

# Ocean–Atmosphere Interactions on Interannual to Decadal Time Scales

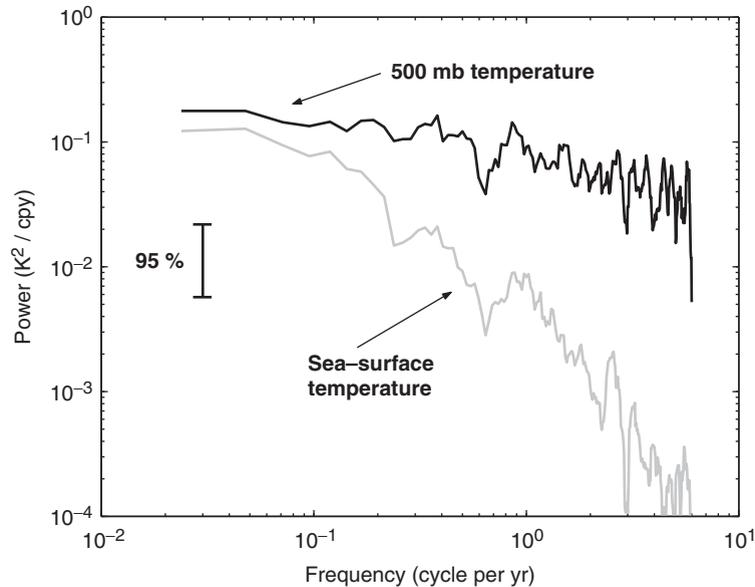
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## 1 INTRODUCTION TO OCEAN– ATMOSPHERE INTERACTIONS

Climate variability on interannual to interdecadal time scales results from multiple interactions among different subcomponents of the climate system, of which the atmosphere and the ocean are arguably the two most important. While here we will focus exclusively on ocean–atmosphere interactions, it is important to keep in mind that climate variability also results from interactions and feedbacks between the atmosphere and the cryosphere, and the land surface and the biosphere. Ocean–atmosphere interactions include exchanges of energy (i.e., radiative transfer and heat fluxes), momentum (i.e., wind stress), water (i.e., precipitation and evaporation) or trace constituents (e.g., CO<sub>2</sub> exchange). The two components, ocean and atmosphere, are characterized by very different time scales of variability. The lower troposphere exhibits a large diurnal cycle and rapidly adjusts to changes in the underlying heat source, with mixing and transport occurring within a matter of days (continental scale) to weeks (global scale). The ocean, on

the other hand, demonstrates a diurnal to annual cycle, but large-scale mixing occurs on time scales from years to decades or even hundreds to thousands of years for mixing of the deep-ocean circulation. The ocean has a much larger heat capacity: its adjustment to temperature change is ~10 months for a 100-m mixed-layer depth, and ~40 years for a 5-km depth, with slow deep-ocean circulation increasing this to ~1000 years (Bigg, 1996). Even if the ocean had no circulation (no transport of heat and salt around the globe), it would still significantly buffer climate variability and change. Figure 22.1 shows the frequency spectrum of sea-surface temperature (SST) and 500 hectopascal (hPa) temperature in the North Atlantic. Clearly the SST spectrum is much redder (the SST variability at short time scales is dampened out), while the atmospheric temperature shows high power at all time scales.

This variation in behavior is the fundamental cause of many delayed ocean–atmosphere oscillations as the slower component (the ocean) tends to act as buffer that delays change and transitions to a new state. The built-in memory of a perturbation leads to a



**Figure 22.1** Log-log plot of the power spectrum of atmospheric temperature at 500 mbar (black) and SST (gray) associated with the North Atlantic Oscillation. Reproduced from Czaja et al. (2003) with permission from the American Geophysical Union.

negative feedback, responsible for the system transition from one state to the other. The memory effect is equally important since a feedback that acts quasi-instantaneously will only dampen a signal but not lead to an oscillatory evolution. Hence a negative feedback and a memory both represent necessary elements for the generation of the oscillatory character of a climate mode.

Interactions and feedbacks between ocean and atmosphere show different characteristics at low and high latitudes (e.g., Wang et al., 2004). In general, changes in the atmosphere tend to precede oceanic fluctuations at mid to high latitudes. Climate variability on interannual time scales at mid latitudes is therefore considered to primarily reflect the slow response of the ocean to forcing imposed by atmospheric variability occurring on much shorter synoptic time scales. This so-called passive variability can be described as a slower component (the ocean) being modulated by random (stochastic) variability of the faster component

(the atmosphere). In the tropics changes tend to be more in phase due to the sensitivity of the atmosphere to moist convection triggered by high SSTs. This can lead to active variability and coupled interactions in which changes in one system mutually reinforce changes in the other through positive feedbacks and amplification of the signal.

