saturated level) some distance away from the eastern boundary. The distance from the eastern boundary at which the magnitude of the solution stabilizes increases with the forcing period. Even if the stochastic forcing does not vary with space, variability in different regions can exhibit different spectral peaks. Consequently, there is no single preferred oscillation time scale over the basin. However, if the stochastic forcing exhibits a significant spatial variation of subbasin scale, with at least one cycle within the basin (tier 2b), the stochastic forcing can generate variability of a preferred period through the mechanism of spatial resonance (Saravanan and McWilliams 1997). In the North Atlantic, this propagating mechanism could offer an explanation for the decadal SST propagation along the North Atlantic Current in the observations (Sutton and Allen 1997). Spatial resonance occurs because the local forcing can act in phase with the downstream propagating variability, reinforcing its magnitude. However, later studies did not find clear evidence of spatial resonance in CGCMs (e.g., Pierce 2001; Delworth and Greatbatch 2000). This arises, in part, because the dominant atmospheric variability modes are usually of basin scale in the zonal direction, making them ineffective in exciting spatial resonance.

At the next level with basinwide ocean dynamics included, stochastic forcing can generate climate variability with preferred frequency simply as the resonance of ocean dynamic modes (tier 3)—a point to be revisited later. Finally and most generally, stochastic forcing can excite coupled climate modes with preferred time scales (tier 4) (e.g., Kleeman and Power 1994).

It should be pointed out that, although we tend to think of interdecadal variability with the paradigm of a damped oscillator (including the limiting case with zero frequency), it is generally difficult to unambiguously distinguish observed variability linked to a damped oscillator (tier 2 or 3) from variability associated with red noise (tier 1). This occurs because our time series are usually too short to contain sufficient interdecadal episodes, especially for multidecadal variability (Figs. 2a, 3a). Extreme caution is therefore needed in attributing the cause of interdecadal variability (Wunsch 1999).

[2] Non-Dopper-shift effect refers to the scenario inwhich the wave propagation speed is not affected by the mean flow. In the case of Rossby waves here, the non-Doppler-shift effect is caused by the cancelation between the advection by the mean flow and the opposite wave propagation induced by the mean potential vorticity gradient (or the so-called beta effect) associated with the mean flow.

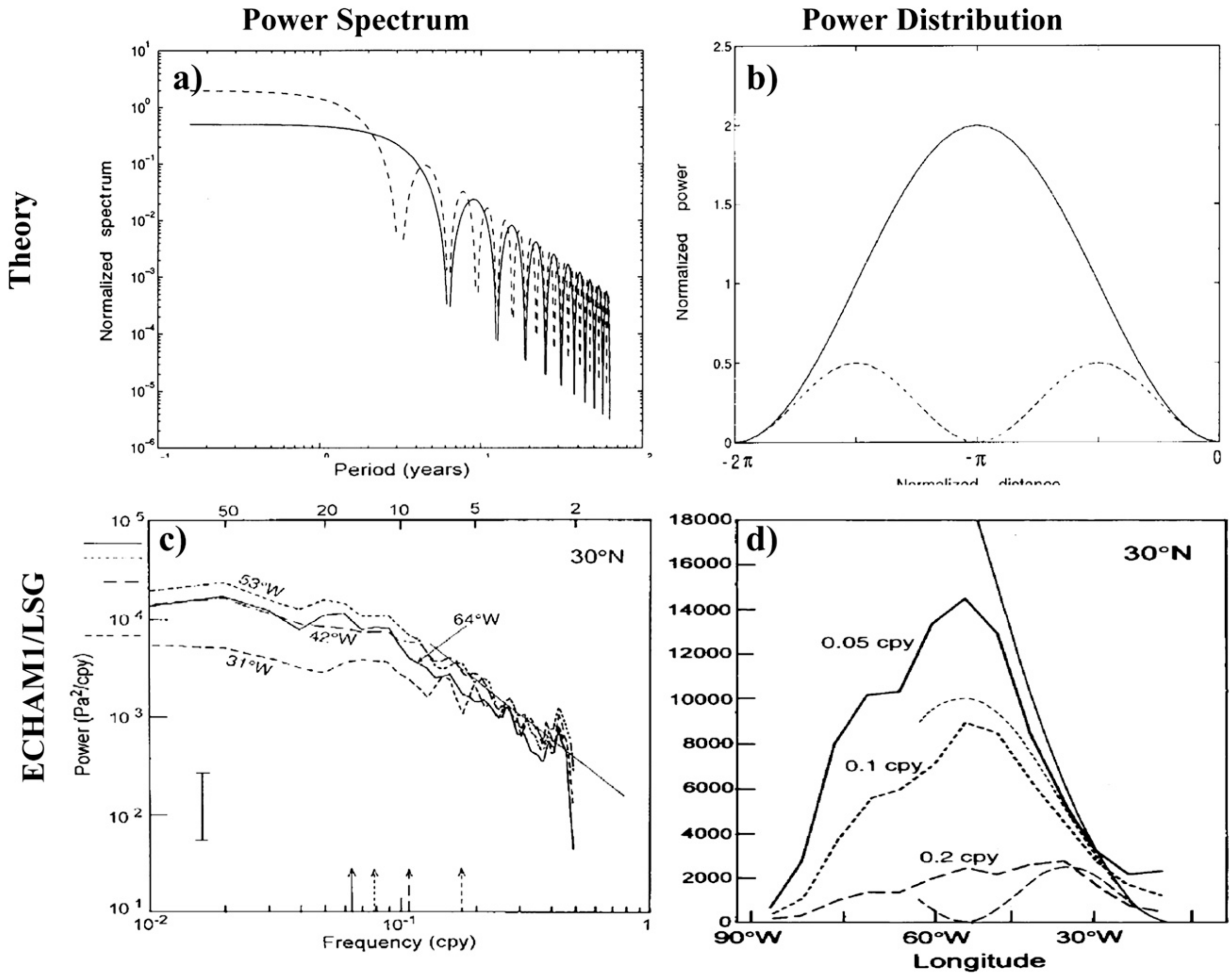

FIG. 6. Wind-driven response in propagative stochastic model. (a) Power spectrum of the baroclinic pressure response as a function of frequency at two locations in the interior ocean. (b) Power of the baroclinic response as a function of distance from the eastern boundary for frequencies ω (solid line) and 2ω (dashed line) where $2\pi\omega$ is the basin crossing time of the planetary wave. (c) Frequency spectrum of a CGCM (ECHAM1/LSG) baroclinic pressure at 250 m, 30°N for various longitudes. The model predictions are given as thin lines for the high-frequency slope. The 95% confidence interval is indicated. (d) Power of the ECHAM1/LSG baroclinic pressure [Pa^2 $(cpy)^{-1}$] at 250 m, 30°N vs longitude in three frequency bands centered around 0.05, 0.1, and 0.2 cpy. The model predictions are given as thin lines. (For details, see Frankignoul et al. 1997.)

c. Role of ocean dynamics

For interdecadal variability, the slow ocean dynamics provides the leading candidate mechanism for the oscillation. Since interdecadal climate variability is usually of basin scale, the relevant ocean dynamics are usually thought to involve large-scale baroclinic Rossby waves in the extratropics. These waves are often called planetary waves because they have spatial scales from hundreds to thousands of kilometers, much larger than the baroclinic deformation radius. These extratropical planetary waves travel slowly, taking years to decades to cross the ocean basin and, therefore, could be relevant for interdecadal climate variability. In comparison, equatorial Rossby waves and Kelvin waves are of seasonal time scales, only providing the memory for interannual variability such as ENSO [see discussion below on (2)]. Extratropical synoptic oceanic Rossby waves generated through instability also have shorter time scales (on the order of seasons) and would act as random noise forcing to interdecadal climate variability.

In the discussion above on Table 2, we mainly discussed the general mechanism giving rise to preferred time scales. The role of ocean dynamics is highlighted as the leading mechanism for the oscillation time scale of interdecadal variability. Here, we further study the specific ocean dynamical processes responsible for interdecadal variability. For convenience we will classify the

TABLE 3. Ocean dynamics for time-scale selection.

Level	Ocean dynamics	Reference
1	Planetary wave propagation	Latif and Barnett 1994; Frankignoul et al. 1997; Jin 1997
2	Basin mode	Cessi and Louazel 2001; Liu 2002a, 2003
3	Advection	Gu and Philander 1997; Yin and Sarachik 1995
4	Planetary wave instability	Colin de Verdière 1986
5	Thermohaline instability	Weaver and Sarachik 1991a,b
6	Positive ocean–atmosphere feedback and coupled instability	Latif and Barnett 1994; Gu and Philander 1997; Chang et al. 1997

role of ocean dynamics in several levels (Table 3). The simplest ocean dynamics is associated with the westward propagation of the first baroclinic planetary wave in the extratropical ocean (level 1). The speed of the planetary wave is proportional to the static stability,

$$N^2 = -\frac{g}{\rho}\frac{\partial \rho}{\partial z},$$

but inversely proportional to the Coriolis parameter f as

$$C_P \sim \beta N^2/f^2 \qquad (2)$$

In which $\beta = df/dy$ is the latitudinal gradient of f. The maximum wave speed is achieved along the equator with a cross-basin time scale of a few months, providing the memory for interannual variability, notably ENSO (Philander 1990; McCreary and Anderson 1991; Neelin et al. 1994, 1998). The wave phase speed decreases rapidly with increasing latitude (i.e., increasing $|f|$) and, to a lesser extent, with reduced stratification N^2. Therefore, in the extratropics planetary waves cross the basin in years to decades, providing a natural time scale for interdecadal variability.

Planetary wave basin modes can be established in an ocean basin because they are not subject to the strong momentum damping along the western boundary (Cessi and Louazel 2001). The time scale of the basin modes is determined by the planetary wave propagation at the poleward boundary (Liu 2002a). Basin-scale stochastic wind forcing can excite basin modes through resonance, generating interdecadal variability with a preferred time scale basinwide (Cessi and Paparella 2001; Cessi and Primeau 2001; Liu 2003; Yang et al. 2004) (Fig. 7), providing another means of interdecadal variability (level 2). However, planetary wave basin modes are dispersed by the differential wave speed with latitude. For the realistic Pacific and Atlantic, which have large meridional extent and therefore larger contrast in wave speed over the basin, planetary wave basin modes are heavily "damped" (Cessi and Primeau 2001; Liu 2003). This damping is further enhanced by baroclinic instability (LaCasce and Pedlosky 2004). Therefore, planetary wave basin modes seem to be inefficient for generating distinctive spectral peaks on interdecadal time scales (Liu 2003; Emile-Geay and Cane 2009), unless longitudinally

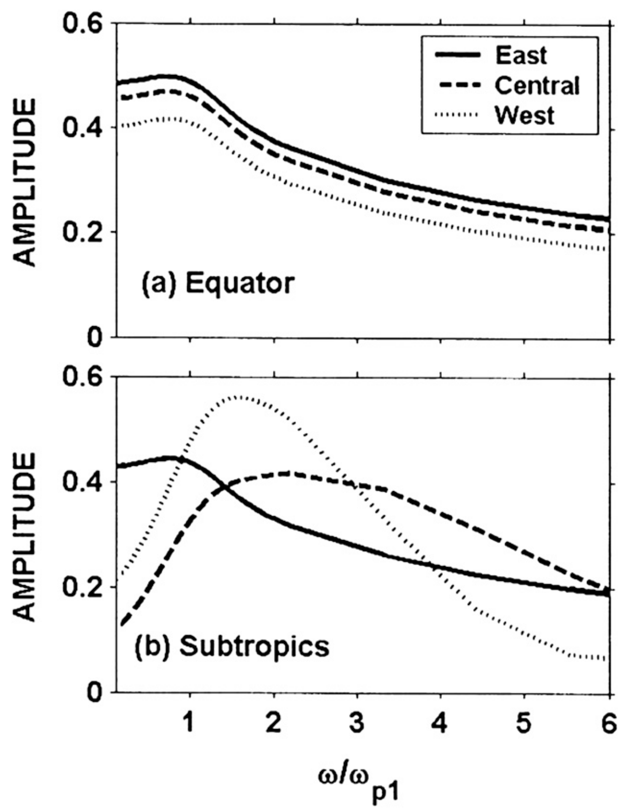

FIG. 7. A numerical solution of the full linear shallow water system with an additional very small linear momentum damping but without the long wave approximation. The model is solved using the finite difference scheme in C grid. The forcing is zonally uniform with the maximum Ekman pumping in the subtropics. Power spectra of the oceanic variability are shown along the (a) equator and (b) subtropics near the eastern boundary (solid line), midbasin (dashed line), and western boundary (dotted line). The forcing frequency ω is normalized by the frequency ω_p. This latter frequency corresponds to the time for the baroclinic wave to cross the basin along the poleward boundary of the basin and is also the frequency of the first basin mode. (Adapted from Liu 2003.)

localized wind forcing is important. Even without basin modes, decadal variability can still be generated locally in the western boundary region, such as the Kuroshio/Oyashio Extension region (KOE), because of the interference of wave propagation across the basin at different latitudes of different wave speeds (for details, see Qiu 2003; Qiu et al. 2007).

