Sea Surface Temperature Variability: Patterns and Mechanisms

Clara Deser,[1] Michael A. Alexander,[2] Shang-Ping Xie,[3] and Adam S. Phillips[1]

[1]Climate Analysis Section, Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado 80305; email: cdeser@ucar.edu, asphilli@ucar.edu

[2]Earth System Research Laboratory, National Oceanic and Atmospheric Association, Boulder, Colorado 80305; email: Michael.Alexander@noaa.gov

[3]International Pacific Research Center, School of Ocean and Earth Science and Technology, University of Hawaii, Honolulu, Hawaii 96822; email: xie@hawaii.edu

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Abstract

Patterns of sea surface temperature (SST) variability on interannual and longer timescales result from a combination of atmospheric and oceanic processes. These SST anomaly patterns may be due to intrinsic modes of atmospheric circulation variability that imprint themselves upon the SST field mainly via surface energy fluxes. Examples include SST fluctuations in the Southern Ocean associated with the Southern Annular Mode, a tripolar pattern of SST anomalies in the North Atlantic associated with the North Atlantic Oscillation, and a pan-Pacific mode known as the Pacific Decadal Oscillation (with additional contributions from oceanic processes). They may also result from coupled ocean-atmosphere interactions, such as the El Niño-Southern Oscillation phenomenon in the tropical Indo-Pacific, the tropical Atlantic Niño, and the cross-equatorial meridional modes in the tropical Pacific and Atlantic. Finally, patterns of SST variability may arise from intrinsic oceanic modes, notably the Atlantic Multidecadal Oscillation.

1. INTRODUCTION

The oceans play an important role in the climate system owing in part to their large heat-storage capacity: Approximately 3.5 m of water contains as much energy as an entire atmospheric column. The oceans' thermal inertia is communicated to the atmosphere via turbulent and radiative energy exchange at the sea surface. These energy fluxes in turn depend on a single oceanic quantity, the sea surface temperature (SST), as well as several atmospheric parameters including wind speed, air temperature, humidity, and cloudiness. SSTs thus play a key role in regulating climate and its variability. In particular, slow variations in SST provide a source of potential predictability for climate fluctuations on timescales of seasons and longer. In this review, we describe the dominant spatial and temporal patterns of nonseasonal SST variability observed over the past ~150 years and discuss the current state of knowledge regarding their underlying physical mechanisms and climate impacts.

The rest of the paper is outlined as follows: Section 2 provides some background on the physical processes governing SST variability, including the role of atmospheric circulation forcing. The North Atlantic SST tripole mode is used to illustrate the latter. Section 3 introduces the SST data sets commonly used for climate studies, with an emphasis on issues related to spatial and temporal coverage. Section 4 discusses the geographical distribution of nonseasonal SST variability and provides a brief introduction to Empirical Orthogonal Function analysis, a commonly used technique for identifying patterns of variability. Section 5 presents the primary modes of nonseasonal SST variability in each ocean basin. These include the following (in addition to the North Atlantic SST tripole): (*a*) the El Niño-Southern Oscillation (ENSO) phenomenon in the tropical Pacific, with teleconnections to the other ocean basins; (*b*) the tropical Indian Ocean dipole mode; (*c*) the tropical Atlantic Niño and meridional modes; (*d*) Southern Ocean variability; (*e*) the Pacific Decadal Oscillation; and (*f*) the Atlantic Multidecadal Oscillation. Summary Points and Future Issues sections conclude this article.

2. PHYSICAL BACKGROUND

2.1. Heat Budget of the Upper-Ocean Mixed Layer

SSTs are governed by both atmospheric and oceanic processes. On the atmospheric side, wind speed, air temperature, cloudiness, and humidity are the dominant factors regulating the exchange of energy at the sea surface. On the oceanic side, heat transport by currents, vertical mixing, and boundary layer depth influence SST. Mathematically, the heat budget for the upper-ocean mixed layer may be written as

$$\partial T/\partial t = Q_{net}/(\rho Cp H) + (\vec{U}_{geo} + \vec{U}_{ek}) \cdot \vec{\nabla} T + (W_e + W_{ek})(T - T_b)/H, \tag{1}$$

where ρ is the density of seawater, Cp is the specific heat of seawater, H is the mixed layer depth, T is the mixed layer temperature (equal to the SST), Q_{net} is the net surface energy flux, U_{geo} is the geostrophic current velocity, U_{ek} is the Ekman current velocity, W_e is the vertical entrainment rate, W_{ek} is the Ekman pumping velocity, and T_b is the temperature of the water at depth that is entrained into the mixed layer. Q_{net} is defined as

$$Q_{net} = Q_{sh} + Q_{lh} + Q_{sw} + Q_{lw}, \tag{2}$$

where Q_{sh} is the sensible heat flux, Q_{lh} is the latent heat flux, Q_{sw} is the downward solar radiative flux minus the portion that penetrates through the mixed layer, and Q_{lw} is the longwave radiative flux. The turbulent ($Q_{sh} + Q_{lh}$) energy flux is linearly proportional to the wind speed and the air-sea temperature or humidity difference; the radiative fluxes ($Q_{sw} + Q_{lw}$) are functions of air temperature, humidity, and cloudiness. The bulk formulas for the air-sea fluxes may be found in standard texts such as Peixoto & Oort (1992), Pond & Pickard (1983), and Hartmann (1994). Ekman and geostrophic currents contribute to the heat budget of the mixed layer through horizontal advection, whereas entrainment and Ekman pumping alter the SST through vertical advection. A complete discussion of these terms may be found in standard oceanography texts such as Pond & Pickard (1983), Vallis (2006), and Stewart (2005).

2.2. Atmospheric Circulation Forcing

Much of the large-scale organization of SST anomalies results from the large-scale organization of atmospheric circulation anomalies and attendant changes in the turbulent and radiative energy fluxes at the air-sea interface and the local wind-driven Ekman currents (Cayan 1992, Alexander & Scott 1997, Deser & Timlin 1997, Marshall et al. 2001, Visbeck et al. 2003, Alexander & Scott 2008). In the middle latitudes, eastward-moving storms dominate surface energy flux variations on synoptic timescales (3 to 10 days), whereas blocking events, slowly moving cutoff lows, Rossby waves, and semistationary patterns contribute to surface energy flux variations on timescales longer than approximately 10 days (Alexander & Scott 1997). In general, atmospheric variability with timescales longer than approximately 10 days is more effective at driving SST anomalies than is synoptic atmospheric variability, because of the thermal inertia of the upper-ocean mixed layer (Frankignoul & Hasselmann 1977, Deser et al. 2003). The frequency dependency of the thermal response of the mixed layer to atmospheric forcing is discussed further below.

Extra-tropical atmospheric circulation variability on timescales longer than approximately 10 days tends to be organized into recurring large-scale patterns anchored to particular geographical regions. These large-scale patterns often consist of anomalies of opposite sign over distant parts of the globe, commonly referred to as teleconnections in the meteorological literature (Wallace & Gutzler 1981, Barnston & Livezey 1987, Kushnir & Wallace 1989, Trenberth et al. 1998). Prominent examples of large-scale teleconnection patterns of extratropical atmospheric circulation variability include the North Atlantic Oscillation (NAO) and the related Northern Annular Mode (NAM), the Pacific-North American and Pacific-South American patterns, and the Southern Annular Mode (SAM). Nigam (2003) reviewed these and other atmospheric circulation teleconnection patterns.

The NAO refers to a redistribution of atmospheric mass, manifest as sea-level pressure (SLP) changes of opposite sign, between the Arctic and subtropics of the Atlantic sector (e.g., **Figure 1**). This SLP oscillation (in space, not time: Walker & Bliss 1932) is accompanied by changes in wind, storminess, air temperature, and precipitation across the North Atlantic as well as over eastern North America and Europe (for a recent review, see Hurrell & Deser 2009). The related NAM incorporates the spatial pattern of the NAO with a third center of SLP variability over the North Pacific that fluctuates in phase with that over the North Atlantic. The NAM's Southern Hemisphere counterpart, the SAM, also refers to a north-south redistribution of atmospheric mass between high (poleward of 50°S) and middle (30°–50°S) latitudes, with accompanying changes in wind, air temperature, and precipitation (Marshall 2003). The SAM is a zonally symmetric pattern of atmospheric circulation variability that spans all longitudes, whereas the NAM exhibits more zonal variation due to continental effects (Thompson & Wallace 2000). Finally, the Pacific-North American and Pacific-South American are wave-like teleconnection patterns consisting of SLP anomaly centers that approximately follow a great circle route over the Pacific and American sectors of their respective hemispheres (Wallace & Gutzler 1981, Karoly 1989). Most prominent during winter, these patterns affect wind, air temperature, and precipitation over the North and South Pacific.