The internal variability exhibited within the coupled ocean–atmosphere system is influenced by additional factors such as inherent variability (deterministic or random) or external forcings (solar, volcanic, greenhouse gases), which may lead to changes in its behavior over time and thereby complicate the diagnosis of the mechanism involved. Studying modes of ocean–atmosphere variability on different time scales and under different environmental background conditions in the past is therefore essential to fully understand the processes involved.

The following sections discuss the characteristics of the main modes of climate variability on interannual to decadal time scales

that arise from, or are otherwise affected by, ocean-atmosphere interactions in both the tropics and high latitudes. The main emphasis is on characterizing the spatiotemporal variability of these modes, their dynamic behavior and their potential for prediction, as many are of high societal relevance with significant impacts on natural and human systems. Given the short time span for which observational data is available (typically less than a century), particularly over the oceans, and the rapid pace at which we are altering the chemical composition of our atmosphere, these results need to be put in a longer temporal context. This is achieved by reviewing our current understanding of past ocean-atmosphere variability based on reconstructions from proxy data and by discussing the most recent results from model-based projections of changes in climate mode behavior under various twenty-first-century greenhouse gas emission scenarios.

## 2 THE EL NIÑO SOUTHERN OSCILLATION (ENSO)

### 2.1 ENSO theory and dynamics

The mean state of the tropical Pacific coupled ocean-atmosphere system is characterized by warm surface water (29–30°C) in the west ('warm pool') and much colder SST in the east (22–24°C). The warm pool is a region of intense rainfall and atmospheric heating with a relatively deep thermocline (100–200 m below the surface), while the cold, dry eastern Pacific is the result of cold water upwelled from below, forced by the trade winds converging along the equator. Collectively, this east-west SST gradient and the associated easterly trade winds constitute a state of quasi-equilibrium for the ocean-atmosphere system. The thermocline depth is a direct result of wind stress forcing by the overlying trade winds. Close to the equator trade winds advect surface waters westward causing water to 'pile up' on the western side

of the basin, leading to a tilt in the thermocline. Away from the equator the Coriolis force leads to opposing diverging forces acting on trades in each hemisphere, resulting in Ekman transport away from the equator. This divergence of surface water is compensated by equatorial upwelling, reflected as a 'cold tongue' in eastern Pacific SST (e.g., Neelin et al., 1998; Wallace et al., 1998).

As a result of the east-west SST gradient the western Pacific warm pool region is characterized by low surface pressure, vertical ascent and deep convection over the region of warmest SST, while the cold, eastern Pacific is characterized by high surface pressure and large-scale subsidence. The equatorial trades are maintained by this pressure gradient, blowing from the South American coasts westward, but at the same time also act to sustain this gradient through their influence on the SST distribution. Hence the state of the tropical Pacific is maintained by a coupled positive feedback where colder temperatures in the east drive stronger easterlies which in turn drive greater upwelling, pull the thermocline up more strongly, and thereby further cool temperatures. This feedback mechanism was first described by Bjerknes (1969) who noted the peculiar character of the equatorial Pacific, where the equatorial oceans all receive the same solar insolation, yet the Pacific is 4–10°C colder in the east than in the west (Bjerknes, 1969). It is now widely known as the 'Bjerknes feedback'.

Approximately every 3–7 years this state of quasi-equilibrium breaks down, usually after the easterly winds relax due to a brief disturbance (e.g., due to westerly wind bursts along dateline). As a result of the reduced Ekman transport away from the equator the upwelling of cold water is reduced, the thermocline deepens in the east and SSTs in the east warm up. This warming reduces the zonal gradients of SST and sea-level pressure (SLP), further weakening the trade winds. Eventually the ocean warms up across the entire eastern equatorial Pacific and the warm pool is displaced to the east, causing

shifts in precipitation and disruption of climate patterns at higher latitudes (Figure 22.2). These warm episodes, known as El Niño events, were first described by Peruvian fisherman, who observed warm ocean currents that ran southward along the coast of Peru and Ecuador around Christmas (hence Niño, Spanish for 'the Christ-child'). By now, however, these events are recognized as basin-wide phenomena. In its opposite phase (La Niña) SSTs in the central and eastern equatorial Pacific are unusually cold, the zonal SST and SLP gradients are enhanced and easterly trade winds are anomalously strong (Figure 22.2).