Oceanic advection provides another means of interdecadal time selection (level 3). Ocean ventilation has been proposed to provide a mechanism for interdecadal variability (Gu and Philander 1997). Oceanic ventilation of density anomalies is equivalent to the advection of high baroclinic modes of planetary waves (Liu 1999a,b). It has also been argued that the passive ventilation of the spiciness, which refers to temperature and salinity anomalies with compensating effects on density, leads to interdecadal variability in the tropical Pacific (Schneider 2004). The advection of subsurface temperature anomalies by the overturning circulation has also been proposed as a mechanism for driving interdecadal variability in the North Atlantic thermohaline circulation (Greatbatch and Zhang 1995; Yin and Sarachik 1995).

Theoretically, basin-scale circulation can be destabilized by long-wave baroclinic instability (Colin de Verdière 1986), providing a source of interdecadal variability (level 4). Unlike synoptic eddies generated by short-wave baroclinic instability (Charney and Eady modes), which is dominated by interaction between the barotropic mode and the first baroclinic mode, planetary wave instability is caused by the coupling among baroclinic modes (Liu 1999b) and therefore depends more on ambient stratification. While synoptic wave instability occurs mostly in the exit region of western boundary currents where there is strong eastward shear, planetary wave instability tends to occur in regions of westward shear, such as the southwestern part of a subtropical gyre. In a more realistic 3D mean flow with stratification, however, unstable planetary wave basin modes show more complex structures with some resemblance to the planetary wave basin mode. These modes are reminiscent of the interdecadal variability that appears in complex ocean models (Colin de Verdière and Huck 1999; Huck and Vallis 2001; Te Raa and Dijkstra 2002). There are also simple model studies suggesting the importance of oceanic synoptic eddies and the subsequent nonlinear oceanic dynamics in the generation of decadal variability (Kravtsov et al. 2007). In a fully coupled climate model, or the real world, however, it remains difficult to identify the role of planetary wave instability unambiguously, because of the predominance of synoptic waves and associated nonlinear wave–wave interaction.

The interplay between temperature and salinity variability can also lead to thermohaline instability and is a potential mechanism for interdecadal variability (level 5). As far as ocean–atmosphere interaction is concerned, temperature and salinity differ significantly in that temperature is damped heavily through surface turbulent heat loss, while salinity is not. Therefore, the overturning circulation can transport salinity poleward near the surface more effectively than heat, leading to a basin-scale thermohaline instability associated with a positive feedback between the salinity transport and overturning circulation. Stronger overturning transports more salinity to high latitudes, which increases high-latitude density and convection and then strengthens the overturning circulation even more (Stommel 1961). Locally in the subpolar region, the characteristic stratification of fresh cold water over saline warm water also favors the role of salinity variability in the surface. This, in turn, leads to changes in convective instability (Lenderink and Haarsma 1994), which can excite interdecadal variability. This process is sometimes referred to as the heat–salt oscillator mechanism (Welander 1982; Yin and Sarachik 1995), a point to be returned to later.

Finally, at the top level (level 6), ocean dynamics can be coupled actively with the atmosphere to produce self-excited coupled modes, similar to ENSO. One important paradigm is the delayed oscillator paradigm that was originally proposed for ENSO (Suarez and Schopf 1988):

$$\frac{dT(t)}{dt} = aT(t) - bT(t-\delta) - gT^3(t). \qquad (3)$$

The oscillation is excited by a strong positive ocean–atmosphere feedback ($a > 0$), and the cycle is caused by a delayed negative feedback with a delay time δ associated with oceanic wave propagation. The oscillation period is usually longer than twice the delay time. In the case of ENSO, the delay is primarily caused by equatorial Rossby waves with δ being less than a year, giving an interannual ENSO. The delayed oscillator model can be generalized by combining the stochastic theory of Hasselmann (1976) as a stochastically forced model. Now, quasiperiodic variability can be sustained even in the absence of positive ocean–atmosphere feedback ($a < 0$) (e.g., Liu 2002b; Power 2010). The delayed oscillator paradigm, including the generalized version with stochastic forcing, has had a great influence on the study of interdecadal variability because of its conceptual simplicity. The self-excited delayed oscillator paradigm (3) has been used in the early stage to help explain interdecadal variability, with the delay provided by slow extratropical planetary waves (e.g., Latif and Barnett 1994; Marshall et al. 2001). The generalized delayed oscillator with stochastic forcing was also used, even in the case of weak air–sea feedback and, in turn, a damped coupled mode. Indeed, often the stochastically forced delayed oscillator is becoming a default paradigm for decadal variability. In this regard, it should be pointed out that a positive ocean–atmosphere feedback, even if insufficient for generating self-exciting variability, is still helpful in reducing the damping rate of the variability, sharpening the interdecadal spectral peak and, therefore, contributing toward a more distinctive interdecadal variability. Obviously, the nature of ocean dynamics and interdecadal variability is much more complex than the levels described here. Nevertheless, this hierarchy of paradigms still provides a useful and coherent framework for thinking about the mechanisms responsible for interdecadal variability.

4. Pacific interdecadal variability: Tropical origin versus extratropical origin

We now turn specifically to variability in the Pacific and Atlantic. In the Pacific, the study on interdecadal variability has developed along two major lines. One line largely follows the classical ENSO idea in which interdecadal variability is viewed as a self-exciting oscillation generated by a positive ocean–atmosphere feedback and a delayed negative feedback associated with extratropical oceanic Rossby waves. This ''ENSO analog'' reflects, in part, the early thinking of ENSO as a self-exciting oscillation (Philander et al. 1984; Cane and Zebiak 1985; Suarez and Schopf 1988); it also reflects the similarities of the observed patterns of Pacific interdecadal variability to ENSO variability (Zhang et al. 1997). The other line adopts the stochastic climate theory (Hasselmann 1976) and considers interdecadal variability to be forced by stochastic atmospheric internal variability. Regardless of the generation mechanism, however, it is generally agreed that the preferred interdecadal time scale, if there is one, originates from ocean dynamics in the Pacific.

a. *Tropical origin*

One important feature of Pacific interdecadal variability is a strong covariance between the tropics and extratropics. As shown in Figs. 1 and 2, major episodes of decadal and multidecadal climate variability in the North Pacific can also be identified in the tropics. The pattern of interdecadal SST variability resembles that of the interannual variability of ENSO, characterized by two centers of opposite signs in the tropical and North Pacific, although the magnitude is stronger in the North Pacific than in the tropical Pacific for interdecadal variability. The strong negative correlation between tropical and North Pacific SSTs is caused by an effective atmospheric teleconnection from the tropics into the North Pacific through the Pacific–North America (PNA) teleconnection (Alexander 1990; 1992). The presence of strong ENSO variability in the tropical Pacific and the subsequent atmospheric teleconnection toward the extratropics have led to the hypothesis that Pacific interdecadal variability originates in the tropical Pacific (e.g., Trenberth and Hurrell 1994; Newman et al. 2003; Deser et al. 2004).

A tropical-origin mechanism, at first sight, appears to be supported by the statistical modeling of Newman et al. (2003), who showed that the North Pacific interdecadal SST variability can be simulated as a first-order auto-regression (AR1) process after including the remote forcing of ENSO:

$$T_{\mathrm{NP}}(t+1) = aT_{\mathrm{NP}}(t) + bT_{\mathrm{ENSO}}(t) + w(t). \quad (4)$$

This statistical model can generate annual SST variability (Fig. 8a) and a power spectrum consistent with observations (Fig. 8b), implying that the North Pacific SST variability can be generated largely by the local stochastic climate model plus the remote forcing of ENSO. A similar model was proposed for SST variability in the South Pacific (Power and Colman 2006). Note, however, that such models are linear, and any interdecadal variability in the North Pacific linked to ENSO requires decadal variability forcing in the tropical Pacific. Nevertheless, one is still left with the question how interdecadal variability is generated in the tropical Pacific itself and, ultimately, what is the origin of the interdecadal variability.

One simple proposal for a tropical origin of decadal variability attributes the tropical Pacific decadal variability to random changes in ENSO activity from decade to decade. ENSO-like decadal patterns of variability in the tropics can be induced by the nonlinear modulation of ENSO (Münnich et al. 1991; Vimont 2005; Power and Colman 2006) or nonlinear asymmetry between El Niño and La Niña (Burgers and Stephenson 1999). This tropical decadal variability can then propagate into the extratropics through teleconnections, as suggested in the residual model (4). However, robust differences between the interannual and decadal patterns seem to suggest that physics above the simple residual model (4) is needed to explain the difference (Power and Colman 2006). Other mechanisms have also been proposed to account for decadal variability in the tropical Pacific. For example, a diagnosis of sensitivity experiments in a CGCM suggests that the decadal memory in the tropics may arise from higher baroclinic modes of equatorial waves (Wu et al. 2003).

Gu and Philander (1997) proposed a tropical-origin mechanism for multidecadal variability in which the slow time scale is derived from equatorward ventilation from the midlatitude North Pacific. Warm North Pacific surface water reaches the equator in ~20 yr through

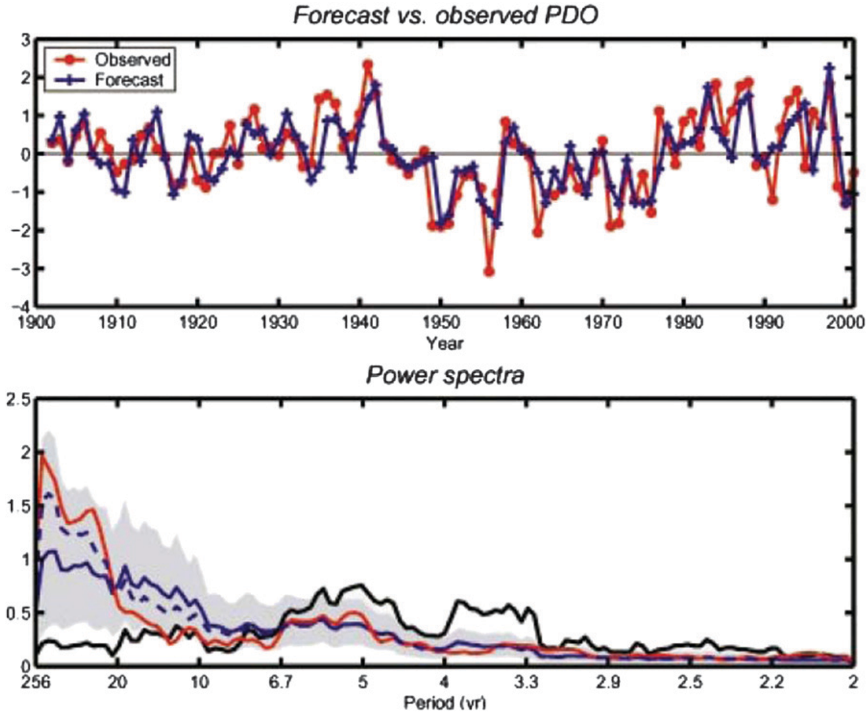

FIG. 8. Null hypothesis model of annual mean PDO as in (3): (top) time series of "forecast" and observed PDO and (bottom) power spectra of observed PDO (red line), ENSO (black line), and ensemble mean "model" PDO (blue line) indices. The 95% confidence interval, shown by gray shading, is determined from the two-sided distribution of one thousand 100-yr samples of the model PDO. The dashed blue line indicates the mean of the 20% (two hundred 100-yr samples) of the model PDO spectra with minimum standard deviation from the observed PDO power spectrum. (Adapted from Newman et al. 2003.)

the equatorward subduction (Fig. 9) (Liu et al. 1994; McCreary and Lu 1994), upwelling along the equator, and warming of the surface equatorial ocean; this surface warming reduces local easterlies, further warming the surface through the positive Bjerknes feedback. The warm equatorial Pacific forces deep atmospheric convection that then propagates into the North Pacific through the PNA teleconnection, intensifying the Aleutian low, midlatitude surface westerlies, and, in turn, turbulent heat flux loss—thus cooling the North Pacific. This forms a delayed negative feedback on the North Pacific SST, which eventually leads to a Pacific multidecadal variability of ~50 yr. This Gu–Philander hypothesis is basically a delayed oscillator paradigm, as in (3), with the positive feedback provided by the Bjerknes feedback in the equatorial Pacific and the delayed negative feedback by the equatorward ventilation. This tropical-origin mechanism, however, was later questioned by Schneider et al. (1999) because the equatorward ventilation of North Pacific warm water (Deser et al. 1996) does not seem to reach the equator. The Gu–Philander hypothesis, however, has not been disproved in the South Pacific because the possibility of equatorward ventilation of an interdecadal temperature anomaly cannot be excluded from the South Pacific (Wang and Liu 2000; Luo and Yamagata 2001).