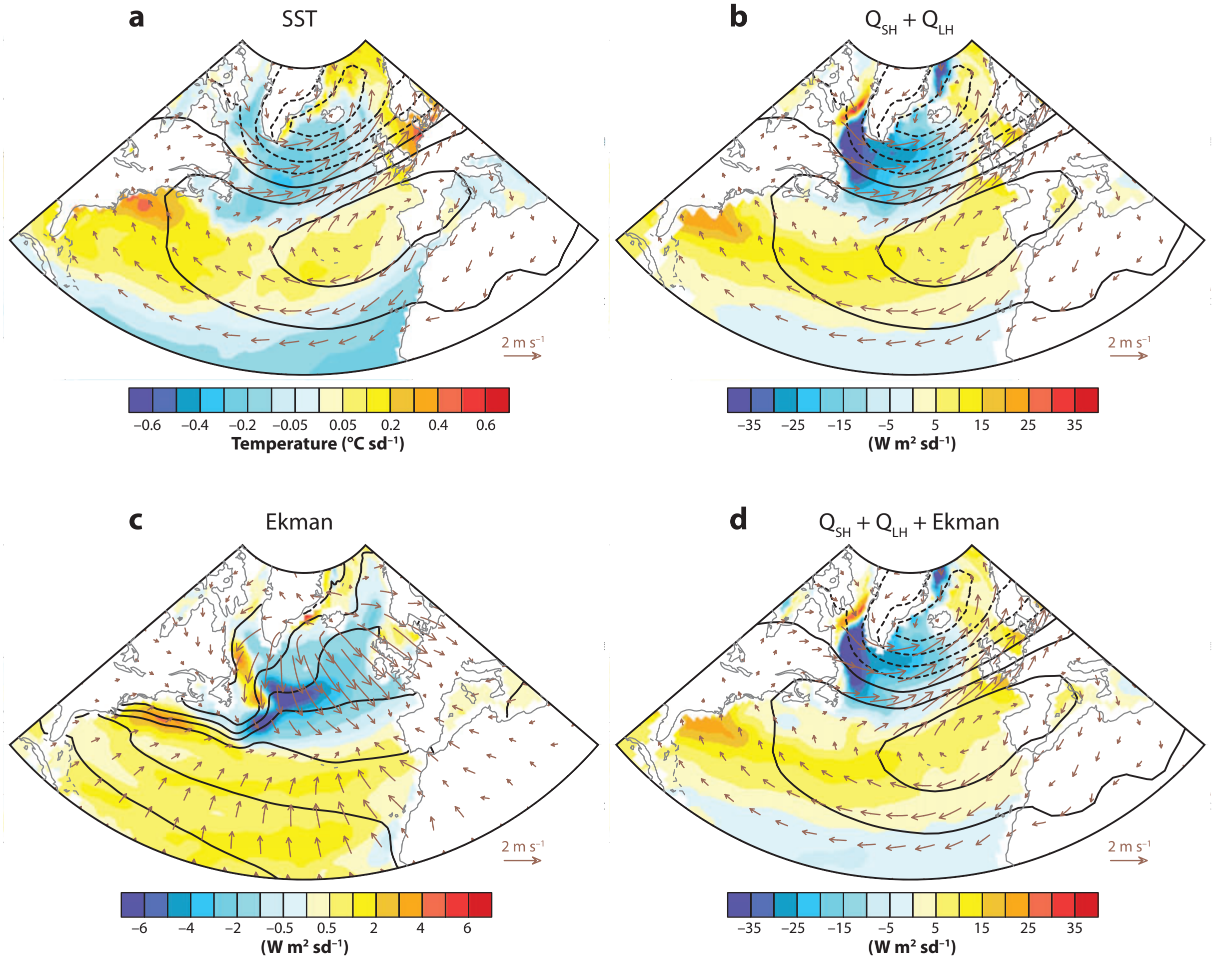

Figure 1

Anomaly patterns associated with a +1 standard deviation departure of the North Atlantic Oscillation (NAO) Index during winter (December–March) defined using the station index of Hurrell et al. (2003). (*a*) Sea surface temperature (SST) (*shading*), sea-level pressure (SLP) (*contours*), and surface wind (*arrows*). (*b*) Sensible plus latent energy flux ($Q_{sh} + Q_{lh}$) (*shading*), SLP (*contours*), and surface wind (*arrows*). (*c*) Ekman heat transport expressed as an equivalent surface energy flux (*shading*), long-term mean SST (*contours*), and Ekman currents (*arrows*). (*d*) Sum of the sensible, latent, and Ekman energy fluxes (*shading*), SLP (*contours*), and surface wind (*arrows*). The SLP contour interval is 1 hPa, with negative values dashed.

The spatial structures of these large-scale patterns of extratropical atmospheric circulation variability are driven primarily by internal nonlinear dynamical processes. That is, they are intrinsic to the atmosphere and require no external forcing to exist. The temporal evolution of these atmospheric circulation patterns is generally consistent with a stochastic first-order autoregressive process with an e-folding timescale of approximately 10 days (Feldstein 2000; N.C. Johnson & S.B. Feldstein, unpublished manuscript). That is, these patterns tend to follow a random sequence akin to white noise on timescales longer than approximately 10 days. Although these teleconnection patterns are intrinsic to the atmosphere, they may also be excited by external forcing factors such as changes in SSTs, sea ice, atmospheric chemical composition (for example, volcanic aerosols as well as ozone and greenhouse gas concentrations), and solar output, in analogy with the excitation of the "normal modes" of a drum. These external forcings may thus alter the temporal evolution of atmospheric teleconnection patterns, enhancing their persistence and potential predictability. External forcing may also generate new patterns of atmospheric circulation response.

In contrast to the extratropics, large-scale patterns of atmospheric circulation variability in the tropics result primarily from interaction with the ocean: That is, they do not exist in the absence of SST variability. The most prominent large-scale tropical atmospheric pattern of variability is the Southern Oscillation, an east-west redistribution of atmospheric mass between the Indian Ocean/western Pacific and the eastern Pacific. In the positive (negative) phase of the Southern Oscillation, the SLP gradient across the tropical Pacific is stronger (weaker) than normal, accompanied by enhanced (diminished) trade winds and equatorial easterly winds. This pattern owes its existence to tropical ocean-atmosphere interaction associated with the ENSO phenomenon and is thus an intrinsically coupled phenomenon. The roles of intrinsic and ocean-coupled atmospheric variability are discussed further in connection with specific patterns of SST variability.

The large-scale nature of atmospheric teleconnection patterns is imprinted upon the SST anomaly field via the surface energy fluxes and Ekman currents. This is illustrated for the NAO in **Figure 1**. The SST field exhibits a tripolar structure marked by a cold anomaly in the subpolar North Atlantic, a warm anomaly in the middle latitudes, and a cold subtropical anomaly between the equator and 30°N; warm anomalies are also found in the Greenland and North Seas. Maximum SST anomaly amplitudes are on the order of 0.5°C. A similar tripolar pattern is seen in the turbulent energy fluxes (**Figure 1*b***), where the sign is defined as positive downwards (e.g., from the atmosphere to the ocean). The ocean loses energy to the atmosphere over the subpolar and tropical Atlantic as a result of strengthened westerly winds and northeast trade winds, respectively. It gains energy in the middle latitudes as a result of the diminished wind speeds (easterly anomalies superimposed on a background of mean westerlies) and anomalous warm-air advection associated with southerly wind anomalies along the eastern seaboard of the United States. Thus, the large-scale structure of the SST anomaly tripole is driven by the turbulent energy flux anomalies associated with the NAO (e.g., Cayan 1992, Marshall et al. 2001, Visbeck et al. 2003). Whereas the mixed-layer heat-budget equation relates the fluxes to the time tendency of SST and not the SST per se (recall Equation 1), the winter-mean SST anomaly closely resembles the SST anomaly tendency across the winter season and the SST anomalies lag the atmospheric forcing by 2–3 weeks (e.g., Deser & Timlin 1997).

Although the surface turbulent energy fluxes drive the overall tripolar pattern of SST anomalies associated with the NAO, horizontal temperature advection by the anomalous Ekman currents, expressed as an equivalent surface heat flux anomaly, also contributes in regions of strong mean SST gradients such as the Gulf Stream and North Atlantic Current (**Figure 1*c***). Other factors such as the radiative fluxes at lower latitudes and spatial variations in mixed-layer depth also play a role in determining the magnitude of the SST anomalies. For example, the winter mixed layers are substantially deeper in the subpolar North Atlantic compared with the tropical Atlantic ($\sim$300 m versus $\sim$60 m), which reduces the amplitude of the SST response (recall Equation 1). This effect is evidenced by the fact that the NAO-related SST anomalies are of comparable magnitude between high and low latitudes, whereas the heat flux anomalies differ by approximately a factor of 5 (**Figure 1**). Although the SST anomalies associated with the North Atlantic tripole are primarily forced by the atmosphere, there is some evidence for a weak feedback on the NAO (Watanabe & Kimoto 2000, Czaja & Frankignoul 2002).

An important concept for understanding the nature of SST variability is the stochastic climate model paradigm (Frankignoul & Hasselmann 1977). According to this paradigm, ocean mixed-layer temperature anomalies are forced by random atmospheric variability (e.g., atmospheric circulation variability that decorrelates within a week or two) via surface energy fluxes and Ekman currents, and decay by damping back to the atmosphere via turbulent energy and longwave radiative fluxes (modeled as a negative linear feedback term). Put another way, the ocean mixed layer integrates the "white noise" atmospheric forcing to yield a "red noise" SST response. In this view, the predictability or persistence of SST anomalies is limited to the timescale associated with the thermal inertia of the mixed layer, a timescale determined by the depth of the mixed layer and the rate at which the SST anomaly damps to the atmosphere via turbulent heat fluxes. The simple stochastic climate model has been widely adopted as the leading paradigm for the "null hypothesis" of SST variability in middle and high latitudes where random atmospheric forcing is a good approximation.

An illustration of the simple stochastic climate model paradigm is given in **Figure 2**, which shows the mixed-layer temperature response to random variations in surface heat flux forcing over a 60-year period. Two cases are considered: a shallow (50 m) and a deep (500 m) mixed layer. These values span the typical range of observed mixed-layer depths for which 50 m is representative of annual mean conditions in the North Pacific and 500 m typical of the northern North Atlantic during winter. Slow variations in the mixed-layer temperature are apparent for both cases, with fluctuations on the order of a few years for the shallow case and on the order of decades for the

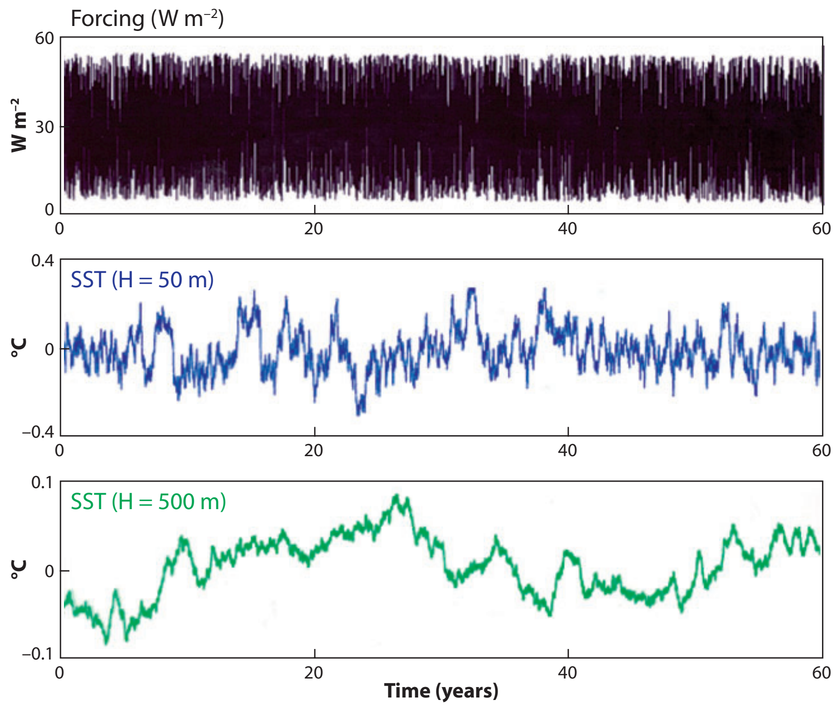

Figure 2

Illustration of the stochastic climate model paradigm showing a random (e.g., "white noise") atmospheric heat flux forcing time series (*top graph*) and the upper-ocean mixed-layer temperature response for a mixed-layer depth of 75 m (*middle graph*) and 500 m (*bottom graph*). Note the slow fluctuations in the ocean mixed-layer temperature response.

deep mixed-layer case. This simple example illustrates that slow SST variations may be induced by random (e.g., unpredictable) atmospheric forcing simply due to the thermal inertia associated with a deep mixed layer. Note that the stochastic climate model does not generate spectral peaks; rather, it enhances the low-frequency component of variability according to a first-order autoregressive process. The stochastic climate model paradigm contributes to SST variability in the North Pacific associated with the Pacific Decadal Oscillation (Newman et al. 2003) and to the long (decadal timescale) persistence of SST fluctuations in the North Atlantic (Deser et al. 2003, DeCoetlogon & Frankignoul 2003). This powerful null hypothesis of SST variability makes it difficult to separate the contributions of oceanic and atmospheric forcing solely on a timescale basis.