The explanation of ENSO through the Bjerknes feedback locks together the eastern Pacific SST and the SLP gradient (the Southern Oscillation) into a single mode of the ocean-atmosphere system. The oscillation in sea level pressure between the eastern and western Pacific had already been discovered much earlier by Walker (1923) in his search for clues to predict monsoon variability in India. He first referred to this pressure seesaw as 'Southern Oscillation' (SO), but he did not have the data to recognize its coupling with Pacific SST. It was not until the discovery by Bjerknes (1969) that the circular argument involved in ENSO could be explained. But while the Bjerknes mechanism explains why the system has two favored states (El Niño and La Niña), it does not explain why it oscillates between the two. As discussed previously, every oscillation must contain some element that is not perfectly in phase with the other and for ENSO this element is the sluggish thermocline (or equivalently the upper ocean heat content), whose changes are not in phase with the winds driving them.

The Ekman pumping associated with westerly winds in the western and central Pacific at the onset of an El Niño event leads to downwelling along the equator and upwelling to the north and south. The resulting mass surplus at the equator travels eastward as a downwelling ocean wave trapped along the equator (a so-called Kelvin wave), while the

mass deficit areas to the north and south form westward-propagating, upwelling ocean waves (known as Rossby waves) (Battisti, 1988; Battisti and Hirst, 1989; Cane and Zebiak, 1985; Suarez and Schopf, 1988). The wave-induced thermocline depth perturbations influence SST in the upwelling regions of the central/eastern equatorial Pacific, such that the mixed layer needs to deepen in the east and become more shallow in the west, further amplifying the SST signal of the El Niño event and providing the necessary coupling with the atmosphere. Upon reaching the basin boundaries Kelvin and Rossby waves reflect and start to travel in the opposite direction while maintaining their surface height anomaly and down- and upwelling properties. Equatorial ocean waves in this delayed oscillator theory thereby offer a mechanism to reverse the phase of perturbations to the thermocline depth, albeit with a significantly dampened amplitude. The El Niño cycle is disrupted once cold SSTs appear in the central Pacific, promoting reappearance of easterlies and initiating El Niño's demise. Models based on this delayed oscillator theory show that the Bjerknes feedback combined with equatorial wave theory can generate a self-sustaining oscillation containing the typical ENSO characteristics (Chen et al., 2004).

Several theories have built upon this mechanism and stressed certain components that may be necessary prerequisites for El Niño events to occur. For example, observations show that prior to El Niño the heat content in the warm pool region is always anomalously high (e.g., McPhaden, 2004); that a significant correlation exists between excess heat stored and the severity of ensuing El Niño events, and that the time between El Niños may be determined in part by the time it takes to recharge the warm pool with excess heat (Jin, 1997).

The potential role of westerly winds in the western and central Pacific as triggers of El Niño has received wide attention (e.g., Lengaigne et al., 2004), but their role is not universally accepted. Indeed, observations

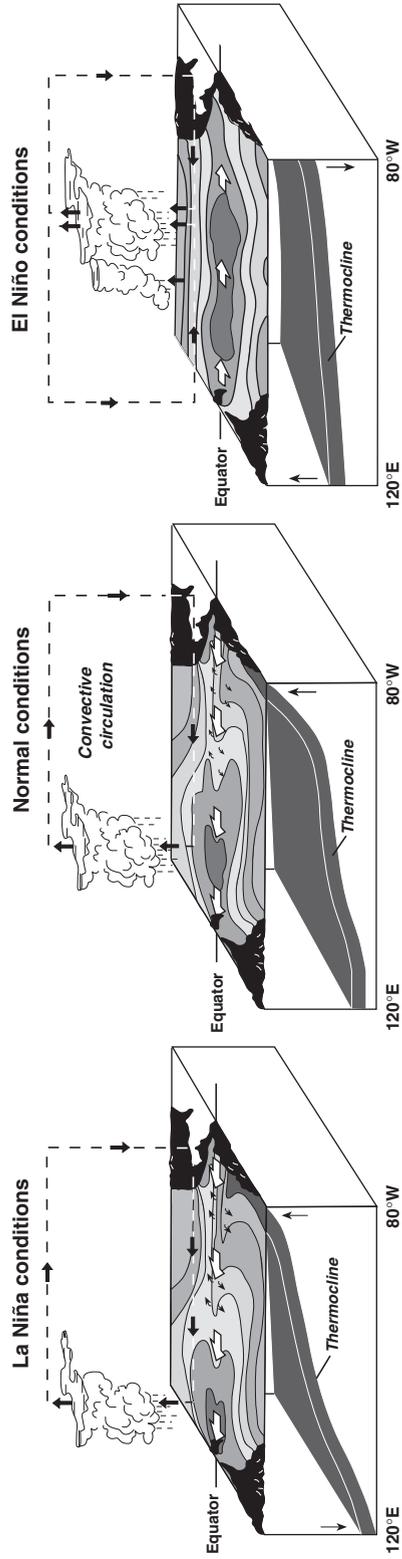


Figure 22.2 Conceptual diagram of ENSO mechanisms (courtesy of NOAA/PMEL/TAO Project Office).