In short, despite various hypotheses for the tropical origin of interdecadal variability, the proposed mechanisms are difficult to confirm with observations and have not been tested thoroughly in CGCMs.

b. *Extratropical origin for Pacific decadal variability*

Some early studies of PDV followed that of the delayed oscillator paradigm of ENSO, except replacing equatorial Rossby waves with midlatitude Rossby waves and replacing the Bjerknes feedback with midlatitude ocean–atmosphere feedbacks. Latif and Barnett (1994) first proposed a mechanism that attributed the origin of the North Pacific decadal variability to the extratropics. Their CGCM simulated a quasi-20-yr variability in the Pacific (Figs. 10a,b), with some resemblance to the observed PDV (Zhang et al. 1997; Power et al. 1999). The decadal variability shows a clear westward propagation in the subtropical thermocline, reminiscent

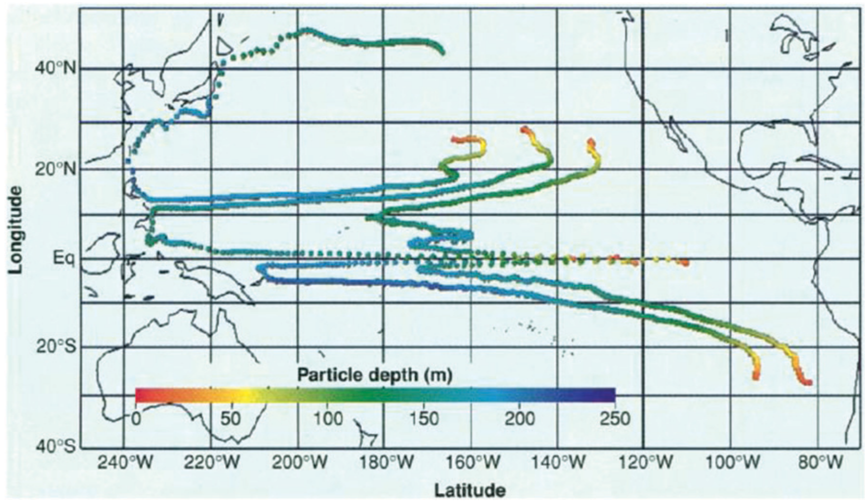

FIG. 9. Paths of water parcels over a 16-yr period after subduction off the coasts of California and Peru as simulated by a realistic OGCM forced by observed climatological winds. From the colors, which indicate the depth of parcels, it is evident that parcels move downward, westward, and equatorward unless they start too far west off California, in which case they join the Kuroshio. Along the equator they rise to the surface while being carried eastward by the swift Equatorial Undercurrent. (Adapted from Gu and Philander 1997.)

of planetary waves (Fig. 10c). Furthermore, a warm Kuroshio–Oyashio Extension SST seems to force a significant ridge response over the Aleutian low. These two features led to the original Latif–Barnett mode hypothesis (Fig. 11a): a warm KOE SST induces a high pressure anomaly in the North Pacific, weakening the Aleutian low and, in turn, the surface westerly in midlatitudes. The reduced wind reduces the turbulent heat flux loss and further warms the North Pacific, leading to a positive ocean–atmosphere feedback. In the meantime, the reduced wind also spins down the subtropical gyre, and in turn the Kuroshio, after a decadal time scale of planetary wave propagation across the basin (Anderson and Gill 1975). The weaker Kuroshio reduces the northward transport of warm water, leading to a delayed cooling in the KOE region and eventually reversing the

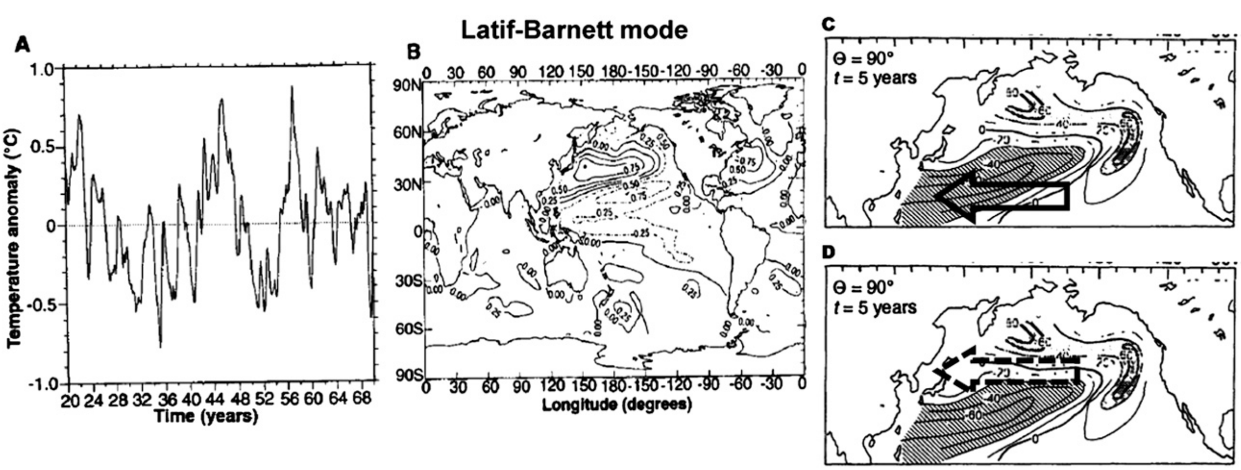

FIG. 10. (a) Time series of the coupled-model anomalous SST (°C) averaged over the region from 25° to 35°N, 150°E to 180°. The time series was smoothed with a 9-month running mean filter. (b) Spatial distribution of linear regression coefficients between the index time series in (a) and SST values. The pattern was scaled to maximum SST anomalies at 1°C. (c) Reconstruction of anomalous heat content (°C m^{-1}) from the leading EOF mode at a particular phase of the decadal cycle. (The phase angle measures the phase of cycle: full cycle = 360°.) The solid double arrow represents schematically the Rossby wave propagation in the subtropics. (d) As in (c), but for the revised Latif–Barnett model, where the dashed double arrow represents the Rossby wave in the midlatitude forced by stochastic forcing. (Adapted from Latif and Barnett 1994.)

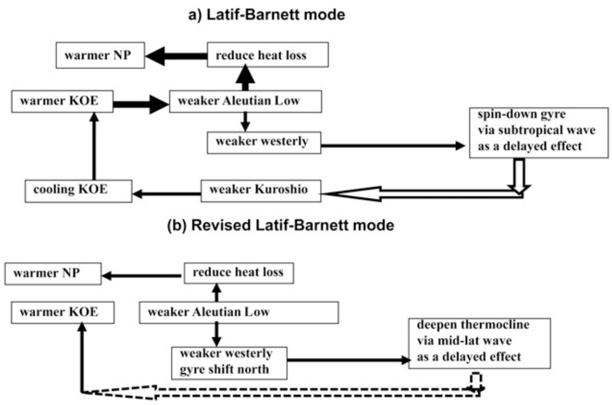

FIG. 11. Schematic diagram for the mechanism of the (a) original and (b) revised Latif–Barnett mode. Double arrows represent the slow wave propagation process for the delayed negative feedback in the (a) subtropics (solid) and (b) midlatitudes (dashed), corresponding to those shown in Figs. 10c and d, respectively.

SST anomaly there into the cold phase. This Latif–Barnett hypothesis is basically a delayed oscillator paradigm, as in (3), with the positive feedback provided by the midlatitude air–sea feedback and the delayed negative feedback by the spindown of the subtropical gyre associated with the planetary wave propagation.

This original Latif–Barnett mode was, however, later found to be largely invalid. Schneider et al. (2002) reexamined the decadal variability simulation of Latif and Barnett (1994) and found that the KOE SST variability lags, rather than leads, the wind variability in the central Pacific by ~5 yr. In the CGCM and observation, in contrast to the eastward advection mechanism by KOE as hypothesized in the original Latif–Barnett mode, the KOE SST variability is induced by westward propagating planetary waves in the midlatitudes, forced by the wind stress curl in the central Pacific (Deser et al. 1999; Schneider and Miller 2001). Reduced westerly wind stresses in midlatitudes then shift the intergyre boundary northward (Seager et al. 2001), inducing a downward Ekman pumping along the intergyre boundary and deepening the thermocline in the central Pacific. The thermocline in the KOE region is then subsequently deepened through the westward propagation of planetary waves. The warmer thermocline water surfaces in late winter owing to deepening of the mixed layer, or the reemergence mechanism (Alexander et al. 1999), thus increasing the KOE SST. Further studies also find that the atmospheric response to typical KOE SST anomalies is not robust (Kushnir et al. 2002; see section 6 of this paper), leaving the positive feedback in the original Latif–Barnett mode questionable as well.

Schneider et al. (2002) modified the Latif–Barnett mode to fit in a propagative stochastic model (tier 2a in Table 2), shown schematically in Fig. 11b. In this revised model, the reduced surface wind still warms the central Pacific by reducing the turbulent heat flux loss, but this warming is caused by stochastic atmospheric variability associated with the Aleutian low instead of the atmospheric response to the warming in the KOE region. This modified Latif–Barnett mode essentially describes a stochastically driven oceanic mode, rather than a coupled mode as described in the original hypothesis. However, both modes share a common feature: decadal variability originates from the extratropical North Pacific (Figs. 10c,d and 11a,b). This modified Latif–Barnett mode seems to be largely consistent with the analysis of variability in the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 2 (CCSM2) (Kwon and Deser 2007), although the role of ocean–atmosphere feedback is thought to be more significant in the latter. The stochastically driven nature of the Pacific decadal variability is also demonstrated in a CGCM study with an interactive ensemble coupling scheme (Yeh and Kirtman 2004, 2006). There, as the internal atmospheric variability is reduced by the ensemble mean of the atmosphere model, the coupled variability is reduced proportionally.

In short, there is evidence in CGCMs that the modified, noncoupled Latif–Barnett mechanism is responsible for some North Pacific decadal variability. In this mechanism, stochastic forcing is important for the generation of the variability, and midlatitude planetary waves determine the decadal time scale. It should be pointed out, however, that in spite of diagnoses consistent with Rossby wave propagation, the role of Rossby waves in Pacific decadal variability has not been explicitly tested in a CGCM.

c. *Extratropical origin for Pacific multidecadal variability*

In contrast to the intensive studies on the decadal (10–20 yr) variability in the Pacific, much less attention has been paid to the multidecadal variability (PMV) (over ~50 yr) in the North Pacific (Fig. 2). It is difficult to explain the multidecadal variability directly using the Latif–Barnett mechanism because the latter involves planetary waves in the subtropics and midlatitudes that cross the basin in less than 10 yr. This time scale, although sufficient for decadal variability, is too short to account for the multidecadal variability.