In addition to atmospheric processes, oceanic processes such as upwelling, entrainment, and lateral advection also contribute to SST variability. For example, vertical advection plays a prominent role along the coastal and equatorial upwelling zones, and horizontal advection is important along the western-boundary current regions (e.g., the Gulf Stream and Kuroshio Current). Oceanic processes also play an indirect role in SST variability by affecting the depth of the upper-ocean mixed layer. Even the mean seasonal cycle of mixed-layer depth can induce winter-to-winter memory or persistence of SST anomalies via the "re-emergence" mechanism whereby thermal anomalies stored in the deep winter mixed layer persist at depth through summer and become partially reentrained into the mixed layer during the following winter (Namias & Born 1974, Alexander & Deser 1995, Deser et al. 2003). Several of these processes have been incorporated to extend the framework of the stochastic climate model, including the re-emergence mechanism (Deser et al. 2003) and horizontal advection (Frankignoul & Reynolds 1983, DeCoetlogon & Frankignoul 2003).

3. SEA SURFACE TEMPERATURE DATA SETS

The patterns of variability sampled in any particular SST data set will depend to some extent on the spatial and temporal resolution of the data and the length of record. For example, satellite-based SST archives of high spatial resolution but short duration may reveal patterns with enhanced meso-scale definition but lack information on multidecadal variability. Conversely, historical ship-based archives of coarse and uneven spatial sampling but long duration may reveal patterns with timescales of multiple decades and longer but lack detailed spatial information. Here we provide a brief overview of the most commonly used SST data sets in climate studies, including information on their spatial resolution and temporal coverage as well as any filling procedures used to account for missing data. Our discussion is limited to those data sets that are available on a regular grid over the world oceans (**Table 1**).

The International Comprehensive Atmosphere-Ocean Data Set (ICOADS) is an extensive and widely used digital collection of quality-controlled surface weather observations (including SST). The majority of the measurements come from ships of opportunity, supplemented in recent years by research vessels, moored environmental buoys, drifting buoys, and near-surface measurements from hydrographic profiles. The data are monthly averaged and binned into 2° latitude by 2° longitude boxes beginning in 1800 (1° by 1° beginning in 1960) and extending through 2007 with updates every few years.

Due to the uneven distribution of commercial shipping routes and changes in those routes over time, data coverage is poor in certain regions and periods (**Figure 3**). Broadly speaking, the North Atlantic, western South Atlantic, and northern Indian oceans contain the highest density of observations, with reasonable coverage back to approximately 1870. Data coverage is limited

Table 1 Commonly used SST data sets for climate studies

Name and Web address	Period of record	Coverage	Comments
ICOADS http://icoads.noaa.gov/	1800–present	Monthly	In situ data
		2° × 2° since 1800 1° × 1° since 1960	No processing beyond quality control
HadISST	1870–present	Monthly	In situ/satellite data
http://badc.nerc.ac.uk/data/hadisst/		1° × 1°	Missing values filled
NCDC	1854–present	Monthly	In situ/satellite data
http://www.ncdc.noaa.gov/oa/climate/research/sst/ersstv3.php		2° × 2°	Missing values filled
NOAA Optimum Interpolation SST Analysis	Nov. 1981–present	Weekly/monthly	Satellite estimates
http://www.emc.ncep.noaa.gov/research/cmb/sst_analysis/		1° × 1°	
NOAA Optimum Interpolation 1/4 Degree Daily SST Analysis http://www.ncdc.noaa.gov/oa/climate/research/sst/oi-daily.php	Sept. 1981–present	Daily 0.25° × 0.25°	Satellite estimates
Kaplan Extended v2 SST Anomaly Data	1856–present	Monthly	In situ/satellite data
http://ingrid.ldeo.columbia.edu/SOURCES/.KAPLAN/.EXTENDED/.v2/.ssta/		5° × 5°	Missing values filled

in the North Pacific before around 1946 and in the tropics before around 1960; the Southern Ocean remains poorly sampled throughout the record. The uneven and changing spatial coverage of SST measurements from historical ship-based archives must be taken into account in any analyses.

Individual SST measurements in the ICOADS are subjected to quality-control procedures to remove outliers and duplicates. However, the screened values are not corrected for changes in instrumentation, observing practice, ship type, etc.; missing grid boxes are not filled in; and no "analysis" of the data is performed (e.g., no spatial or temporal smoothing or interpolation). An example of errors introduced by changes in instrumentation is the spurious warming trend around 1941 as a result of switching from bucket to engine-intake measurements (Thompson et al. 2008). The lack of spatial and temporal smoothing in the ICOADS, along with large uncertainties in individual monthly mean values due to inadequate sampling, makes it difficult to produce comprehensible maps of a particular climatic variable for a specific month and year without additional processing of the data.

Various empirical and statistically optimal procedures have been employed to improve upon the sampling uncertainties, temporal inhomogeneities, and missing data in the ICOADS and related archives. These globally complete monthly SST analysis products (e.g., HadISST, Kaplan, and NCDC in **Table 1**) are tremendously useful for certain applications, for example, in studies of global SST variability and as lower-boundary conditions in atmospheric model simulations, although they remain constrained by the quality, quantity, and distribution of the original measurements (c.f., Hurrell & Trenberth 1999). The HadISST data are reconstructed using a two-stage reduced-space optimal interpolation procedure, followed by superposition of quality-improved gridded observations onto the reconstructions to restore local detail (Rayner et al. 2003). Similar methods are used for Kaplan and NCDC (see Web addresses in **Table 1** for detailed information). In this review, we make use of the HadISST data set in part because of the higher spatial resolution

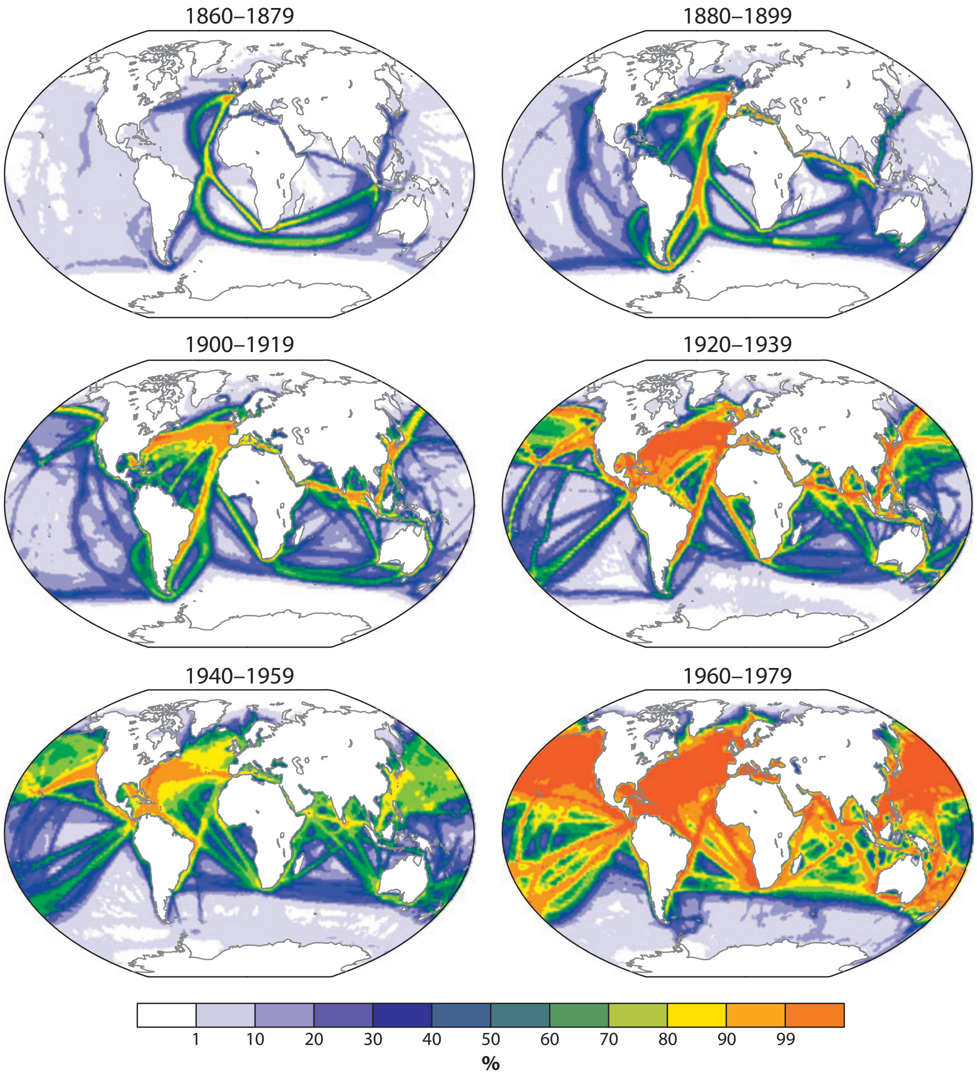

Figure 3

Distribution of sea surface temperature observations from the International Comprehensive Ocean Atmosphere Data Set for each 20-year period since 1860. Color shading indicates the percentage of months with at least one measurement in a 2° latitude by 2° longitude grid box.

(1° latitude by 1° longitude) compared with the Kaplan (5° latitude by 5° longitude) and NCDC (2° latitude by 2° longitude) archives.

With the advent of satellite retrievals of SST beginning in the early 1980s, there is now full SST coverage over the world oceans at high temporal (every few days) and spatial (~10–25 km) resolution. Various satellite retrieval methods are available, including infrared sensors from the Advanced Very High Resolution Radiometer and microwave measurements from the Advanced Microwave Scanning Radiometer. The satellite data have also been blended with conventional in situ data to account for biases and inhomogenieties associated with changing satellites and orbits through time (NOAA Optimum Interpolation products; see **Table 1**).

4. GEOGRAPHICAL PATTERNS OF NONSEASONAL SEA SURFACE TEMPERATURE VARIABILITY

4.1. Standard Deviations

A common approach to quantifying the magnitude and spatial distribution of nonseasonal SST variability is to map the standard deviation (σ) of the monthly SST anomalies (defined as the departure of the SST for a particular month from the long-term monthly mean) at each location. **Figure 4*b*** shows the σ distribution based on satellite-derived SST values during 1982–2008. The long-term mean SST distribution is also shown for reference (**Figure 4*a***). Three regions stand out as having the largest nonseasonal variability: the upwelling zones along the equatorial Pacific and coastal Peru, the western-boundary current regions of the Kuroshio and Gulf Stream, and the Southern Ocean from the tip of South Africa eastward into the Indian Ocean. Maximum values of the SST anomaly standard deviations in these regions exceed 1.5°C. In the extratropics, the regions of maximum SST anomaly standard deviation coincide with regions of maximum north-south mean SST gradients (termed frontal zones). In the tropics, the highest standard deviations occur where there is a local minimum in the long-term mean SSTs due to equatorial and coastal upwelling. Apart from the regions of maximum SST anomaly standard deviation, there is relatively high variability (standard deviations around 0.75°C–1°C) across the North Pacific and North Atlantic and the South Pacific, South Indian, and South Atlantic and relatively low variability across the tropical ocean basins (standard deviations around 0.3°C–0.5°C). A similar map based on the COADS archive lacks the sharpness of the SST variance maxima seen in the satellite data owing to the coarser sampling in both space and time.