show that the onset of El Niño is often marked by a series of prolonged westerly wind bursts in the western Pacific, which persist for 1–3 weeks and trigger downwelling equatorial Kelvin waves (e.g., McPhaden, 1999). The Madden–Julian oscillation (MJO), an eastward traveling atmospheric wave of enhanced and suppressed convective activity, responsible for much of the intraseasonal precipitation variability in the tropics and commonly observed over the tropical Indian and Pacific Oceans, may serve as such a stochastic forcing by producing westerly wind bursts that may trigger El Niño events. The diverging views of El Niño as one phase of a continual oscillation *versus* a sporadic phenomenon initiated by random westerly wind bursts may be reconciled by a hybrid view of El Niño as one phase of a damped oscillation that is sustained by random noise. Which theories are closer to reality is not just of academic interest but has implications for ENSO prediction. Predictability arises first and foremost from large-scale wave dynamics (oscillations), while random ‘noise’ driven by stochastic forcing will limit predictability.

Definitions on what constitutes an El Niño or a La Niña event vary widely, but some indices and definitions are more commonly used than others. Many are based on SST anomalies over specific regions of the equatorial Pacific (Table 22.1). Trenberth (1997), for example, defined an El Niño event as a time when the five-month running means of

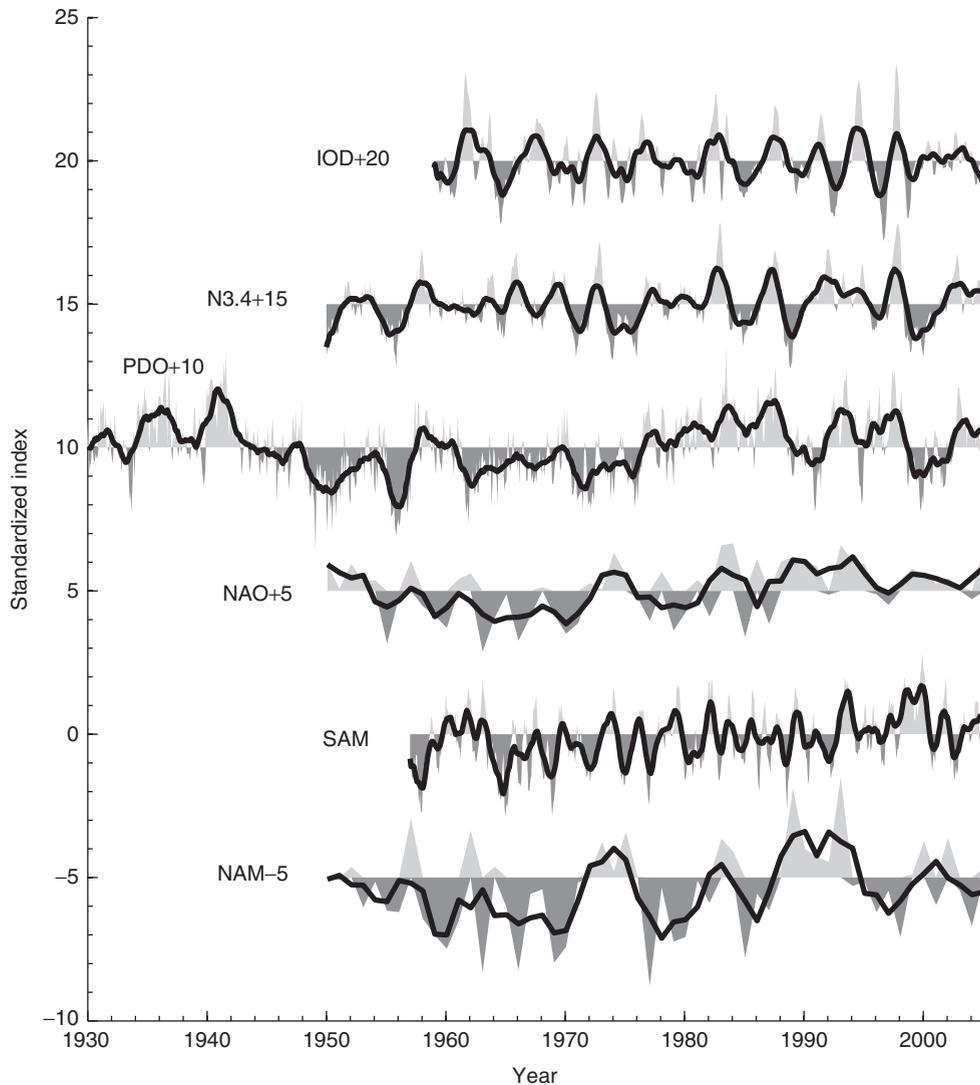
SST anomalies (SSTA) exceed  $0.4^{\circ}\text{C}$  for six months or more in the Niño 3.4 region. The NOAA Climate Prediction Center uses a modified definition in which a positive SSTA (from the 1971–2000 base period) in the Niño 3.4 region larger than  $0.5^{\circ}\text{C}$ , averaged over five consecutive months, constitutes an El Niño event. Other indices such as the Southern Oscillation Index (SOI) or the Multivariate ENSO Index (MEI) are based on the standardized SLP difference between the eastern and the western tropical Pacific or make use of a combination of different measures of ENSO or are defined as the first Empirical Orthogonal Function (EOF) of tropical Pacific SSTA (Table 22.1). The decision of which ENSO index to use is relevant as the result of an analysis may vary significantly depending on the choice of index. It is also worth noting that El Niño and La Niña events are nothing unusual. According to Trenberth (1997) the tropical Pacific experienced El Niño 31 percent of the time and La Niña’s 23 percent of the time over the past 50 years. Hence the Pacific is in either an El Niño or a La Niña phase more often than not (Figure 22.3), although these numbers depend on the definition used.