To identify the origin of the North Pacific multidecadal variability, sensitivity experiments have been performed in two different CGCMs, with consistent results (Liu et al. 2002; Wu et al. 2003; Zhong et al. 2008). Here, we only show the results in CCSM3. First, to isolate the key region of ocean–atmosphere feedback,

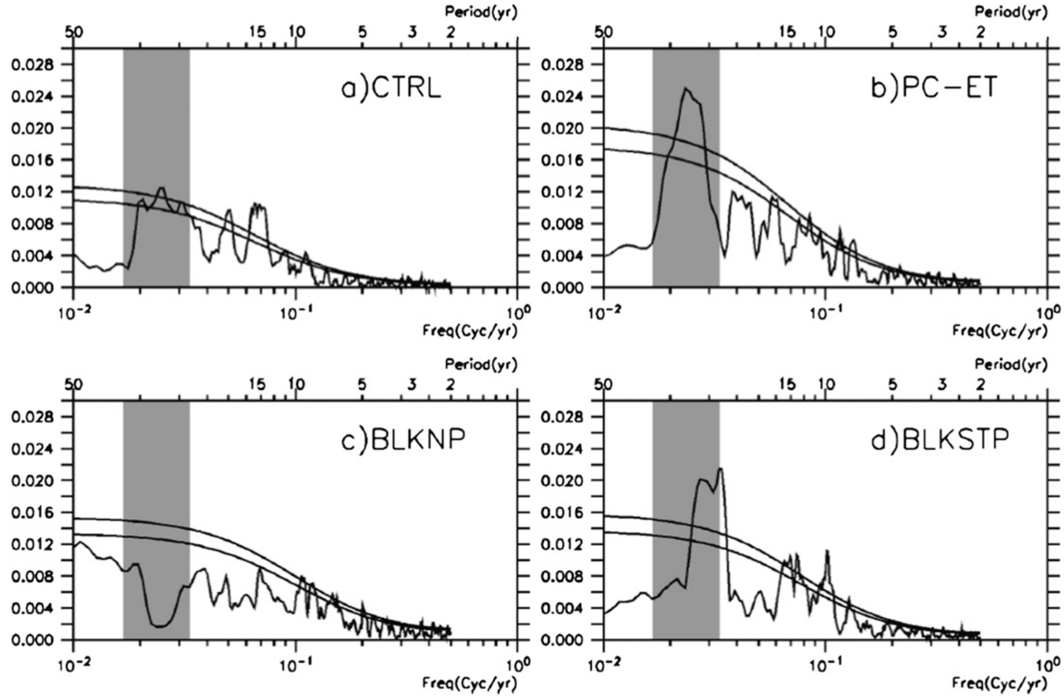

FIG. 12. Power spectra of monthly SST averaged over the western-central North Pacific in four NCAR CCSM3 modeling surgery experiments (each of 400 yr). These experiments are designed to identify the origin of the North Pacific multidecadal variability. (a) Control simulation, which shows a distinct multidecadal spectral peak, (b) PC_ET, in which ocean–atmosphere coupling is suppressed in the tropics (<20°N), (c) BLKNP, as for PC_ET but with an additional sponge layer in the entire North Pacific along the date line, and (d) BLKSTP, as for BLKNP, but with the sponge layer confined in the subtropical North Pacific (<40°N). Vertical shading bar indicates the range of 30–40 yr. The multidecadal peak persists even if ocean coupling is disabled in the tropics [in (b)] and wave propagation is blocked in the subtropics [in (d)], but disappears when the subpolar ocean is also blocked [in (c)]. Similar results can be obtained if the PC1 of the North Pacific SST is plotted. [Adapted from Zhong et al. (2008) and Zhong and Liu (2009).]

a partial-coupling (PC hereafter) sensitivity experiment is performed in which ocean–atmosphere coupling is suppressed in the tropics. The distinctive North Pacific multidecadal variability in the control run (Fig. 12a) remains (Fig. 12b) although the SST variability has been suppressed in the tropics. In contrast, in a complementary PC experiment in which ocean–atmosphere interaction is suppressed in the extratropics, the North Pacific multidecadal mode disappears completely although tropical variability, including ENSO, remains largely unchanged (not shown). These two PC experiments demonstrate that tropical ocean–atmosphere interaction, albeit essential to ENSO, is unnecessary for the multidecadal variability in the North Pacific. To further isolate the role of ocean dynamics in decadal variability, a partial-blocking (PB) sensitivity experiment is performed in which ocean wave propagation between the tropics and extratropics is blocked using a vertical sponge wall in the ocean. Despite the presence of this sponge wall, Pacific multidecadal variability continued (not shown). These experiments demonstrate unambiguously that the origin of the PMV, that is, the process that generates the multidecadal time scale of the variability, in this model is confined to the North Pacific coupled ocean–atmosphere system.

To further understand the role of planetary waves at multidecadal time scales, two additional PB experiments were performed with a north–south oceanic sponge wall across the central North Pacific to block the westward propagation of planetary waves. The North Pacific multidecadal variability disappears over the entire basin when the sponge wall blocks the North Pacific (Fig. 12c). Note that the multidecadal variability disappears not only west of the sponge wall, but also east of the wall. This suggests an active role of the planetary wave in the formation of the multidecadal variability. In contrast, the variability remains similar to the variability in the control run when the sponge wall is confined to the subtropics south of 40°N (Fig. 12d). These two PB experiments demonstrate that the multidecadal variability originates from wave propagation in the subpolar gyre. Planetary waves are slow in the subpolar gyre because of both large

Coriolis parameter and weak stratification [see (2)]. In fact, such waves take more than 20 yr to cross the basin (Zhong and Liu 2009). This provides a sufficiently long time scale for the role of planatary waves in driving multidecadal variability. Further analysis found that a warm KOE SST induces a ridge response and, in turn, downward Ekman pumping in the subpolar gyre. The downward Ekman pumping forces a southward Sverdrup flow and, in turn, a cold anomaly in the thermocline. This cold anomaly appears to migrate westward and is eventually transported southward by the Oyashio toward the KOE region, generating a delayed negative feedback. This analysis shows evidence of a positive ocean–atmosphere feedback associated with the atmospheric response to the KOE SST, although it remains unclear if the net ocean–atmosphere feedback in the midlatitudes is positive or not. Therefore, this mode is likely to be stochastically driven, with ocean–atmosphere feedback only modifying the variability.

It is interesting to note that the variability at the surface in the model used by Zhong et al. (2008) is highly correlated between the North and tropical Pacific for PMV (Zhong et al. 2008), as in the observations (Deser et al. 2004). This is an important feature that led to the tropical-origin hypothesis for PMV as discussed earlier. However, the experiments of Zhong et al. explicitly demonstrated otherwise in their model (the NCAR CCSM3). One lesson is that we should be cautious in judging the origin of interdecadal variability from the diagnostic relationship of surface climate. Indeed, there are studies that imply a teleconnection of the opposite direction for the multidecadal variability: from the extratropics to the tropics, as discussed below.

Extratropical climate variability can also impact the tropics through teleconnections via atmospheric, oceanic, and coupled processes (Liu and Alexander 2007, and references therein). Extratropical climate variability can affect tropical climate through the atmospheric transport of latent and sensible heat (e.g., Liu and Yang 2003; Chiang and Bitz 2005). Boundary layer ocean–atmosphere coupling through the wind speed–evaporation–SST (WES) feedback can generate a rapid equatorward propagation in the trade wind regime (Liu and Xie 1994), enabling extratropical climate variability to affect the tropics in the so-called seasonal footprinting mechanism (Vimont et al. 2003). These fast atmospheric and coupled teleconnections apply not only to multidecadal variability but also to decadal, interannual (Vimont et al. 2003), and even seasonal (Liu and Xie 1994) variability in the Pacific and the other oceans (Tanimoto and Xie 1999). These equatorward teleconnections are particularly important at the decadal time scale because they can transmit the long-term variability signal from the extratropics into the tropics, generating a long-lasting impact. These equatorward teleconnections are likely to be responsible for the global hyper mode, a decadal variability mode of global extent, but generated by ocean–atmosphere coupling in the absence of ocean dynamics (Dommenget and Latif 2008). At the interdecdal time scale, tropical variability can be further affected by the extratropical climate changes through the equatorward oceanic ventilation (Liu and Yang 2003; Wu et al. 2007) and coastal trapped waves (McGregor et al. 2008).

d. Summary

Pacific climate exhibits multiple modes of interdecadal variability. All interdecadal variability modes exhibit coherent variability in the tropical and North Pacific although, relatively, the decadal variability tends to be more prominent in the tropical–subtropical region while the multidecadal variability tends to be more prominent in the North Pacific. The fact that variability in the tropical Pacific and North Pacific is coherent has led to two groups of hypotheses: one attributes the origin of the variability to the tropics and the other to the extratropics. The relative importance of tropical and extratropical processes in driving interdecadal variability is difficult to clarify through the diagnosis of the surface climate alone. It is much easier to do this in CGCMs using sensitivity experiments. Such experiments show that the multidecadal variability in the North Pacific originates from the extratropics, and more precisely the subpolar North Pacific. The origin of the Pacific decadal variability mode, on the other hand, remains less clear. Finally, all interdecadal variability is most likely stochastically driven, rather than self-exciting.

5. Atlantic variability: Thermohaline and wind driven

Unlike the Pacific where interdecadal variability occurs largely in the upper-ocean wind-driven circulation, much of the interdecadal variability in the Atlantic is often thought to be associated with an active thermohaline circulation. Indeed, early work on interdecadal variability in the Atlantic arises from studies of thermohaline equilibria and stability. As such, these studies point to yet another source of interdecadal variability.

a. Multidecadal variability and thermohaline mechanism

1) OGCMs WITH IDEALIZED SURFACE BOUNDARY CONDITIONS

Early studies of interdecadal variability in the thermohaline were carried out in OGCMs with two types of idealized surface boundary conditions (BCs): mixed boundary conditions [a restoring BC for SST and flux BC for sea surface salinity (SSS)] and the flux–flux boundary condition (flux BC for both SST and SSS). Mixed boundary conditions are based on an understanding that a SST anomaly can affect the atmosphere and, in turn, the surface turbulent heat flux, leading to a strong thermal damping of the SST anomaly [a few months for a typical surface ocean mixed layer, Haney (1971)]. In contrast, a SSS anomaly has no direct impact on the atmosphere. Mixed BCs were originally designed to study thermohaline instability and multiple equilibria (Stommel 1961; Rooth 1982) in OGCMs (e.g., Bryan 1986; Marotzke and Willebrand 1991; Rahmstorf 1995; Power et al. 1994; Power and Kleeman 1993, 1994). In these OGCMs, a weak (strong) freshwater flux at high latitude tends to force a stable thermal (saline) mode that sinks in the high (low) latitude. However, for an intermediate strength of freshwater forcing, self-sustained oscillation could emerge. Weaver and Sarachik (1991a,b) first examined self-exciting interdecadal Atlantic meridional overturning circulation (AMOC) variability in an idealized sector basin of a flat bottomed OGCM under mixed BCs. Similar interdecadal variability was produced later in many OGCMs (e.g., Weaver et al. 1991, 1993; Myers and Weaver 1992; Yin and Sarachik 1995; Tziperman et al. 1994; Kravtsov and Ghil 2004; Arzel et al. 2006) (Fig. 13).

The flux–flux boundary condition originates from the idea that low frequency variability involves basin-scale SST anomalies that tend to be damped weakly by longwave radiation, rather than strongly by turbulent surface heat fluxes (Bretherton 1982). Therefore, the surface forcing by SST should act more like heat flux than a restoring (Zhang et al. 1993; Power and Kleeman 1993; Rahmstorf and Willebrand 1995). From the heat capacity viewpoint, the flux–flux boundary condition is equivalent to an atmosphere of zero heat capacity, which seems to be more reasonable than the mixed boundary condition, which is equivalent to an atmosphere of infinite heat capacity. Under the flux–flux boundary condition, the temperature and salinity equations can be combined approximately into a single equation for density, effectively eliminating the thermohaline instability under the mixed boundary condition. Self-exciting interdecadal variability has also been found in many models under the flux–flux boundary condition (e.g., Greatbatch and Zhang 1995; Huang and Chou 1994; Zhang et al. 1995; Winton 1996; Chen and Ghil 1996; Arzel et al. 2006) (Fig. 13). Interdecadal oscillations exhibit different spatial patterns under the two types of BCs. The pattern of variability under flux–flux BCs seems to resemble the interdecadal variability in complex CGCMs with the maximum centered in the northwest North Atlantic (Greatbatch and Zhang 1995; Arzel et al. 2006).