4.2. Methods for Identifying Patterns of Variability

The standard deviation map of nonseasonal SST anomalies presented in **Figure 4** does not give any information on how SST variations at one location are related to those at another. To ascertain the covariability of SST anomalies at different locations, alternative analysis methods are needed. One commonly used approach is empirical orthogonal function (EOF) analysis and extensions thereof (von Storch & Zwiers 1999). Conceptually, EOF analysis determines a spatio-temporal pattern of variability that accounts for the maximum covariance between the SST anomaly time series at all pairs of grid points in the data set. Then, the remaining covariability is subject to the same decomposition with the added constraint that the second spatio-temporal EOF pattern is orthogonal (e.g., uncorrelated) in both time and space to the leading EOF pattern. This procedure is repeated until all n EOF patterns have been computed, where n is equal to the number of grid points. In practice, only the first few leading modes are robust as a result of the orthogonality constraint. EOF rotation may be used to circumvent this constraint; however, it may also result in patterns that are overly localized in space. Each EOF pattern is associated with a principal component (PC) time series, which describes the temporal evolution of the EOF pattern. The PC time series may be obtained by projecting the EOF pattern onto the original SST anomaly field at each time step so as to find the sign and amplitude of the pattern at any given time. It should be noted that the sign of the EOF is arbitrary; however, the product of the EOF and the PC time series recovers the correct polarity of the mode at any given grid box and time.

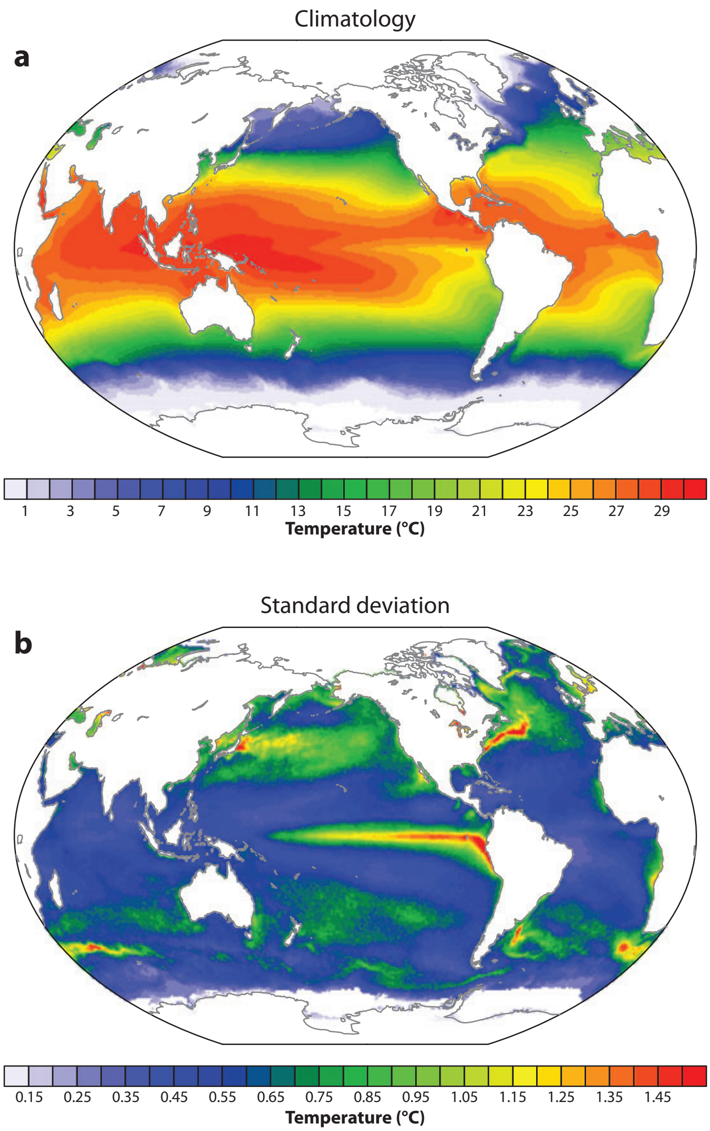

Figure 4

(*a*) Long-term mean sea surface temperature (SST) distribution from satellite passive microwave measurements during 1982–2008. (*b*) The standard deviation of monthly SST anomalies (deviations from the long-term monthly means) based on the same data set.

Although useful, EOF analysis is not foolproof. The EOF modes depend on the spatial domain considered, are subject to orthogonality constraints, and may not be separable if they account for similar percentages of the total variance (North et al. 1982). Also, EOFs are empirically determined "modes" and thus are not necessarily equivalent to the dynamical modes of the system. It is always prudent to confirm EOF patterns with simpler techniques such as compositing, one-point correlation maps, and linear regression analysis. Some examples are given in the next section.

5. DOMINANT PATTERNS OF NONSEASONAL SEA SURFACE TEMPERATURE VARIABILITY

We first describe the dominant patterns of SST variability in the tropics, followed by those in the extra tropics.

5.1. Tropics

In this section, we consider ENSO in the Pacific, the basin-wide and zonal dipole modes in the Indian Ocean, and the Niño and meridional dipole modes in the Atlantic.

5.1.1. ENSO. The leading EOF of monthly SST anomalies over the globe is the ENSO mode. Here we show the leading EOF based on linearly detrended monthly anomalies during the period 1900–2008 from the HadISST data set (**Figure 5*a***), but similar results are found for shorter time periods and other SST data sets. This single mode accounts for 19% of the nonseasonal variability of SSTs over the global oceans during the past 109 years. The spatial pattern associated with the warm phase of ENSO consists of positive SST anomalies across the eastern two-thirds of the equatorial Pacific Ocean, flanked by weaker negative anomalies over the far western tropical Pacific and extending in a horseshoe-like fashion to the North and South Pacific. Positive SST anomalies are also found along the west coasts of North and South America. ENSO-related SST anomalies also occur over the Atlantic, Indian, and Southern oceans.

A commonly used index for representing SST variability associated with ENSO is the area average of monthly SST anomalies in the region 5°N–5°S, 170°–120°W (outlined by the box in **Figure 5*a***), referred to as the Niño-3.4 SST index for historical reasons having to do with the configuration of ship tracks in the region (Rasmusson & Carpenter 1982). The Niño-3.4 SST index is nearly identical to the PC time series associated with the leading EOF: Their correlation coefficient is 0.93. The Niño-3.4 time series exhibits an irregular series of warm and cold "events" typically lasting about 1–1.5 years and recurring approximately every 3–8 years (**Figure 5*b***). The largest warm events such as those that occurred in 1982–1983 and 1997–1998 have amplitudes of approximately 2°–2.5°C, whereas more moderate warm events range from 1°C to 2°C. Cold events tend to be somewhat weaker and longer lasting than their warm counterparts. A power spectrum of the Niño-3.4 time series shows that the dominant range of periods is 2.5–8 years, with some sensitivity to the period of record sampled (**Figure 5*c***). ENSO is seasonally dependent as indicated by the seasonal cycle of the standard deviation of the Niño-3.4 SST anomaly time series (**Figure 5*d***), which shows minimum values (~0.55°C) during April–June and maximum values (~0.95°C) near the end of the calendar year (November–January).

ENSO-related SST anomalies evolve substantially during the course of a warm or cold event. A composite ENSO evolution over a period of six seasons is shown in **Figure 6**, starting with conditions during March–May before the peak and ending with conditions in June–August of the following year. The warm-event composite is based on a selection of the years for which the Niño-3.4 SST index in September–November exceeds 1 standard deviation based on the period 1950–2008. Similarly, the cold-event composite is based on a selection of the years for which the September–November Niño-3.4 SST index is less than -1 standard deviation. The cold-event composite is then subtracted from the warm-event composite. The composite ENSO evolution is depicted for linearly detrended SST and SLP anomalies in **Figure 6**. Note that this approach is independent of the EOF analysis shown earlier.

The ENSO composite SST evolution exhibits an incipient warming along the equator and coast of South America during MAM0 that subsequently grows in amplitude over the next several