## 2.2 ENSO teleconnections, impacts and predictability

Latent heat release associated with anomalous tropical precipitation over the equatorial

**Table 22.1** Definition of main ENSO indices

<i>Index</i>	<i>Definition</i>	<i>References</i>
Niño3	SSTA over the box $5^{\circ}\text{N}$ – $5^{\circ}\text{S}$ , $90^{\circ}$ – $150^{\circ}\text{W}$	Deser and Wallace, (1990)
Cold Tongue Index (CTI)	SSTA over the box $6^{\circ}\text{N}$ – $6^{\circ}\text{S}$ , $90^{\circ}$ – $180^{\circ}\text{W}$	Deser and Wallace, (1990)
Niño3.4	SSTA over the box $5^{\circ}\text{N}$ – $5^{\circ}\text{S}$ , $120^{\circ}$ – $170^{\circ}\text{W}$	Trenberth (1997)
Southern Oscillation Index (SOI)	Standardized SLP difference between Darwin, Australia ( $12^{\circ}28'\text{S}$ , $130^{\circ}50'\text{E}$ ) and Tahiti ( $17^{\circ}32'\text{S}$ , $149^{\circ}37'\text{W}$ )	Ropelewski and Jones (1987)
Multivariate ENSO index (MEI)	Weighted average of several ENSO features, including SST, Outgoing longwave radiation and winds	Wolter and Timlin (1998)
ENSO-EOF	First Empirical Orthogonal Function (EOF) of tropical Pacific SSTA	Hoerling and Kumar (2000)



**Figure 22.3** Time series of ENSO, PDO, NAO, NAM, SAM and IOD. Note that ENSO, PDO, SAM and IOD are based on monthly data, while NAO and NAM show wintertime averages (DJF) only.

Pacific during ENSO events excites atmospheric equatorial Kelvin and Rossby waves which propagate zonally and lead to enhanced subsidence, warming the troposphere and suppressing precipitation. Rossby waves also propagate into the extratropics, especially in the winter hemisphere, where they cause large-scale circulation anomalies affecting

mid-latitude weather (Trenberth et al., 1998; Wallace et al., 1998). The mid-latitude response is barotropic, with pressure anomalies of the same sign throughout the troposphere. This wave train of alternating high and low pressure systems (stationary Rossby waves) is a forced response to changes in location and intensity of tropical heating.

Hence it is not the SSTs per se which provide the forcing, but the related tropical rainfall patterns. The two wave trains, the Pacific North American (PNA) and Pacific South American (PSA) patterns, show considerable hemispheric symmetry, with shifts in the mid-latitude jets and associated changes in storm tracks which ultimately cause the ENSO-related extratropical precipitation and temperature anomalies. However, it is important to keep in mind that ENSO teleconnections explain only a modest fraction of extratropical interannual variability and may suffer from nonstationarities (e.g., Diaz et al., 2001). While they form the cornerstone for seasonal prediction in the PNA and PSA regions, other sources of variability can mask ENSO signals on a case-by-case basis.

Atmospheric teleconnections are to first order approximately linear (stronger El Niños yield a stronger atmospheric response and the response to La Niña events is approximately inverse). However, during the ENSO peak phase teleconnections tend to show a stronger sensitivity to the warm than the cold phase (e.g., Hoerling and Kumar, 2000). This nonlinear response can be explained by (1) the spatial shift in rainfall between El Niño and La Niña due to the different spatial SSTA; (2) the fact that convection responds to the total rather than the anomalous SST; and (3) the SST threshold for convection (Hoerling et al., 2001).