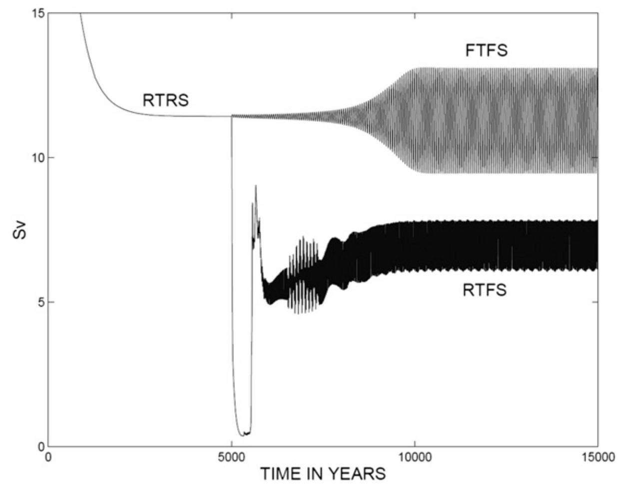

FIG. 13. Maximum meridional overturning streamfunction (Sv) for the ocean model running under restoring boundary conditions [restoring for temperature and restoring for salinity (RTRS)] until year 5000. Mixed boundary conditions [restoring for SST but flux for SSS (RTFS)] and constant flux [flux for temperature and flux for salinity (FTFS)] boundary conditions are applied at year 5000 of the integration, both showing significant decadal variability. (Adapted from Arzel et al. 2006.)

Interdecadal oscillations are caused by different mechanisms under the two types of boundary conditions and will be reviewed only briefly here [see Arzel et al. (2006) for a comparison study]. For mixed boundary conditions, decadal variability is generated by an enhanced convective instability in the subpolar ocean and a slow northward advection of warm subsurface water as a delayed negative feedback [for more details, see Yin and Sarachik (1995)]. This combined advective–convective mechanism can be traced back to the flip-flop heat–salt oscillator in previous theoretical models (Welander 1982; Lenderink and Haarsma 1994; Yin 1995). In contrast, under the flux–flux boundary condition, large-scale baroclinic instability (Colin de Verdière 1986) seems to be responsible for the excitation of the variability (Colin de Verdière and Huck 1999; Huck and Vallis 2001; Te Raa and Dijkstra 2002; Arzel et al. 2006), while the oceanic heat transport and the resulting oceanic warming at high latitude provides a delayed negative feedback on the AMOC (Huang and Chou 1994; Greatbatch and Zhang 1995).[3] There is also growing recognition that significant decadal thermohaline variability can be generated by stochastic wind and buoyancy forcing as damped oceanic

[3] The mechanism for the interdecadal variability under the flux–flux boundary condition, such as large-scale baroclinic instability, in principle should also work in the Pacific and other oceans because it does not require an active thermohaline circulation.

modes (Mikolajewicz and Maier-Reimer 1990; Weisse et al. 1994; Griffies and Tziperman 1995; Saravanan and McWilliams 1997; Neelin and Weng 1999; Delworth and Greatbatch 2000; Eden and Greatbatch 2003).

In spite of intensive studies on thermohaline interdecadal variability in OGCMs of idealized surface forcing, the relevance of such variability—especially those self-exciting thermohaline oscillations—to the real world, or even a fully coupled CGCM, has remained elusive. For example, some studies suggest that self-exciting thermohaline variability found under these idealized surface boundary conditions can be damped severely by bottom topography (Winton 1997). However, self-exciting thermohaline interdecadal variability has also been generated in some OGCMs of realistic topography under the mixed BCs (Weaver et al. 1994; Weisse et al. 1994; Griffies et al. 2009) and flux–flux BCs (Huck et al. 2001; Te Raa et al. 2004). Further study on this issue is needed.

2) Multidecadal variability in a CGCM

The study of multidecadal variability in a CGCM is much more challenging than in an OGCM under idealized surface forcing because of the complex nature of ocean–atmosphere coupling and the relatively higher computational cost. Delworth et al. (1993) presented the first thorough analysis of multidecadal variability in the Atlantic in a CGCM (GFDL_R15). The variability they described resembled the observed Atlantic multidecadal oscillation (AMO) with a period of ~50 yr. The multidecadal SST variability was driven by the AMOC, with a stronger AMOC increasing the northward heat transport and leading to a subsequent monopole warming over the North Atlantic (Fig. 3a). The increased AMOC also exported heat out from the South Atlantic, cooled the South Atlantic, and formed an interhemisphere dipole in SST. The variability of the AMOC in the model of Delworth et al. appeared to be driven by density anomalies in the sinking region over the Laborador Sea, which is contributed predominantly by salinity and leads the AMOC transport by ~5 yr (Fig. 14a). The poleward salt transport that is associated with an upper-ocean recirculation gyre is suggested to be important in initiating the variability because it leads both the heat and overturning transports significantly; this is consistent with the nearly in-phase salinity and overturning transport and the fact that salinity leads temperature by ~10 yr (the negative of density attributed by temperature) (Fig. 14a). The oscillation cycle appeared to be generated by the interplay of heat and salt transports. A weaker AMOC reduced the heat transport, resulting in a cold and dense pool in the upper subpolar region. This cold and dense pool induced a cyclonic gyre, which enhanced the northward salt transport, increasing the salinity and, in turn, the density in the sinking region, thus intensifying the AMOC. The role of ocean–atmosphere feedback is further clarified in sensitivity experiments using the ocean-alone model forced by the surface fluxes diagnosed from the coupled simulation (Delworth and Greatbatch 2000). They found that major features of the multidecadal variability can be reproduced by the surface heat flux forcing (Fig. 15a), but not the freshwater flux. This is consistent with a regression analysis on the overturning transport, which shows a regression on the surface heat flux comparable with that on the heat transport, but a regression on the freshwater flux negligible relative to that on the salt transport. Furthermore, the preferred multidecadal time scale can be reproduced when the surface heat flux is randomized in time (Fig. 15b). Delworth and Greatbatch therefore concluded that the AMO is forced by the stochastic surface heat flux as a damped oscillatory ocean mode, with ocean–atmosphere feedback playing a minor role. This conclusion is consistent with a theoretical study in a four-box ocean model under mixed BCs where the variability of various fields exhibits similar lead–lag relationships as in the CGCM (Fig. 14c, Griffies and Tziperman 1995). In this model, sensitivity experiments show that the heat flux is the dominant stochastic forcing.

Atlantic multidecadal variability is later simulated in other CGCMs but has been proposed to be caused by different mechanisms. Some models suggested a damped oceanic mode forced by stochastic forcing (e.g., Jungclaus et al. 2005; Danabasoglu 2008), as in Delworth and Greatbatch (2000). Others highlight the role of ocean–atmosphere feedback. In the ECHAM1/LSG coupled climate model, Timmermann et al. (1998) simulated Atlantic multidecadal variability similar to that of Delworth et al. (1993), with the density in the sinking region leading the overturning transport and the density contributed mainly by salinity (Fig. 14b). However, their analysis suggests a critical role of ocean–atmosphere feedback. A stronger AMOC increases salt transport into the sinking region and then enhances the AMOC further, forming a positive feedback. The increased AMOC also increases heat transport, warming the North Atlantic. The warm North Atlantic forces an atmospheric response with a strengthened NAO and, in turn, a reduced evaporation and Ekman transport off Newfoundland and in the Greenland Sea. The reduced salinity then weakens the convection and eventually the AMOC, forming a delayed negative feedback. Vellinga and Wu (2004) studied centennial variability in the third climate configuration of the Met Office Unified Model (HadCM3) and suggested another critical role of air–sea coupling. An increased AMOC warms the North Atlantic, which shifts the ITCZ northward, leading to

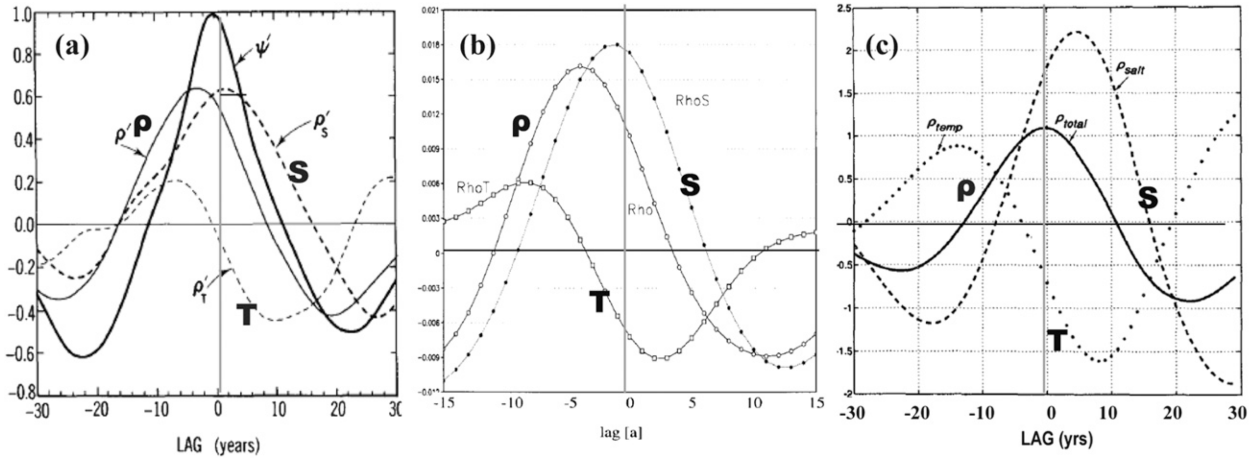

FIG. 14. Lagged correlation/regression between the AMOC or annual surface total density and its contribution by salinity and temperature in the North Atlantic sinking region (averaged vertically and horizontally). (a) Correlation in the GFDL_R15 model (Delworth et al. 1993), (b) regression in the ECHAM1/LSG (Timmermann et al. 1998), and (c) regression in a four-box model (Griffies and Tziperman 1995). The correlation with the AMOC transport is also shown in (a) as the thick black curve. The thin vertical line in each panel marks the zero lag: a negative lag means a lead time prior to the AMOC. See the respective paper for details.

greater rainfall in the tropical North Atlantic. The resulting freshwater is eventually transported to the subpolar North Atlantic, providing a delayed negative feedback that weakens the AMOC.

The analysis of ocean–atmosphere feedbacks in a control simulation is important for understanding decadal variability, even if the variability is caused by a damped oceanic mode. However, there is a lesson here, as there was in our earlier discussion of the Pacific: it is generally difficult in a CGCM to clarify the role of ocean–atmosphere feedback and to distinguish a coupled mode from a damped oceanic mode unambiguously, based on the diagnosis of a control simulation alone. For example, in Timmermann et al. (1998), the role of atmospheric feedback and an active coupled mode is not demonstrated explicitly, unlike in Delworth and Greatbatch (2000). Instead, the evidence of atmospheric response to SST variability is a simultaneous coherence between the atmospheric variability in the North Atlantic and North Pacific. Similar comments can be made regarding the analysis by Vellinga and Wu (2004). Even using the ocean-only model, the results could vary depending on the formulation of the surface boundary condition. For example, Weaver and Valcke (1998) found little multidecadal variability in the OGCM of Delworth et al. (1993) under idealized surface forcing and therefore

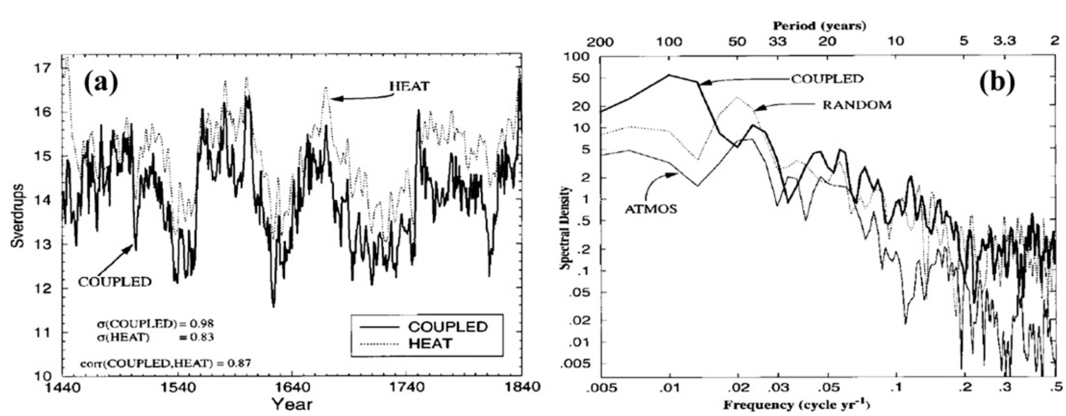

FIG. 15. Comparison of the AMOC in the coupled model with the corresponding OGCM simulations forced by various forcing schemes. (a) Time series of the AMOC from the coupled model (thick solid line, COUPLED) and the ocean model driven by time series of annual mean heat flux anomalies from the coupled simulation (dashed line, HEAT). (b) Power spectra of the coupled model (thick solid line; COUPLED), the ocean models driven by randomized surface flux forcing (dotted line, RANDOM), and the atmospheric surface fluxes to the ocean in the coupled model (thin solid line, ATMOS). (Adapted from Delworth and Greatbatch 2000.)

concluded that the model multidecadal varaibility is a truly coupled mode excited by positive ocean–atmosphere feedback, opposite to Delworth and Greatbatch (2000). However, the variability of the flux forcing in Weaver and Valcke is an order of magnitude smaller than that in Delworth and Greatbach, owing to a more idealized formulation of the surface boundary condition in the former.