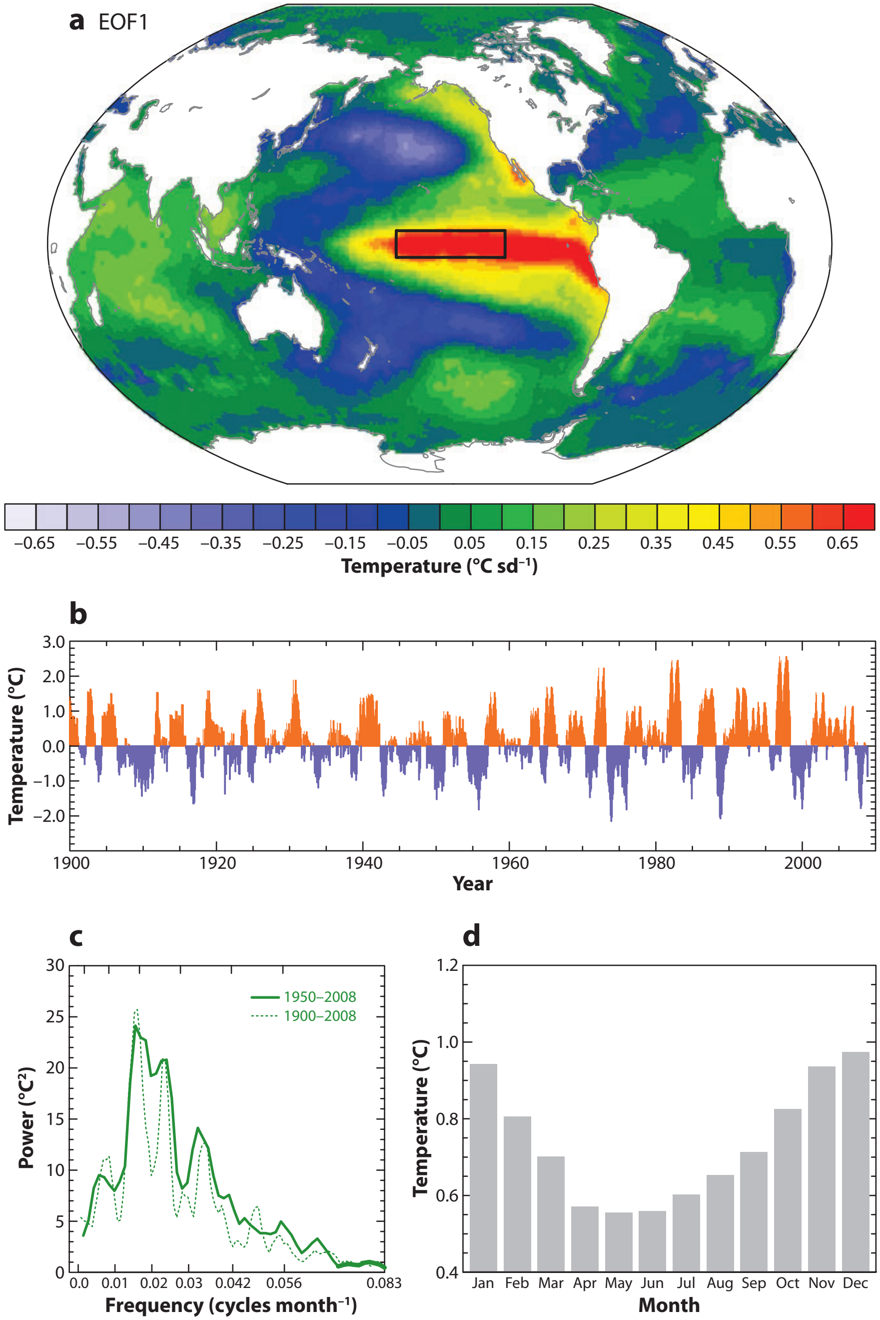

Figure 5

(*a*) Leading empirical orthogonal function (EOF) of detrended monthly sea surface temperature (SST) anomalies over the global oceans based on the HadISST data set during 1900–2008. This mode, which accounts for 19% of the variance, depicts the El Niño-Southern Oscillation phenomenon. (*b*) Monthly SST anomaly time series in the Niño-3.4 region (5°N–5°S, 170°–120°W: outlined by the rectangle on the EOF pattern). (*c*) Power spectrum of the Niño-3.4 SST anomaly time series based on 1950–2008 (*solid curve*) and 1900–2008 (*dashed curve*). (*d*) Monthly standard deviation of the Niño-3.4 SST anomaly time series.

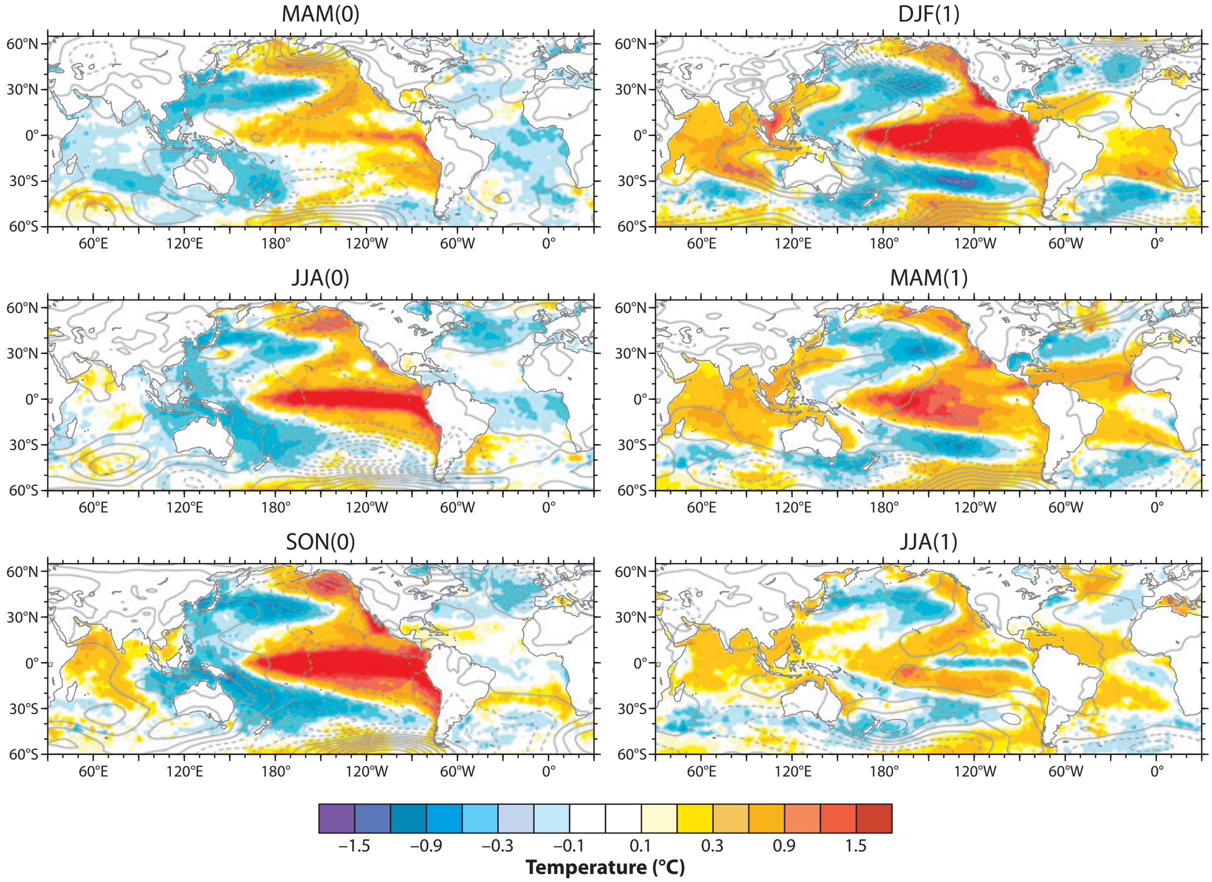

Figure 6

Composite ENSO evolution of seasonal sea surface temperature (*shading*) and sea-level pressure (SLP) (*contours*) anomalies starting in March–May of year 0 (MAM0) and ending with June–August of the following year (JJA1). The SLP contour interval is 1 hPa, with negative values dashed.

seasons, reaching its peak around SON0 and DJF1. (Here, the first year of the ENSO event is denoted by a zero and the subsequent year by a one.) The warm event diminishes during MAM1 and begins to transition to a weak cold event in JJA1. Accompanying the warming is the simultaneous development of negative SST anomalies in the northwest and southwest tropical Pacific. The tropical Indian Ocean as well as portions of the tropical Atlantic exhibit a delayed warming relative to that in the Pacific, reaching peak strength around DJF1 and MAM1 but lingering through JJA1 and beyond. Similar results are shown in Harrison & Larkin (1998), Rasmusson & Carpenter (1982), Trenberth et al. (2002), and Klein et al. (1999), among others. Tropical SLP anomalies develop in tandem with the SST anomalies, reaching maximum strength in DJF1. This large-scale SLP anomaly pattern, with positive values west of the International Date Line and negative values to the east, is associated with the negative phase of the Southern Oscillation (see also Deser & Wallace 1990, Harrison & Larkin 1996).

Not all ENSO events follow the canonical evolution depicted in **Figure 6**. For example, some events develop mainly in the central equatorial Pacific while others are confined to the coast of South America (Deser & Wallace 1987, Kao & Yu 2009). In addition, some events show an eastward spreading of the equatorial Pacific SST anomalies (Guan & Nigam 2008) in contrast to the westward development of the composite ENSO.

ENSO is understood to be a natural mode of variability of the coupled ocean-atmosphere system in the tropical Pacific. Over the past 25 years, two broad paradigms have emerged to explain the governing dynamics of ENSO. One paradigm is the delayed oscillator or recharge oscillator theory, which holds that ENSO is an unstable mode of the coupled tropical ocean-atmosphere system exhibiting self-sustained, regular oscillations (Zebiak & Cane 1987, Battisti 1988, Schopf & Suarez 1988). Irregularity in the ENSO period can occur through stochastic forcing (noise) in the form of energetic weather events such as the Madden-and-Julian Oscillation and westerly wind bursts or through low-order chaos (Munnich et al. 1991, Jin et al. 1994, Blanke et al. 1997). The other leading paradigm is the stochastic forcing theory, which holds that ENSO is a damped stable mode of the system requiring energy from stochastic forcing to maintain the oscillation (Penland & Sardeshmukh 1995, Chang et al. 1996, Moore & Kleeman 1999, Thompson & Battisti 2001). Regardless of the precise nature of the ENSO mode, a series of positive and negative feedbacks between the atmosphere and ocean within the tropical Pacific lead to the growth and decay of an event. The dominant positive feedback is the so-called Bjerknes feedback between the wind anomalies in the western and central equatorial Pacific and the SST anomalies in the eastern Pacific. According to the Bjerknes feedback mechanism, westerly wind anomalies produce positive SST anomalies through downwelling equatorial oceanic Kelvin waves, and these SST anomalies in turn weaken the zonal SLP gradient, which leads to westerly wind anomalies. The dominant negative feedback involves the dynamical adjustment of the equatorial Pacific thermocline to the overlying wind field with a time delay that eventually leads to the demise of an event. Much progress has been made on the predictability of ENSO, leading to the routine issuance of ENSO forecasts (e.g., Palmer et al. 2004).

5.1.2. ENSO teleconnections: the atmospheric bridge. While air-sea interactions responsible for ENSO are centered over the equatorial Pacific Ocean, changes in tropical precipitation from deep convection associated with ENSO influence the global atmospheric circulation, in part through the excitation of Rossby waves that emanate into the extra tropics of both hemispheres. The wave energy tends to follow great circle routes, initially extending poleward and eastward, refracting (turning) away from the pole and returning to the tropics. These planetary waves form in preferred locations, which include high pressure in the subtropics and low pressure in midlatitudes of the North and South Pacific. The overall structure of the ENSO response is influenced by additional factors, including the detailed structure of the climatological wind field that governs the path of the wave energy, sources/sinks of Rossby waves outside of the tropical Pacific, and changes in the storm tracks (see reviews by Trenberth et al. 1998, Liu & Alexander 2007). The ENSO-driven circulation patterns alter the near-surface air temperature, humidity, and wind as well as the distribution of clouds far from the equatorial Pacific. The resulting variations in the surface heat, momentum, and freshwater fluxes induce changes in SST as well as salinity, mixed-layer depth, and ocean currents. Thus, the atmosphere acts as a bridge spanning from the equatorial Pacific to the remainder of the global oceans.

Atmospheric teleconnections to the North Pacific peak during DJF1, manifest by large amplitude negative SLP anomalies centered over the Gulf of Alaska (**Figure 6**). This pattern, the surface expression of the Pacific-North American, results in anomalous northwesterly winds that advect relatively cold dry air over the western/central North Pacific, anomalous southerly winds that advect warm moist air along the west coast of North America, and enhanced surface westerlies over the central North Pacific. The resulting anomalous surface heat fluxes and Ekman transport create negative SST anomalies in the central and western North Pacific and positive SST anomalies along the west coast of North America (Alexander et al. 2002).