ENSO also affects other ocean basins through perturbations in the Walker circulation which induce changes in cloud cover, evaporation, surface winds, and hence the net heat flux entering these remote oceans. This 'atmospheric bridge' leads to increased heat flux and positive SSTA in remote ocean basins such as the South China Sea, the Indian Ocean, and the tropical North Atlantic, approximately ~3–6 months after SSTA peak in the tropical Pacific (e.g., Alexander et al., 2004; Klein et al., 1999). The remote warming of tropical oceans leads to a coherent zonal mean warming throughout the troposphere from 30°N to 30°S during El Niño events. The peak warming, however, lags

behind tropical Pacific SST by approximately four months, due to the lagged response of the remote oceans (Chiang and Sobel, 2002).

While many regions see opposite climatic impacts during El Niño and La Niña events, there are some asymmetries in precipitation responses to El Niño and La Niña. It should therefore not be *a priori* assumed that the typical climate anomaly of one ENSO extreme is likely to be the opposite of the other extreme. The most notable climate anomalies during El Niño events include an eastward shift in tropical precipitation over the equatorial Pacific associated with the perturbed Walker circulation, anomalous descending motion over tropical South America and the tropical Atlantic; enhanced ascending motion over the equatorial western Indian Ocean and East Africa; a southward displacement of the Intertropical Convergence Zone (ITCZ) over the tropical eastern Pacific; a northeastward shift of the South Pacific Convergence Zone (SPCZ) and a weakened Asian monsoon, although the relationship between the Asian monsoon and ENSO has weakened substantially over the past decades (Kumar et al., 1999, 2006). El Niño-induced changes in the atmospheric circulation over the North Atlantic region, resembling the negative phase of the North Atlantic Oscillation (NAO), have also been noted (e.g., Emile-Geay et al., 2007; Toniazzo and Scaife, 2006).

The enhanced convection over the equatorial Pacific during El Niño events leads to a strengthening of the upper tropospheric westerly winds over the Caribbean and equatorial Atlantic regions. As a result, tropospheric vertical wind shear is enhanced and the upper-level circulation less anticyclonic, both of which are unfavorable for tropical cyclone genesis and maintenance. Hurricane frequency over the tropical North Atlantic and the Caribbean therefore tends to be reduced during El Niño events (e.g., Smith et al., 2007; Tartaglione et al., 2003). The relationship between ENSO and tropical cyclones (typhoons) over the western Pacific and the

South China Sea on the other hand is more complicated, with the eastern and western part of the basin responding differently to ENSO conditions and both showing a strong seasonal dependence in their response (e.g., Camargo and Sobel, 2005; Chan, 2000).

Changes in temperature and precipitation patterns during ENSO can lead to thermal and moisture stress which affects agricultural crop production and vegetation in general. As a result of these changes in atmospheric circulation described previously, drought tends to prevail over many tropical land areas during El Niño, while La Niña events tend to produce excessive precipitation over tropical land (Lyon and Barnston, 2005). There are also reports of a heightened risk of certain vector-borne diseases such as malaria in regions where ENSO impacts are strong and disease control is limited (e.g., Pascual et al., 2000). More accurate seasonal climate forecasts, several months in advance, could provide early indicators of epidemic risk, particularly for malaria or cholera.

The profound impacts of ENSO, however, are not only felt on land but also in the Pacific Ocean as warm SSTs and reduced upwelling of nutrient-rich water during El Niño lead to reductions in chlorophyll concentrations (Chavez et al., 1999) and mass displacement of fish stock such as skipjack tuna (Lehodey et al., 1997). It is estimated that global economic losses during El Niño reached US\$ 8–18 billion in 1982/83 and US\$ 35–45 billion in 1997/98 (Goddard and Diley, 2005). On the other hand, stronger ENSO events also lead to greater predictability of climate variations. Climate forecasts therefore tend to be more accurate during ENSO, which can help mitigate adverse impacts and reduce cost to life and property in years of ENSO extremes (Goddard and Diley, 2005).

Indeed, significant efforts have been directed toward improving ENSO prediction. During the 1960s and 1970s monitoring of the tropical Pacific was very erratic and El Niños could emerge essentially without warning. The event in 1986/1987 was the

first to be successfully forecast, but it was followed by new failures. Even in 1997/1998 alerts were issued very late and most models severely underestimated the event's enormous magnitude (e.g., Barnston et al., 1999). In 2004 a new study suggested that large ENSO events might be predictable up to two years in advance (Chen et al., 2004), but it was based on a successful hindcast rather than forecast of an event. Models currently in operation are either based on statistical schemes, using empirical linear techniques, or are physically based, consisting of coupled ocean-atmosphere models of varying degrees of complexity. The limits of El Niño predictability are still the subject of much debate. Some scientists argue that predictability is largely limited by the effects of high-frequency atmospheric 'noise' (westerly wind bursts associated with the MJO), which makes ENSO essentially unpredictable, while others believe that limitations arise from the growth of initial errors in model simulations and that predictability will improve once models become more sophisticated.