Finally, it is important to point out that, although the origin of Atlantic multidecadal variability seems to lie in the North Atlantic, it can impact the tropical Atlantic and global climate remotely at decadal and longer time scales through atmospheric waves, oceanic Kelvin waves, and coupled oceanic–atmospheric propagation, as discussed for the Pacific by Kleeman and Power (1999), Dong and Sutton (2002), Sutton and Hodson (2005), Chiang and Bitz (2005), Timmermann et al. (2005), Latif et al. (2007), and Zhang et al. (2007).

b. Decadal variability and wind-driven mechanism

In addition to the thermohaline circulation, there are other mechanisms that rely on wind-driven circulation for decadal variability in the Atlantic. Decadal variability in the Atlantic has been proposed to originate in both the tropics and extratropics. Chang et al. (1997) proposed a mechanism that attributes the decadal variability in the tropical Atlantic to coupled ocean–atmosphere feedback in the tropical Atlantic (Fig. 16). Their mechanism largely follows a delayed oscillator paradigm with the WES feedback providing the positive feedback and advection by the Brazil Current providing the delayed negative feedback. A dipole of the warm/cold SST anomaly in the northern/southern tropical Atlantic induces a pressure gradient force northward across the equator, reducing/increasing the northeasterly/southeasterly trade wind in the northern/southern tropical Atlantic, and in turn, further warming/cooling the northern/southern tropical Atlantic through the WES feedback. The increased southerly wind south of the equator also intensifies the northward Brazil Current, advecting cold water from the southern into the northern tropical Atlantic to provide a delayed negative feedback, leading to a decadal oscillation in the tropical Atlantic. Further AGCM simulations suggest that the tropical dipole SST can force a positive NAO response in the North Atlantic through atmospheric teleconnection (Sutton et al. 2001; Terray and Cassou 2002). This teleconnection enables the tropical decadal variability to impact the entire Atlantic, forming a pan-Atlantic decadal variability (Fig. 4).

Atlantic decadal variability has also been proposed to be forced by the extratropical North Atlantic. The extratropical origin for TAV is consistent with the observations that Atlantic decadal variability is preceded by sea ice variability in Hudson Bay (Deser and Blackmon 1993) and the northern tropical Atlantic SST variability is driven largely by the NAO (Czaja et al. 2002). Indeed, all of the wind-driven mechanisms for North Pacific decadal variability discussed in section 4 could, in principle, be applied to the North Atlantic, after halving the time scale to account for a narrower Atlantic basin. The propagation stochastic model (tier 2 in Table 2) has been used to account for the observed variance of decadal variability in the North Atlantic with some success (Sturges and Hong 1995; Frankignoul et al. 1997; Sturges et al. 1998). The Latif–Barnett mechanism, in its original or modified form, could also be applied to the North Atlantic. In a modified propagation stochastic model, Marshall et al. (2001) proposed a mechanism for North Atlantic decadal variability in a delayed oscillator paradigm, modified from the original Latif–Barnett hypothesis, now with the emphasis on the meridional shift of the intergyre boundary.

Eden and Greatbatch (2003) proposed a combined wind–thermohaline mechanism for decadal variability and interpreted the decadal variability as a damped mode excited by stochastic forcing. In a hybrid coupled model with a simple statistical atmosphere coupled to an OGCM, a SST dipole of warm subpolar/cold subtropical North Atlantic reduces the NAO and produces a cyclonic wind stress curl in the western North Atlantic. This wind curl anomaly forces an instantaneous barotropic gyre circulation that enhances the heat transport across the intergyre boundary, forming a positive feedback on the SST dipole. This positive feedback from the wind-driven circulation reduces the damping rate of the oceanic mode and therefore enhances the variance of the decadal variability. The warm subpolar SST then reduces the AMOC and, in turn, the northward heat transport, leading to a delayed negative feedback.

Finally, decadal variability in the extratropical North Atlantic can propagate equatorward through the coupled WES teleconnection (Liu and Xie 1994), as suggested in the Pacific (Vimont et al. 2003). This results in a pan-Atlantic decadal variability (Tanimoto and Xie 1999, 2002). A review of tropical Atlantic variability can be found in Xie and Carton (2004).

c. Summary

It has been suggested that multidecadal climate variability in the Atlantic region is associated with the changes in thermohaline circulation. Earlier studies with OGCMs under highly idealized surface BCs, mostly with flat bottoms, show self-exciting interdecadal variability generated in the Atlantic through either thermohaline instability under mixed BCs or baroclinic instability under the flux–flux BC. The time scale of the multidecadal variability appears to be determined by the northward advection of the return branch of the AMOC. In CGCM

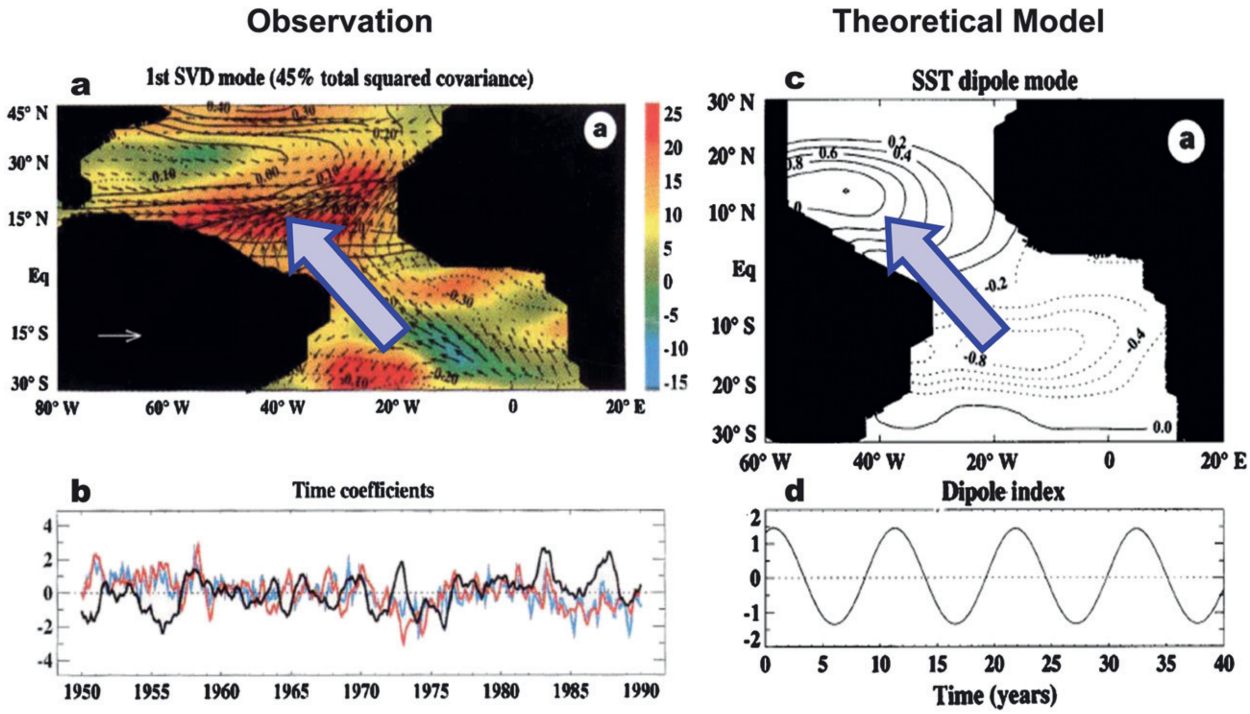

FIG. 16. Observed and modeled Atlantic dipolelike variability. The observed dipole variability is revealed by a joint SVD analysis between SST and surface wind stress. (a) Spatial structure of the first SVD mode; contours represent SST anomaly (°C) and vectors depict wind stress anomalies. (b) The two associated time coefficients (red line, first SVD of the Atlantic SSTA; blue line, first SVD of the atmospheric variation), both normalized by its standard deviation. The black line in (b) is the normalized time coefficient of the first EOF of the global SST anomaly. (c) The simulated dipole variability from an intermediate coupled model is shown, its pattern as regressed from the SST to (d) the dipole index, which is calculated as the difference of SST between the northern and southern tropical Atlantic. The North Brazil Current is depicted schematically by the blue arrow in (a) and (c). (Adapted from Chang et al. 1997.)

studies, multidecadal variability exhibits an interhemisphere dipole in SST and is found to be associated with the fluctuation of the AMOC. The multidecadal variability has been proposed to be caused either by coupled feedback or stochastic forcing on the AMOC, although it is generally difficult in a CGCM to identify the nature of the variability unambiguously. Some sensitivity experiments using CGCMs suggest that the multidecadal variability is generated as a damped oceanic mode forced by stochastic heat flux, with ocean–atmosphere feedback playing a minor role.

Decadal variability in the Atlantic exhibits a pan-Atlantic pattern, with a tripole in the North Atlantic and another pole in the South Atlantic. This decadal variability has been proposed to be caused by several mechanisms, including ocean–atmosphere feedback and wind-driven circulation in the tropics and coupled feedback through the wind or combined wind–thermohaline mechanism in the extratropics.

6. Ocean–atmosphere feedback in the extratropics

A review of interdecadal variability is not complete if it does not address ocean–atmosphere interaction in the extratropics because the mechanism responsible for interdecadal variability often involves ocean dynamics and ocean–atmosphere feedbacks outside the tropics. Furthermore, interdecadal SST variability exhibits large variance in the extratropics (Boer 2000) and therefore its climate impact needs to be understood.

In studying ocean–atmosphere feedbacks in the extratropics, the most challenging issue is the atmospheric response to extratropical SST. In contrast to the linear stationary baroclinic atmospheric response to tropical SST forcing (Gill 1980; Lindzen and Nigam 1987), the atmospheric response to extratropical SST variability remains poorly understood (Frankignoul 1985; Kushnir et al. 2002). This has undermined our understanding of ocean–atmosphere feedback in the extratropics and, in turn, its role in interdecadal variability. The study of the atmospheric response to extratropical SST has been challenging in both observations and modeling. The atmospheric response to extratropical SST anomalies is difficult to assess from observations alone because of the overwhelming climate noise of atmospheric internal variability. Modeling the atmospheric response to extratropical SST variability is also difficult because the response involves synoptic eddies and the associated strongly nonlinear wave–wave and wave–mean flow interactions. These interactions are difficult to simulate in simplified models with sufficient realism. Thus, an AGCM has to be used. Early AGCM modeling studies on the atmospheric responses to extratropical SST anomalies used the approach to prescribe the SST anomaly to force an AGCM (e.g., Palmer and Sun 1985). However, these model simulations often differ significantly from each other, from a warm SST–ridge response of equivalent barotropic structure to a warm SST–low response of baroclinic structure [see Kushnir et al. (2002) for an extensive review].

Recent studies show some evidence of an atmospheric response to extratropical SST variability. Special attention has been paid to the climate impact of the North Atlantic SST variability that is associated with the AMO and, in turn, the AMOC (e.g., Dong and Sutton 2002; Sutton and Hodson 2005; Zhang and Delworth 2005; Latif et al. 2006, 2007). Long-term observations show significant climate anomalies associated with the North Atlantic SST variability. For example, a warm decadal SST over the North Atlantic (Fig. 17, top right) is found to correspond to low sea level pressure and warm surface temperature over North America and western Europe, and a northward migration of the tropical rainbelt, notably over North Africa [Figs. 17a(1–3)] (Sutton and Hodson 2005). This atmospheric pattern is largely reproduced in an AGCM forced by the observed SST variability over the North Atlantic [Figs. 17b(1–3)]. Further sensitivity experiments show that most of the atmospheric anomaly is forced by tropical (equatorward of 30°N) SST anomalies, although there is a significant response over Europe forced by the extratropical SST anomalies (Sutton and Hodson 2005). In general, however, for the extratropics where internal atmospheric variability is dominant, caution is needed when inferring causality from the simultaneous correlation between the atmosphere and SST [e.g., Figs. 17a(1–3)] because it tends to mix the atmospheric response to SST forcing with the SST response to atmospheric variability (Frankignoul et al. 1998), a point to be returned to later. For the same reason, one should also be careful in interpreting the corresponding Atmospheric Model Intercomparison Project (AMIP) experiments with prescribed SST (Bretherton and Battisti 2000).