The wintertime ENSO response also impacts the North Atlantic (**Figure 6**). During El Niño events, a pattern similar but not identical to the NAO develops in DJF1, with anomalous high (low) pressure in high (mid) latitudes in conjunction with a southward displacement and intensification of the storm track along the eastern seaboard. This change in the atmospheric circulation creates a large-scale pattern of SST anomalies through the mechanisms discussed in Section 2. The SST response most closely resembles the tripole pattern in MAM1, with negative anomalies in mid-latitudes flanked by positive anomalies in the subarctic and subtropics. Observational studies and model experiments suggest that the ENSO teleconnection over the North Atlantic is stronger and more robust during La Niña events compared with El Niño events (Pozo-Vazquez et al. 2001).

ENSO-driven atmospheric teleconnections to the Southern Hemisphere are in many ways symmetric to those in the Northern Hemisphere (**Figure 6**). For example, negative SLP anomalies develop over the South Pacific in austral winter (JJA0), in analogy with those over the North Pacific in boreal winter (DJF1). During austral summer (DJF1), the remote response to ENSO includes not only a wave-like response but also a component that projects onto the SAM, with negative SLP anomalies over mid-latitudes and positive SLP anomalies over high latitudes. The Southern Hemisphere SST response to these atmospheric circulation anomalies includes negative anomalies at ∼30°S and positive anomalies near the Antarctic Peninsula during El Niño events. These SST anomalies are primarily driven by surface heat fluxes in mid-latitudes and Ekman transport in high latitudes in both summer and winter (Ciasto & Thompson 2008). The amplitude of the SST anomalies is largest during austral summer, when the mixed layer is shallow and thus the ocean can rapidly respond to thermal forcing.

As mentioned above, the tropical Indian Ocean warms following an El Niño event, with SST anomalies that lag those in the central Pacific by approximately 3–6 months. Part of the basin-wide warming is generated by the diminished wind speeds, which reduce the upward latent heat flux, and by the reduced cloudiness, which allows more solar radiation to reach the surface (Klein et al. 1999). Ocean dynamics also influences El Niño–related SST anomalies in parts of the Indian Ocean: Westward-propagating oceanic Rossby waves generated by anomalous wind-stress curl in the southern Indian Ocean contribute by depressing the relatively shallow thermocline at ∼10°S (Xie et al. 2002). The slow propagation of the oceanic Rossby waves maintains the Indian Ocean SST anomalies through JJA1, with associated atmospheric teleconnections to the subtropical Northwest Pacific and East Asia (Xie et al. 2009).

5.1.3. Non-ENSO variability in the tropical Indian and Atlantic oceans. The two leading EOFs of detrended monthly SST anomalies over the tropical Indian Ocean (20°N–20°S) are the basin-wide mode and the east-west dipole mode (**Figure 7**). The former is closely linked to ENSO as described in the previous section, whereas the latter occurs both in conjunction with and independently of ENSO. The two modes differ not only in their spatial expression, but also in their seasonal dependency: The basin-wide mode peaks in January–March, whereas the dipole mode peaks in July–October (**Figure 7**). The positive phase of the dipole mode exhibits weak positive SST anomalies over the western two-thirds of the basin and strong negative SST anomalies in the vicinity of Indonesia. A positive dipole mode is evident in the ENSO composite during SON0 (**Figure 6**).

Bjerknes feedback appears to contribute to the development of the dipole mode: The eastern cooling induces easterly wind anomalies, which in turn drive equatorial upwelling that shoals the thermocline, thereby reinforcing the SST cooling (Saji et al. 1999). The east-west seesaw in the thermocline depth is part of the transient adjustment in response to wind variability near

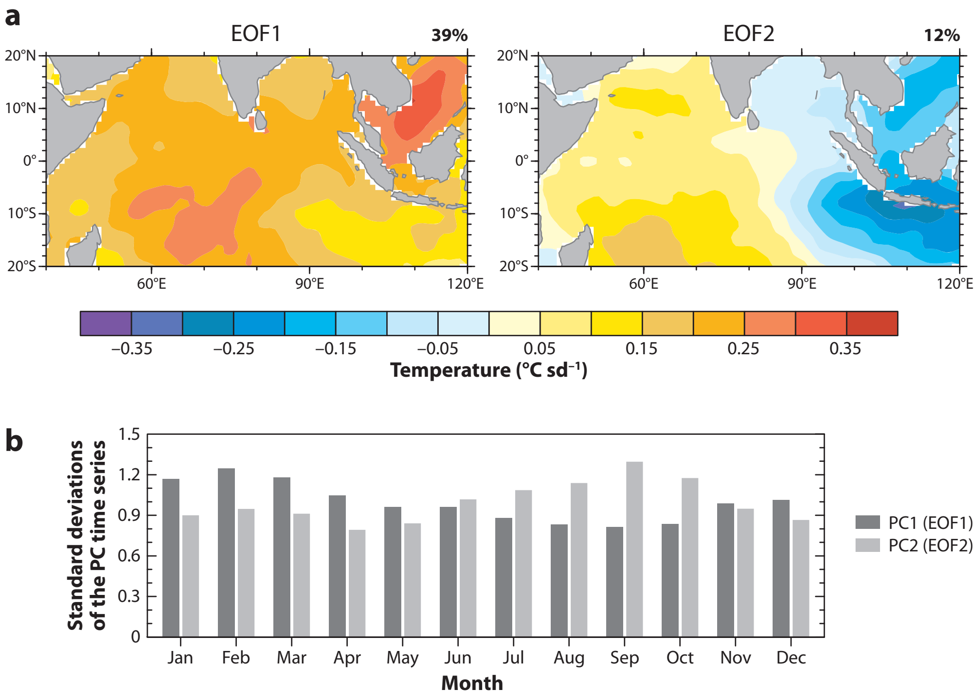

Figure 7

(*a*) First and second empirical orthogonal functions (EOFs) of detrended monthly SST anomalies in the tropical Indian Ocean based on the HadISST data set during 1900–2008. These modes account for 39% and 12% of the variance, respectively. EOF1 depicts the basin-wide mode and EOF2 depicts the dipole mode. (*b*) Monthly standard deviations of the principal component time series associated with EOF1 (*dark gray columns*) and EOF2 (*light gray columns*).

the equator. In fact, the dipole mode stands out as the leading mode in EOF analysis of thermocline depth variability (Shinoda et al. 2004). The SST dipole mode is strongest during July–November when upwelling-favorable southeasterly alongshore winds prevail off the coast of Indonesia, and it is linked to a similar dipole in precipitation (negative in the east and positive in the west during the positive phase). See Schott et al. (2009) for a recent review of Indian Ocean climate variability. Predictability of the Indian Ocean dipole mode is discussed in Luo et al. (2007).

There are two major modes of tropical Atlantic variability: an equatorial mode akin to the Pacific El Niño and a meridional mode characterized by cross-equatorial SST anomaly gradients (a similar mode is also found in the tropical Pacific). Recent reviews of these modes are provided in Xie & Carton (2004) and Chang et al. (2006).

The equatorial Atlantic cold tongue displays significant year to year variability, with typical amplitude ~0.5°C and frequency ~2–3 years, although the latter is dependent on the time period analyzed. This equatorial mode, called the Atlantic Niño, appears as the leading mode of tropical Atlantic variability with strongest amplitude during May–July (**Figure 8**). It is similar to ENSO in that ocean-atmospheric interactions give rise to positive feedbacks similar to those envisioned by Bjerknes for ENSO in the Pacific. The anomalous warming in the eastern equatorial Atlantic relaxes the easterly trade winds, which reinforce the eastern warming by weakening the equatorial upwelling and lowering the thermocline (Zebiak 1993). The Atlantic Niño intensifies atmospheric convection near the equator, and it keeps the summer rain band from advancing into the Sahel

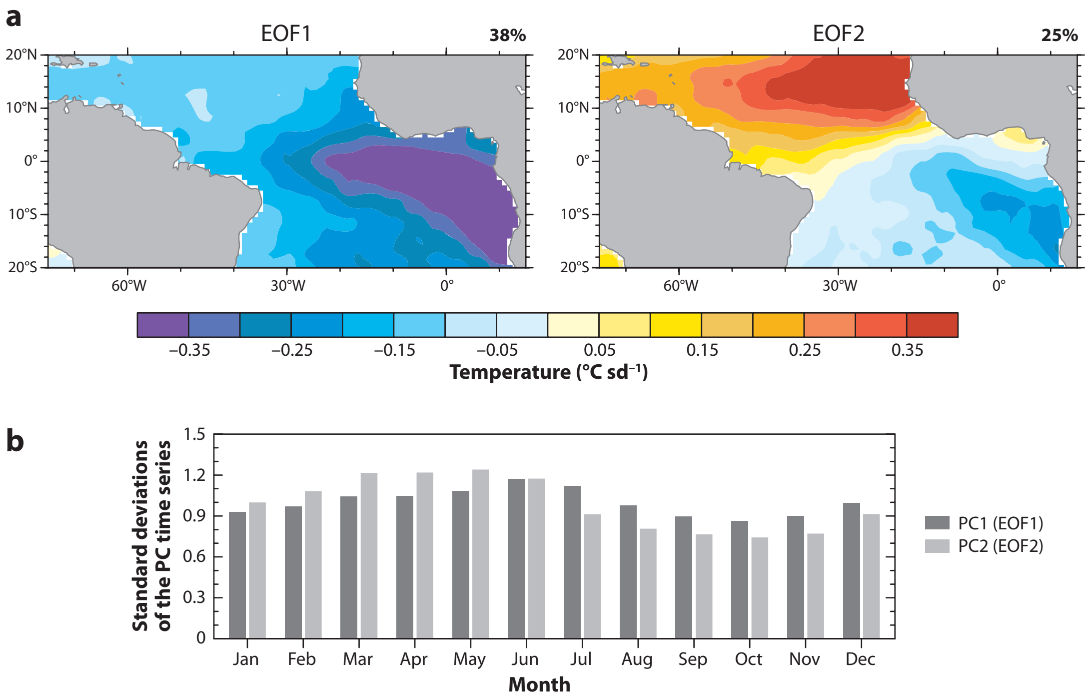

Figure 8

(*a*) First and second empirical orthogonal functions (EOFs) of detrended monthly sea surface temperature anomalies in the tropical Atlantic Ocean based on the HadISST data set during 1900–2008. These modes account for 38% and 25% of the variance, respectively. EOF1 depicts the Atlantic Niño mode and EOF2 depicts the meridional mode. (*b*) Monthly standard deviations of the principal component time series associated with EOF1 (*dark gray columns*) and EOF2 (*light gray columns*).

while increasing rainfall at the Gulf of Guinea coast (Giannini et al. 2003). The Atlantic Niño is weaker and higher in frequency than its Pacific counterpart.