### 2.3 ENSO in the past

Past ENSO variability is documented in a variety of natural and historical archives. Marine observations document anomalous winds and ocean currents affecting travel times of ancient sailing ships off the coast of South America, mass mortality of endemic marine organisms and guano birds, invasions of tropical nekton, rising SST and sea levels and associated effects on coastal fisheries and fish meal production, all of which are commonly ascribed to El Niño events (Quinn et al., 1987). On land, a wealth of archival data documents periods of anomalous amounts of rainfall and flooding along coastal Peru, indicative of El Niño, dating back to Pizarro's travels in 1525. Most studies reviewing such archival material suggest that the seventeenth century appears to be the least active El Niño period, while the 1620s, 1720s, 1810s, and 1870s were the most

active decades (Garcia-Herrera et al., 2008; Ortlieb, 2000; Quinn et al., 1987). One limitation of these studies is that they cannot provide any information on La Niña activity, as precipitation along coastal Peru is insensitive to cold events in the equatorial Pacific.

Coral records from the tropical Pacific are advantageous as they record changes in SST and salinity associated with ENSO *in situ* and because they are equally sensitive to both the ENSO cold and warm phase. Sr/Ca and oxygen isotope records from corals in the tropical Pacific, for example, reflect ENSO-related SSTA and anomalous convective activity through rainfall-induced salinity changes (e.g., Cole, 1993; Evans et al., 1998). Indeed coral-based oxygen-18 records have been successfully used to extend ENSO-indices such as the Niño3 index back in time (e.g., Urban et al., 2000). By dating and analyzing ancient corals, washed onto the top of reefs during storms, stacked records can yield a glimpse of past ENSO variability back hundreds or thousands of years (Cobb et al., 2003; Tudhope et al., 2001). The study by Cobb et al. (2003) suggests that ENSO varied considerably in strength over the past millennium, with changes occurring rapidly, on time scales of decades or so. Compared with earlier periods, events in the twentieth century appear to have been relatively strong, but not exceptionally so.

Tree rings from ENSO sensitive regions are also commonly used to reconstruct the past history of El Niño and La Niña; however, they have the drawback that they implicitly assume that teleconnection patterns have remained stationary through time (e.g., D'Arrigo et al., 2005; Stahle et al., 1998). Several of these proxy-based studies suggest a connection between decreased (increased) radiative forcing and greater (lesser) ENSO variability. Hence the Little Ice Age period appears to have been characterized by enhanced ENSO activity, while opposite conditions prevailed during medieval times.

Longer ENSO reconstructions, spanning the entire Holocene, were first produced from laminated lake sediments, recovered from the Andes of Ecuador (Rodbell et al., 1999). They suggest that ENSO was weak or absent in the early Holocene and became more prominent during the middle and late Holocene (Moy et al., 2002). Ice core records from the Andes similarly offer some potential for reconstructing the history of ENSO, as demonstrated by Thompson et al. (1984) and Bradley et al. (2003).

## 2.4 ENSO and global warming

Several unusual aspects of ENSO behavior in the 1980s and 1990s prompted a flurry of papers debating whether this change had a natural or an anthropogenic origin. There was an increase in frequency of El Niño events; the period from 1990 to 1995 was considered the longest El Niño period on record (this period is not universally accepted as one long event, but it certainly coincides with the longest sustained high pressure in Darwin in the instrumental record) and the event in 1997/1998 is generally considered to be the strongest event of the twentieth century. While some studies concluded that these changes were indeed unprecedented (e.g., Trenberth and Hoar, 1996, 1997), others argued that the changes seen are not that unusual in a historical context (Rajagopalan et al., 1997). In any case the debate prompted renewed interest in how ENSO might change in the future under increased greenhouse gas forcing. While early results suggested enhanced ENSO activity due to greenhouse warming (Timmermann et al., 1999), more recent studies indicate that the simulated response is strongly model dependent and that models with the largest ENSO-like climate change, are those performing the poorest in simulating ENSO variability (Collins, 2005). The most likely scenario, based on a model-skill-weighted approach where models which can simulate modern ENSO conditions reasonably well have more weight than