As an alternative to the SST-forced AGCM experiment, the oceanic influence on the atmosphere is simulated by applying surface heat flux forcing, reflecting the idea that the subsurface ocean influences the atmosphere through the release of heat (e.g., Yulaeva et al. 2001). In Figs. 17c(1–3), a heat flux anomaly confined to the extratropical North Atlantic (30°–60°N) is used to force the ocean in a coupled AGCM–Slab Ocean Model. The response forced by the heat flux in the coupled model captures the major features of the atmospheric responses in the observation and the AMIP experiment, including the reduced sea level pressure and northward migration of the rainbelt (Delworth et al. 2007). Note that, even though the heat flux anomaly is confined to latitudes north of 30°N, the resulting warm surface air temperature (and SST, not shown) extends into the tropics [Fig. 17c(3)]. This is because the response excites both atmospheric and coupled teleconnections, as discussed above. Therefore, the extratropical oceanic heat flux anomaly induces an atmospheric response and, in turn, a heat flux response to the SST anomalies in both the tropics and extratropics. This approach likely accounts for part of the effect of ocean–atmosphere coupling on the atmospheric response.

In observations or a CGCM control simulation, the atmospheric response to extratropical SST is difficult to extract because of the presence of strong internal atmospheric variability and complex interactions between the atmosphere and ocean, as noted above. In principle, the oceanic influence on the atmosphere should be evaluated using lagged covariance calculations where ocean leads the atmosphere because this calculation will remove the forcing effect of internal atmospheric variability on SST (Frankignoul et al. 1998). One approach is lagged singular value decomposition (SVD) analysis, or maximum covariance analysis (MCA) (Czaja and Frankignoul 2002), which shows evidence of a winter NAO atmospheric response to late fall North Atlantic SST variability (Czaja and Frankignoul 2002) and a summer North Pacific atmospheric response to spring North Pacific SST variability (Liu et al. 2006; Frankignoul and Sennéchael 2007). Quantitatively the oceanic influence on the atmosphere can be assessed as a ratio of lagged covariance using the so-called equilibrium feedback analysis (EFA) (Frankignoul et al. 1998) and its multivariant generalization generalized equilibrium feedback analysis (GEFA) (Liu et al. 2008). The statistical assessment of the atmospheric response using EFA can be validated explicitly in sensitivity experiments using a CGCM, as in the example of atmospheric responses to KOE SST (left and middle columns of Fig. 18). An ensemble coupled initial value experiment is desirable because it takes full account of ocean–atmosphere coupling (Liu and Wu 2004). Once the statistical assessment is found sufficiently robust and accurate in a complex CGCM, the statistical method can be used to assess the atmospheric response in the observations with certain confidence. In the case of the atmospheric response to KOE SST variability, the assessed atmospheric response exhibits a warm SST–ridge response (Fig. 18, right column) that is dominated by the winter response (Liu et al. 2007).

Overall, studies so far suggest that the magnitude of the atmospheric response to extratropical SST variability is

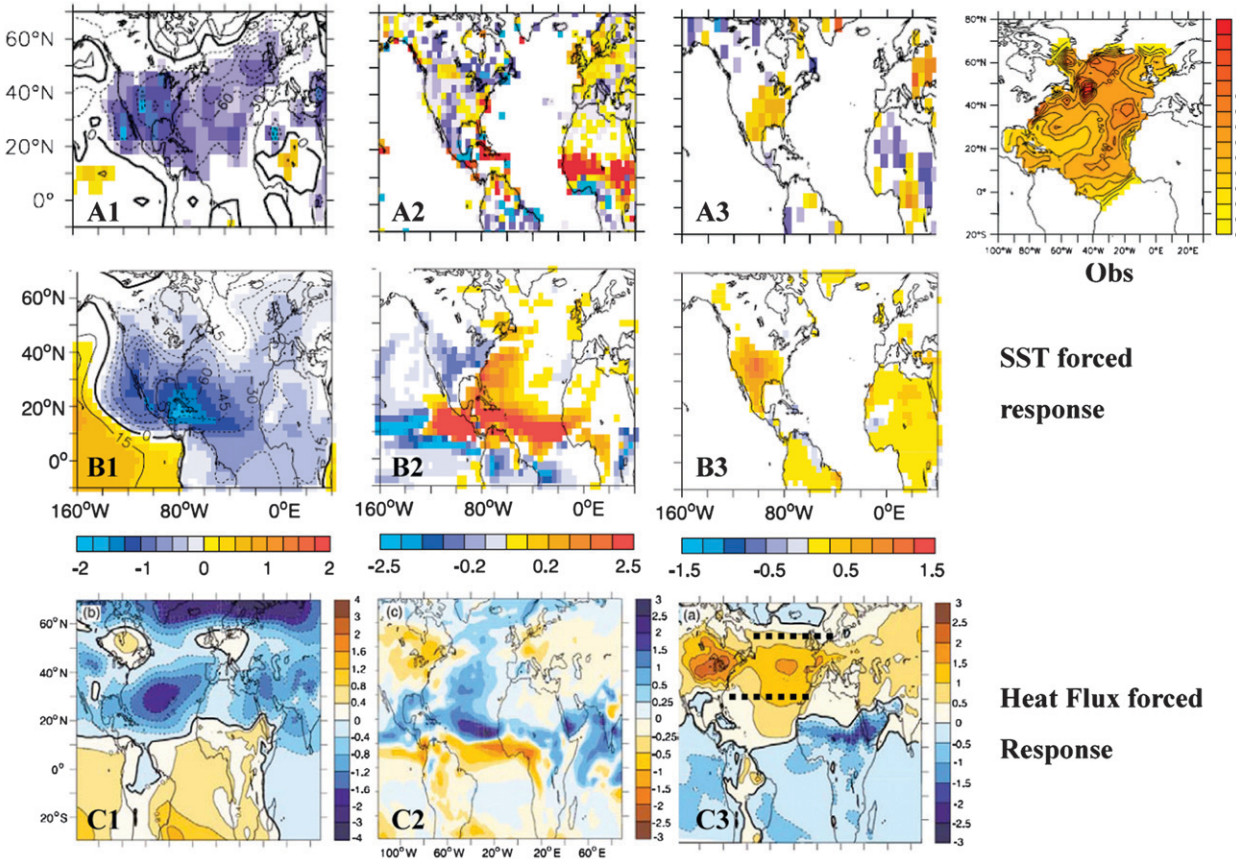

FIG. 17. Observed Jun–Aug atmospheric anomaly of (a1) SLP (contour interval = 30 Pa; color signal/noise ratio), (a2) precipitation, and (a3) surface air temperature between the warm (1930–60) and cold (1961–90) phase of the AMO. Simulated response of Aug–Oct (b1) SLP (contour interval = 15 Pa), (b2) precipitation, and (b3) surface air temperature to (top right) an idealized warm AMO SST anomaly (Obs) in the HADCM3 AGCM. Simulated atmospheric response of (c1) SLP (hPa), (c2) precipitation, and (c3) surface air temperature forced by a heat flux in the region of 30°–60°N in the GFDL2.1 AGCM coupled with a slab mixed layer (dotted lines). Precipitation in mm day^{-1} and surface temperature in K. [Adapted from Sutton and Hodson (2005) and Delworth et al. (2007).]

modest for monthly to seasonal responses, usually ~10 m °C^{-1} in the upper troposphere, which is about 10% of the total variability. This is much less than the impact of tropical SST variability on the atmosphere. However, the mean atmospheric response to a SST anomaly lasting a decade or more can still be significant. This follows because the variance of the random atmospheric internal variability decreases with the average time as a square root, but the variance of the atmospheric response signal to a decadal SST anomaly decreases more slowly. In coupled model experiments initiated with subsurface temperature anomalies, Liu et al. (2007) found that the signal/noise ratio increases from ~10% to ~50%, from the monthly to 4-yr mean responses. This example demonstrates that the atmospheric response to extratropical SST anomalies, although modest for monthly–seasonal responses, can be significant at decadal and longer time scales. This is consistent with coupled sensitivity experiments in which coherent decadal variability is identified in the extratropical atmosphere and ocean even though ocean–atmosphere interaction is only allowed in the extratropics (e.g., Fig. 12).

Recent observations using remotely sensed data of high spatial resolution suggest that the sharp SST front along the Gulf Stream generates a strong pressure gradient in the atmospheric boundary layer and, in turn, lower-layer convergence and, eventually, latent heat release in the atmospheric column (Minobe et al. 2008) (Fig. 19). The atmospheric heating tends to generate a baroclinic perturbation on the mean flow (Li and Conil 2003; Ferreira and Frankignoul 2005) and, in turn, eddies, which then feed back nonlinearly onto the mean flow to form an enhanced atmospheric response (Peng and Whitaker 1999).

Summary

Observational and modeling studies show some evidence of an atmospheric response to extratropical SST

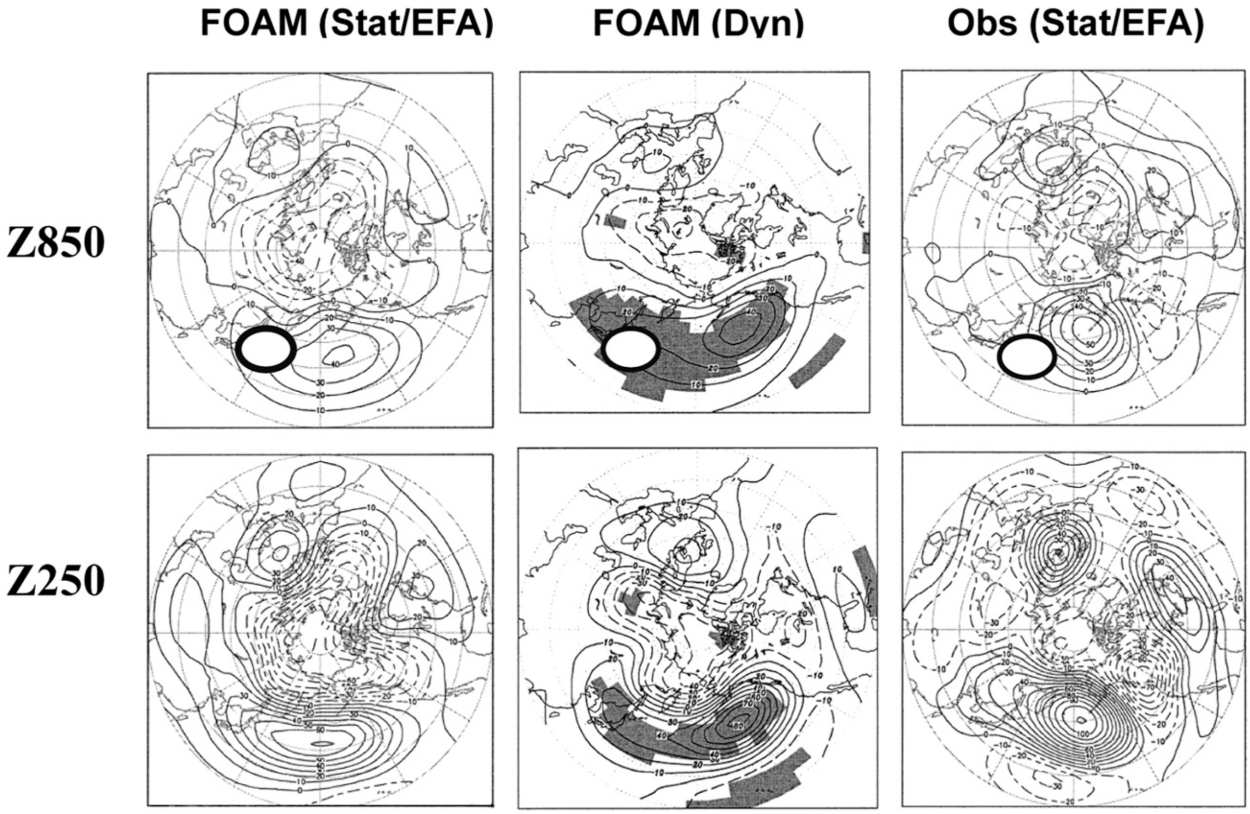

FIG. 18. Winter atmospheric response in geopotential heights at (top) 850 hPa and (bottom) 250 hPa to the Kuroshio Extension warm SST anomaly (marked by the black ellipses, top panels) assessed (left) statistically from the control run and (middle) dynamically in ensemble experiments, in the coupled model Fast Ocean Atmosphere Model (FOAM), and (right) statistically from the National Centers for Environmental Prediction (NCEP) reanalysis. The statistical estimation for both (left) the model and (right) NCEP is the same EFA method (Frankignoul et al. 1998) using the ratio of lag-1 covariance with the KOE SST anomaly. All geopotential height panels show an equivalent barotropic structure with a warm SST–ridge response. [See Liu and Wu (2004) for more details.]