The second EOF mode of tropical Atlantic SST variability shows a meridional dipole pattern with the nodal line displaced slightly north of the equator (**Figure 8**). The warm (cool) pole is associated with the relaxed (intensified) easterly trades (Nobre & Shukla 1996). Such a wind-SST relation suggests the following positive feedback via surface evaporation (Chang et al. 1997). An initial SST dipole induces southerly winds across the equator. The Coriolis force causes these southerlies to gain a westerly (easterly) component in the warm (cool) hemisphere, decelerating (accelerating) the prevailing easterly trades. The resultant changes in surface evaporation act to amplify the initial SST dipole. This wind-evaporation-SST feedback preferentially amplifies equatorial-antisymmetric disturbances while the Bjerknes feedback grows east-west variations on the equator. A similar phenomenon exists in the Pacific (Chiang & Vimont 2004), with growing evidence that it may play an important role in initiating ENSO events (Chang et al. 2007).

The meridional mode is most pronounced during the equatorial warm season March-May, with SST anomalies typically ~0.5°C in the subtropics. The attendant SST gradients across the equator are very effective in displacing the Atlantic intertropical convergence zone (ITCZ) into the warmer hemisphere. Rainfall in the semiarid Nordeste region of Brazil is sensitive to this meridional mode and the ITCZ shifts it induces (Nobre & Shukla 1996). The meridional mode also affects North Atlantic hurricanes through its influence on SST in the main development region (90–20°W, 5–25°N) (Wang et al. 2006, Vimont & Kossin 2007).

Besides wind-evaporation-SST feedback, there are several other factors for variability in the cross-equatorial SST gradient. For example, ENSO affects SSTs over the tropical North Atlantic (recall **Figure 6**), whereas the North Atlantic Oscillation forces the same region by changing the northeast trade winds (recall **Figure 1**) (see also Xie & Tanimoto 1998, Czaja et al. 2002). These and other factors result in a weak correlation between subtropical SST anomalies north and south of the equator. It has been suggested that the North Atlantic tripole (Deser & Blackmon 1993) and a South Atlantic dipole (Venegas et al. 1997) are correlated to form a pan-Atlantic pattern through the coordination of wind-evaporation-SST feedback across the equator (Xie & Tanimoto 1998).

5.2. Extra Tropics

In this section, we consider the dominant patterns of SST variability in the Southern Hemisphere, the Pacific Decadal Oscillation, and the Atlantic Multidecadal Oscillation.

5.2.1. Southern Hemisphere. Given the paucity of SST measurements from ships of opportunity, studies of SST variability in the Southern Hemisphere rely mainly on satellite-based data sets. The leading EOF pattern of SST anomalies south of 20°S from the NOAA Optimum Interpolation SST data set for the period 1981–2005 from Ciasto & Thompson (2008) is reproduced in **Figure 9**. This analysis was performed separately for the warm season, November–April, and the cold season, May–October. Superimposed on the SST EOF patterns are the associated 500 hPa geopotential height anomaly patterns, obtained by regressing seasonal anomalies at each grid box on the SST PC time series. Similar patterns are obtained for SLP (not shown). In both seasons,

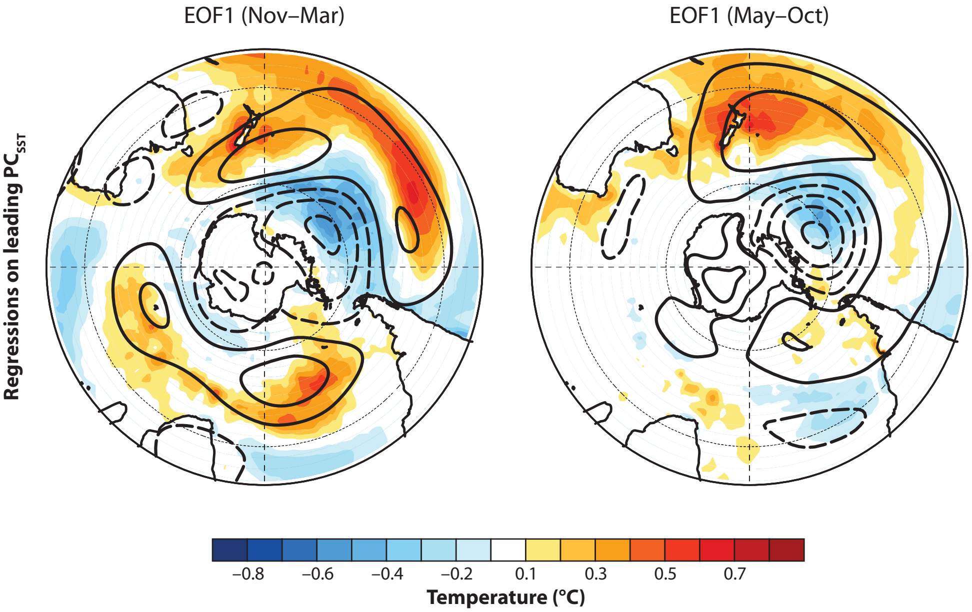

Figure 9

Sea surface temperature (SST) (*shading*) and 500-hPa geopotential height (*contours*) anomaly patterns associated with the leading principal component time series of monthly SST anomalies in the Southern Hemisphere (south of 20°S) during November–March (*left*) and May–October (*right*). The contours are drawn at ±35 m, 25 m, 15 m, and 5 m, with negative values dashed. Adapted with permission from Ciasto & Thompson (2008).

the leading EOF (in its positive polarity) exhibits positive SST anomalies across much of the South Pacific and negative SST anomalies in the Pacific sector of the Southern Ocean. The warm-season pattern also exhibits positive SST anomalies across the Indian and Atlantic sectors of the Southern Ocean, with negative anomalies to the north and adjacent to the coast of Antarctica.

As Ciasto & Thompson (2008) demonstrated, the SST anomalies associated with the leading mode of variability are largely atmospherically forced via changes in the surface turbulent fluxes and to a lesser extent Ekman currents. The associated 500 hPa geopotential height anomaly pattern resembles that due to ENSO (recall **Figure 6**), with an additional component related to the SAM during the warm season. Although the SST anomalies depicted in **Figure 9** are primarily forced by the atmosphere, there is evidence for a back effect onto the atmosphere, manifest as an increase in the persistence time for the SAM above that due to intrinsic atmospheric dynamical processes (Sen Gupta & England 2007).

5.2.2. Pacific Decadal Oscillation. The leading EOF of monthly SST anomalies over the North Pacific (after removing the global mean SST anomaly) and its associated PC time series are termed the Pacific Decadal Oscillation (PDO) after Mantua et al. (1997). This mode, which accounts for 25% of the variance of monthly anomalies in the HadISST data set during the period 1900–2008, is displayed in **Figure 10**. Although the EOF calculation was restricted to the North Pacific (20°–70°N), the pattern is displayed globally by regressing the monthly SST anomalies at each location on the PC time series. The spatial pattern of SST anomalies associated with the PDO is similar to that associated with ENSO (recall **Figure 6**) except for the relative weighting between the north and tropical Pacific. In particular, the amplitudes of the SST anomalies in the equatorial eastern Pacific compared with those in the North Pacific are comparable for the PDO; however, they are considerably larger for ENSO (see also Zhang et al. 1997, Garreaud & Battisti 1999, Dettinger et al. 2000, Deser et al. 2004). The PDO has also been termed the Interdecadal Pacific Oscillation (Power et al. 1999) in recognition of its extension to the South Pacific.

The time series of the PDO exhibits considerable decadal variability, hence its name (**Figure 10**, bottom). Even the raw monthly anomalies predominantly show periods in which one sign lasts for two decades or longer (for example, the periods 1908–1945, 1947–1976, and 1977–1998). The decadal SST transitions are accompanied by widespread changes in the atmosphere, ocean, and marine ecosystems. For example, Mantua et al. (1997) found that the timing of changes in the PDO closely corresponded to those in salmon catch along the west coast of North America. The Aleutian low pressure system also fluctuates in tandem with the PDO SST variations (Deser et al. 2004). Given the relatively short observational record, it is difficult to ascertain whether there is a robust spectral peak in the PDO time series. Some studies suggest nominal timescales of ~20 years and ~50 years (Minobe 1997, Minobe 1999, Deser et al. 2004), whereas others conclude that the PDO record does not differ significantly from a first-order autoregressive process (e.g., red noise: Pierce 2001, Qiu et al. 2007).

Many different mechanisms have been proposed for the PDO, including random forcing of the ocean by the atmosphere according to the stochastic climate model paradigm (recall Section 2), the atmospheric bridge from the tropical Indo-Pacific, extra-tropical ocean processes, and atmosphere-ocean interactions within the North Pacific (see the recent review in Alexander 2009). How can we reconcile these different mechanisms for the PDO? Several studies have used statistical analyses to reconstruct the PDO and determine the processes that underlie its dynamics (Newman et al. 2003, Vimont 2005, Schneider & Cornuelle 2005, Qiu et al. 2007, Newman 2007). Taken together, theses studies indicate that on interannual timescales, random and ENSO-induced fluctuations in the strength of the Aleutian low are almost equally important in determining PDO variability via surface heat flux forcing, with negligible contribution from

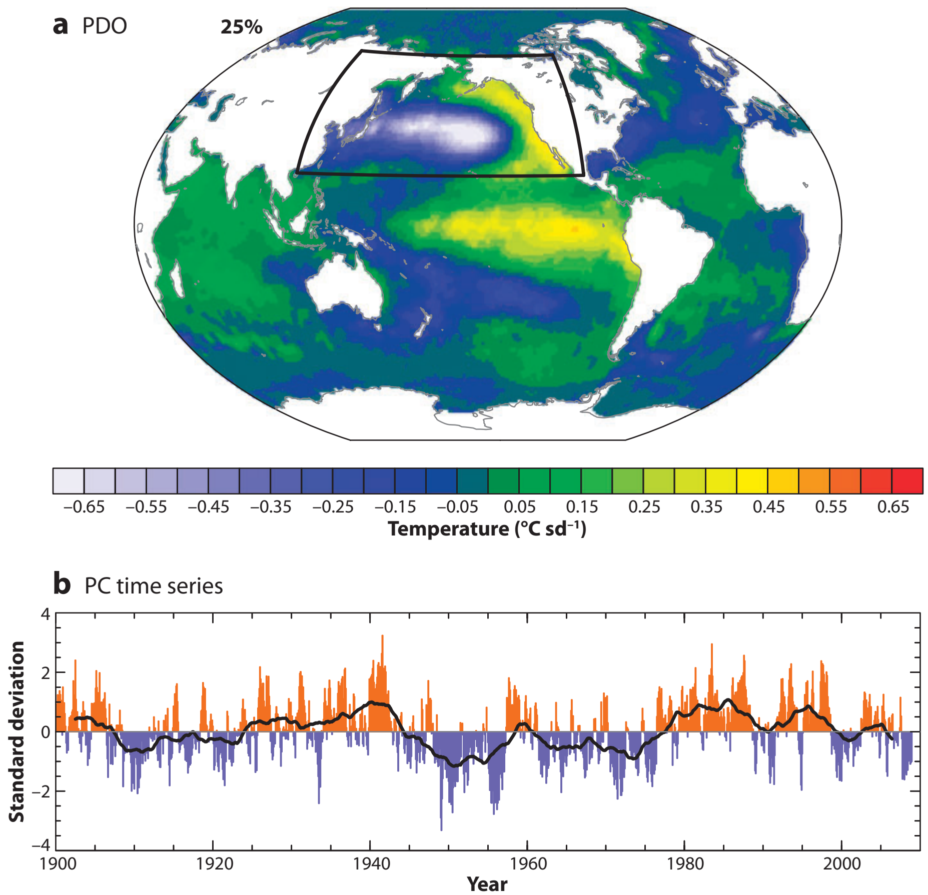

Figure 10

Pacific Decadal Oscillation. (*a*) The leading empirical orthogonal function (EOF) of monthly sea surface temperature (SST) anomalies over the North Pacific (after removing the global mean SST anomaly) based on the HadISST data set during 1900–2008. Although the EOF calculation was restricted to the North Pacific (*region outlined by the black rectangle*), the pattern is displayed globally by regressing the monthly SST anomalies at each location on the principal component (PC) time series. (*b*) Associated PC time series showing the unsmoothed record (*red and blue bars*) and the 5-year running mean record (*black line*).

ocean currents. On decadal timescales, stochastic heat flux forcing, the atmospheric bridge, and changes in the North Pacific oceanic gyre circulation contribute approximately equally. Over the western Pacific east of Japan where a deep mixed layer develops during winter, ocean circulation changes associated with latitudinal excursions of the Kuroshio Current Extension are of primary importance for PDO-related SST variability (Nonaka et al. 2006, Taguchi et al. 2007). A key implication of these analyses is that, unlike ENSO, the PDO is likely not a single physical mode but rather the sum of several phenomena.