(and sea ice) variability. The magnitude of the response appears to be modest at monthly to seasonal time scales, but significant at decadal and longer time scales. Therefore, the atmospheric response to extratropical SST forcing can be important for interdecadal climate variability. Even if the extratropical coupling is not critical for the genesis of coupled interdecadal modes, the atmospheric response to extratropical SST can still be of practical importance for the understanding of the climate impact of interdecadal variability in the atmosphere. It remains to be understood how low-level wind convergence affects atmospheric column heating and how the column heating affects the atmosphere through nonlinear wave–mean flow interaction. Although some model simulations and observations tend to show consistent results, such as the response to the Atlantic multidecadal oscillation, it remains to be understood why the simulated atmospheric responses to prescribed SST anomalies of a subbasin scale differ dramatically among models.

7. Conclusions and perspective

Our understanding of the mechanisms responsible for interdecadal variability has changed significantly over the last two decades. Earlier studies on interdecadal variability in the Atlantic focused on the role of thermohaline instability. In contrast, earlier studies on interdecadal variability in the Pacific were philosophically similar to studies on ENSO variability in applying a delayed oscillator paradigm with the emphasis on positive ocean–atmosphere feedback and self-exciting variability. Stochastic climate theory (Hasselmann 1976), which was proposed long ago for general climate variability without specific emphasis on preferred oscillation time scales, did not receive much attention in the study of interdecadal variability until the last decade or so. Now, the stochastic school has become dominant, and interdecadal variability is considered to be stochastically driven to leading order, with ocean–atmosphere feedback playing a minor role. This stochastic view is now

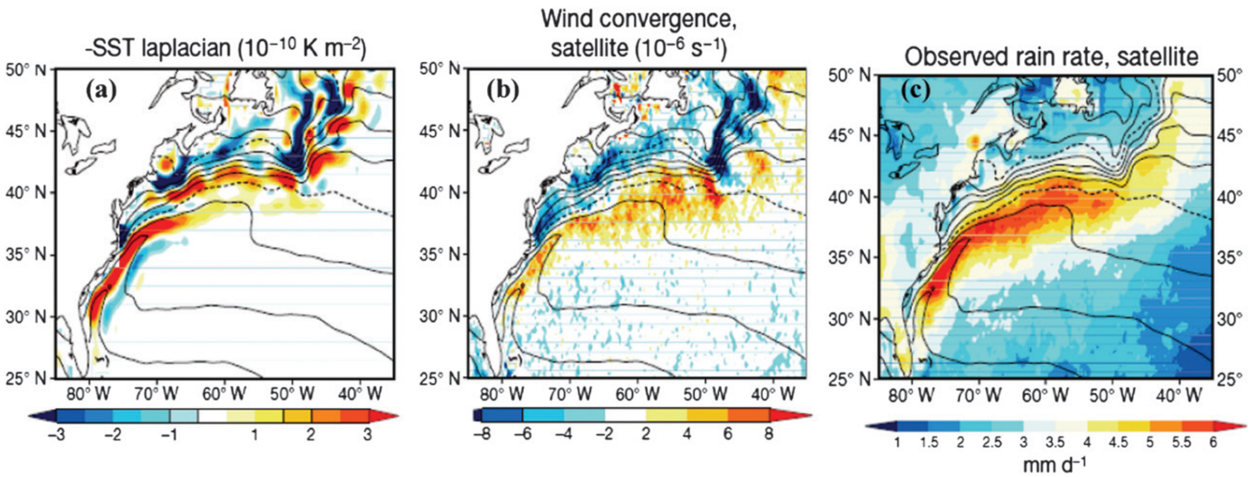

FIG. 19. Annual climatology of satellite-derived (a) sign-reversed SST Laplacian, (b) surface 10-m wind convergence (color), and (c) rain rate. (Adapted from Minobe et al. 2008.)

even reflected in ENSO, which can be predicted successfully either as a self-excited mode (Zebiak and Cane 1987) or as a stochastically driven damped mode (Penland and Magorian 1993; Penland and Sardeshmukh 1995; Kleeman and Power 1994).

Table 1 summarizes our understanding of the mechanisms for interdecadal variability in the Pacific and Atlantic. It also outlines some of the key issues that have not yet been resolved.

a. What we know

- Forcing and origin: stochastic forcing appears to be the major driving mechanism for all interdecadal variability, while ocean–atmosphere feedback appears to play a relatively minor role. Interdecadal variability can be generated independently in the tropics or extratropics and in the Pacific or Atlantic, although multidecadal variability appears more likely to be generated in the subpolar regions.
- Tropics versus extratropics: tropical SST is effective in forcing an atmospheric response, while extratropical oceanic memory is important for determining the long time scale of the variability.
- Time scale and ocean dynamics: ocean waves and ocean circulation are the leading candidates for the mechanism of time-scale selection. Interdecadal variability is associated with the wind-driven upper-ocean circulation in the Pacific, whereas in the Atlantic it is associated with changes in the AMOC, especially for the multidecadal variability. The time scale of interdecadal variability seems to be determined mainly by Rossby wave propagation in the extratropics; it could also be determined by the advection of the returning branch of the AMOC in the Atlantic.
- Ocean–atmosphere feedback: the atmospheric response to tropical oceanic variability is robust at all climate time scales. In comparison, the atmospheric response to extratropical oceanic variability is modest at interannual time scales, but could be significant at decadal and longer time scales and therefore contribute to interdecadal variability.

b. What we do not know

- Time-scale selection mechanism: what determines the time scale of a specific interdecadal variability, such as the Pacific decadal variability (PDV), Pacific multidecadal variability (PMV), Atlantic decadal variability (ADV), and Atlantic multidecadal variability (AMV)?
- Tropics versus extratropics: what are the roles of the tropics and extratropics in driving decadal and interdecadal variability?
- Thermohaline versus baroclinic instability: how is thermohaline instability or planetary wave baroclinic instability relevant to the interdecadal variability in CGCMs and the real world?
- Ocean–atmosphere coupling: what exactly is the role of ocean–atmosphere coupling for a specific interdecadal variability?

c. Further issues

Of all major unknowns for interdecadal variability, the most important is perhaps the mechanism for time-scale selection: what determines the time scales of specific interdecadal variability? In the case of ENSO, it is largely agreed that equatorial Rossby waves determine the interannual time scale of ENSO, although ocean–atmosphere coupling may modify the time scale. For interdecadal variability, in spite of our general sense that the preferred time scale is determined by the extratropical Rossby wave or thermohaline advection, it is far less clear what the specific mechanism is for the time scale of a specific variability mode. Even for the extratropical Rossby wave, the time scale varies greatly with latitude with the cross-basin time scale ranging from years to decades. It is therefore unclear, for example, what exactly determines the time scale of the Pacific decadal variability and what determines the time scale of the Pacific multidecadal variability? With the short observational record, it is even unclear if the interdecadal variability in observations is oscillatory with a preferred time scale or simply red noise variability. This poor understanding of the time scale of interdecadal variability is also reflected in current CGCM simulations, which produce interdecadal variability with a wide range of time scales. For example, across the Intergovernmental Panel on Climate Change (IPCC) climate models, the PDOs (defined as the first EOF of the annual SST in the North Pacific) all show a spatial pattern resembling the observed PDO. However, their time scales range widely from decades to multidecades (Overland and Wang 2007; Furtado et al. 2011). As another example, the GFDL model produces multidecadal variability with a similar pattern to its observational counterpart. The time scale, however, varied from 40 to 50 yr in the early version of model to ~20 yr in later versions. In the NCAR CCSM, the interdecadal variability in the Atlantic also exhibits different time scales at different model resolutions (Bryan et al. 2006; Danabasoglu 2008). A consistent spatial pattern is an indication that the atmospheric internal variability mode is largely correct in the model because the SST pattern is determined mainly by the atmospheric forcing. The disparity of time scales, however, suggests a high sensitivity of the time scale of interdecadal variability on ocean–atmosphere interaction and/or ocean dynamics. Given that climate models still produce ENSOs of various time scales, one should not be surprised that model interdecadal variability varies dramatically on the time scales exhibited. This poor understanding of the time scale of interdecadal variability poses a serious problem for our models and will have a significant impact on our ability to provide decadal climate predictions.

The nature of interdecadal climate variability in the Atlantic, and its relation to the oscillation studied in OGCMs with highly simplified surface boundary conditions, remain unclear. Is the AMOC variability more relevant to the thermohaline mechanism due to salinity transport and convective instability (under mixed boundary conditions), or large-scale baroclinic instability (under the flux–flux boundary condition)? Even though Atlantic multidecadal variability is ultimately generated by stochastic forcing, it is still important to understand how it is maintained against dissipation and what determines its spatial and temporal characteristics. All of this can be better understood if it can be simplified to an OGCM paradigm.

While most studies so far focused on the role of ocean, and ocean–atmosphere interaction, there are other mechanisms that may be important for interdecadal variability. For example, the interaction between the North Atlantic and Arctic Oceans can be important for interdecadal variability (Jungclaus et al. 2005), and sea ice interaction with the ocean may also be important for thermohaline variability (e.g., Yang and Neelin 1993). High-resolution ocean models are needed in CGCMs to explore the role of nonlinear dynamics of the midlatitude ocean in the generation of decadal variability, as suggested in simple model studies (e.g., Dewar 2001; Kravtsov et al. 2007). While most studies have focused on the generation mechanism of interdecadal variability within each basin, there are also studies suggesting the remote influence on interdecadal variability from other oceans. Observational analysis indicates significant correlation when the North Pacific variability leads the North Atlantic variability by 1 yr and when the North Atlantic variability leads the North Pacific by 11 yr (Wu et al. 2011), implying the potential of mutual interaction between the two coupled systems. This interaction can be accomplished by the atmospheric teleconnection, as indicated in modeling studies (Wu and Liu 2005). Latif et al. (2000) and Latif (2001) described some modeling evidence that AMOC variability is forced remotely by tropical Pacific climate variability. Zhang et al. (2007) were able to simulate a significant part of Pacific interdecadal variability in a coupled model forced by heat flux forcing only from the Atlantic, implying a potential impact from the Atlantic on Pacific interdecadal variability, consistent with the behavior evident in the CGCM analyzed by Rashid et al. (2010). Furthermore, factors outside the ocean–atmosphere system may also have impacts on interdecadal variability. For example, there is modeling evidence that the terrestrial ecosystem can enhance interdecadal ocean–atmosphere variability significantly over the tropical Atlantic (Zeng et al. 1999) and North Pacific (Notaro and Liu 2007).

d. What we need to do in the future

Given the limited observations available now and those expected in the near future, as well as the complex nature of interdecadal variability in the coupled system, it will remain a great challenge to understand and verify different mechanisms of interdecadal variability in the real world. Certainly, further observations, especially those on subsurface oceanic variability, are essential for improving our understanding of interdecadal variability. More advanced statistical methods have been developed to assess ocean–atmosphere interaction in the observations (Frankignoul et al. 1998, 2011; Newman 2009; Wen et al. 2010), which will help us clarify the role of ocean–atmosphere feedback in interdecadal variability. In the mean time, climate models are expected to play an increasingly important role. Current models, especially state-of-the-art CGCMs, have been improved such that they can successfully simulate interdecadal variability similar to observed variability. These models allow us to perform sensitivity experiments that can never be performed in the real world, and can therefore give us some information on the mechanisms responsible for interdecadal variability. With increased computing power, it also becomes practical for many research groups to use these CGCMs to perform sensitivity experiments on interdecadal variability. In addition, as more and more CGCMs become available, great insight can be gained with extensive model–model comparisons of interdecadal variability. Finally, these models are the most promising means for studying the predictability of interdecadal variability. This gives us a chance to combine the studies on mechanisms causing interdecadal variability with those on the predictability of interdecadal variability in the future.

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