The PDO is only one measure of SST variability in the North Pacific. Other recurring patterns include the North Pacific mode (Deser & Blackmon 1995, Nakamura et al. 1997, Barlow et al. 2001, Guan & Nigam 2008), which is closely related to the PDO albeit with less amplitude in the tropical Indo-Pacific. A distinct pan-Pacific mode has also been identified, emphasizing variability in the eastern portion of the North Pacific (Guan & Nigam 2008). A North Pacific Gyre Oscillation mode is also evident in sea surface height variability (Di Lorenzo et al. 2008). The mechanisms governing these modes are not well understood.

5.2.3. Atlantic Multidecadal Oscillation. SSTs in the North Atlantic undergo slow variations with a period on the order of 65–80 years (Kushnir 1994, Schlesinger & Ramankutty 1994, Delworth & Mann 2000, Enfield et al. 2001, Ting et al. 2009, Guan & Nigam 2009). This phenomenon has been termed the Atlantic Multidecadal Oscillation (AMO) (Kerr 2000). A simple index of the AMO is shown in **Figure 11** for the period 1870–2008 based on the HadISST data set (note that SST observations are relatively plentiful in the North Atlantic back to 1870; recall **Figure 3**). This index is defined as the area-average SST anomaly over the North Atlantic (0°–70°N) minus the global mean SST anomaly. Warm phases occurred from the late 1920s through the late 1960s and since the mid-1990s, and cool phases occurred from the early 1900s through

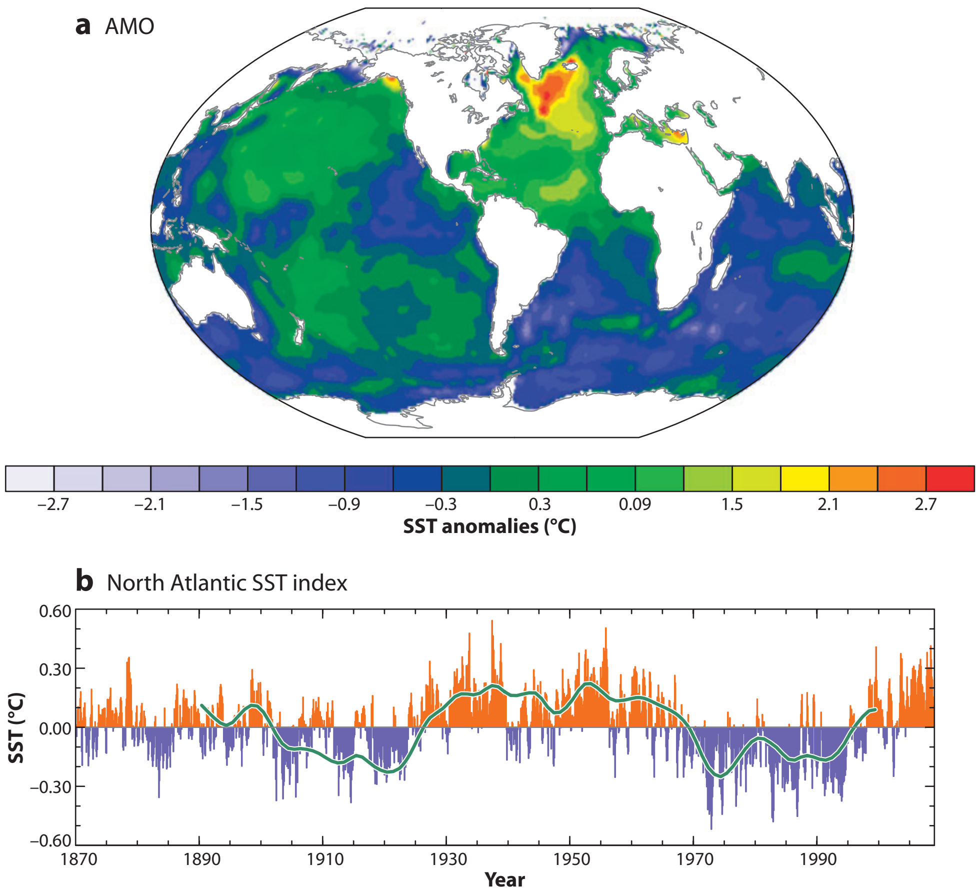

Figure 11

Atlantic Multidecadal Oscillation. (*a*) Regression pattern of monthly sea surface temperature (SST) anomalies (after removing the global mean SST anomaly) on the North Atlantic SST Index, based on the HadISST data set during 1870–2008. (*b*) The North Atlantic SST Index, defined as the average monthly SST anomaly over the North Atlantic (0°–70°N) minus the global mean monthly SST anomaly (*red and blue bars*). The green line depicts an estimate of the natural (e.g., not due to forcing external to the ocean-atmosphere system) component of the 10-year low-pass-filtered North Atlantic SST index from Ting et al. (2009).

the mid-1920s and from the early 1970s through the mid-1990s. The spatial pattern associated with this time series, obtained by linearly regressing the SST anomalies at each location on the AMO index, exhibits positive values over the entire North Atlantic, with the largest magnitudes (approximately 0.5°C) south of Greenland.

The AMO is associated with large-scale precipitation changes, most notably over the Sahel, the southeastern United States, and Brazil (Enfield et al. 2001, Ting et al. 2009), and with a frequency of severe Atlantic hurricanes (Enfield et al. 2001). It should be noted that the AMO is not significantly correlated with the NAO (Guan & Nigam 2009). The precise nature of the AMO is still being refined and clarified. In particular, the role of Pacific SST variability is not well understood. For example, Ting et al. (2009; see also Enfield et al. 2001) showed some connection between the AMO and SST anomalies in the Gulf of Alaska and the tropical Pacific (see also **Figure 11**), whereas the analysis of Guan & Nigam (2009) indicates that there are no significant SST linkages outside of the North Atlantic.

The AMO is considered to be a natural mode of oscillation of the Atlantic Ocean's thermohaline circulation (Delworth & Mann 2000). Modeling studies indicate that this mode is intrinsic to the ocean and stochastically forced by atmospheric buoyancy fluxes (Delworth & Greatbatch 2000). However, the amplitude of the mode is augmented owing to coupled ocean-atmosphere interactions.

Considerable debate concerns the degree to which anthropogenic effects may be contributing to the recent positive phase of the AMO. In particular, there has been a general warming trend over both land and ocean associated with increasing greenhouse gas concentrations. Because the AMO index is a temperature-based record, it may contain the anthropogenic global warming signal and thus partially confound the true state of the AMO. There have been various attempts to remove the anthropogenic global warming signal from the AMO index, including simple linear detrending, removal of the global mean temperature anomaly (as in **Figure 1**), and removal of model-based estimates of the forced component of variability. The latter approach provides the best estimate of the natural component of the AMO (Ting et al. 2009). This natural component (green curve in **Figure 11**) indicates that the AMO is currently in a modest warm phase, not an extreme warm phase as would be inferred from the raw SST index (Ting et al. 2009, Guan & Nigam 2009). However, the Atlantic thermohaline circulation may itself be altered under anthropogenic forcing as projected by some climate models (Dixon et al. 1999, Wood et al. 1999), but this issue is distinct from that of the overall global warming trend in the temperature-based AMO index.

SUMMARY POINTS

1. Interannual and longer timescale SST variations exhibit large-scale organization owing to both atmospheric and oceanic processes.
2. Such large-scale organization may be due to intrinsic modes of atmospheric circulation variability that imprint themselves upon the SST field mainly via surface energy fluxes but also via Ekman currents. Examples include SST fluctuations in the Southern Ocean associated with the Southern Annular Mode, a tripolar pattern of SST anomalies in the North Atlantic associated with the North Atlantic Oscillation, and a pan-Pacific mode known as the Pacific Decadal Oscillation. Oceanic processes also contribute to the latter.
3. A leading paradigm for these atmospherically forced patterns of SST variability is the stochastic climate model whereby random fluctuations in the atmospheric circulation give rise to a low-frequency SST response.

4. Patterns of nonseasonal SST variability may also result from coupled ocean-atmosphere interactions, most prominently the El Niño-Southern Oscillation phenomenon in the tropical Indo-Pacific sector, but also the tropical Atlantic Niño and the cross-equatorial meridional modes of the tropical Pacific and Atlantic.
5. Intrinsic oceanic modes also may give rise to large-scale patterns of SST variability. A leading example is the Atlantic Multidecadal Oscillation associated with changes in the oceanic thermohaline circulation.

FUTURE ISSUES

A number of outstanding issues remain regarding the patterns and mechanisms of nonseasonal SST variability; these are exemplified in the following questions:

1. Does the response to global warming (e.g., increasing greenhouse gas concentrations) project onto known patterns of natural SST variability, or does it create new patterns?
2. How much does feedback from the ocean to the atmosphere influence the atmospherically driven modes of extratropical SST variability?
3. What is the nature of the predictability of the tropical and extratropical modes of SST variability on decadal timescales?
4. What are the linkages among the different natural modes of SST variability, and what mechanisms produce these connections?
5. Do high-resolution satellite data sets reveal new patterns of SST variability?
6. Do empirical methods such as EOF analysis identify the true physical modes of SST variability?
7. How do the patterns of SST variability relate to those in the ocean interior?

DISCLOSURE STATEMENT

The authors are not aware of any biases that might be perceived as affecting the objectivity of this review.

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Errata

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