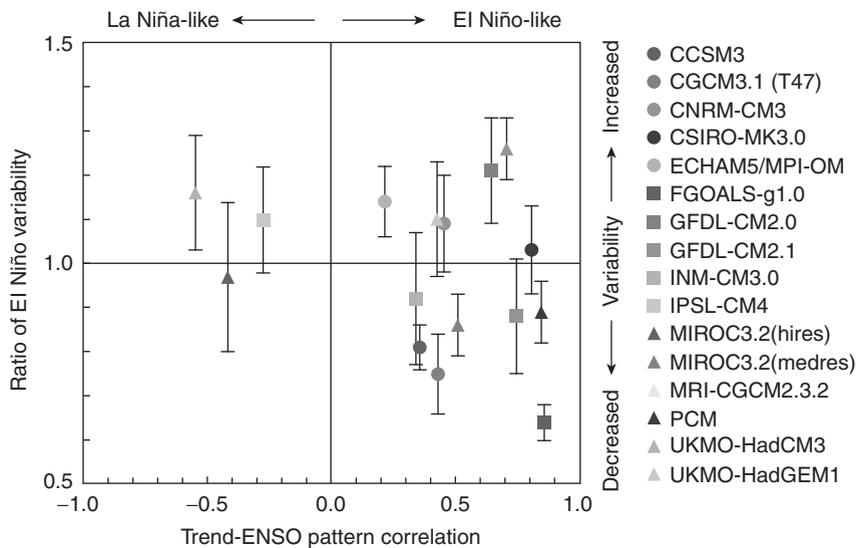
models that fail in this regard, is for no trend towards either mean El Niño-like or La Niña-like conditions (Cane, 2005; Collins, 2005). There is currently no consistent indication of discernible future changes in ENSO amplitude or frequency, even though a majority of models suggests a future Pacific base state with a more El Niño-like pattern (Meehl et al., 2007; Figure 22.4).

### 3 THE PACIFIC DECADAL OSCILLATION (PDO)

#### 3.1 Definitions and indices

Pacific decadal and interdecadal variability has been defined in many different ways with different authors using different indices

over different parts of the Pacific Ocean (Table 22.2). However, these are not necessarily different descriptions of the same phenomenon as Pacific (multi)-decadal variability may indeed consist of more than one mode of variability. Mantua et al. (1997) described a Pacific Decadal Oscillation (PDO) that consists of coherent, interdecadal covariability in the dominant pattern of North Pacific SLP and SST. Hence the PDO can be defined as the leading EOF of SST poleward of 20°N. The PDO shows remarkable multiyear and multidecadal persistence with just two full PDO cycles in the past century: a cool regime from 1890–1924 and again from 1947–1976, and a warm regime from 1925–1946 and from 1977 onward (Figure 22.3). The spatial SST and SLP patterns of the PDO are in many ways reminiscent of the typical ENSO patterns; however, the PDO amplitudes in the



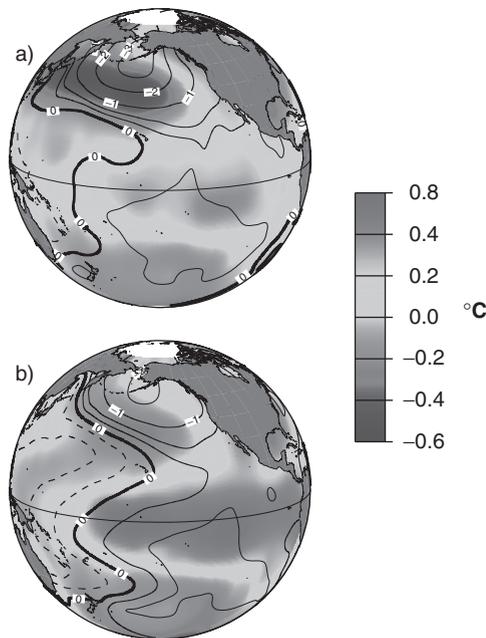
**Figure 22.4** Base state change in average tropical Pacific SSTs and change in El Niño variability simulated by AOGCMs in 1 per cent/year CO<sub>2</sub> increase climate change experiment. Positive correlation values (horizontal axis) indicate that the mean climate change has an El Niño-like pattern, and negative values are La Niña-like. Note that tropical Pacific base state climate changes with either El Niño-like or La Niña-like patterns are not permanent El Niño or La Niña events, and all still have ENSO interannual variability superimposed on that new average climate state in a future warmer climate. Reproduced from Meehl et al. (2007) with permission from the Intergovernmental Panel on Climate Change.

**Table 22.2 Definition of main climate modes other than ENSO**

<i>Index</i>	<i>Definition</i>	<i>References</i>
Pacific Decadal Oscillation (PDO)	1 <sup>st</sup> EOF of Pacific SSTA north of 20°N	Mantua et al. (1997)
ENSO-like interdecadal variability/ Interdecadal Pacific Oscillation (IPO)	1 <sup>st</sup> EOF of low frequency component of N. Pacific SSTA 1 <sup>st</sup> EOF of residual time series (ENSO removed) of N. and S. Pacific SSTA	Zhang et al. (1997) Garreaud and Battisti (1999)
North Atlantic Oscillation (NAO)	SLP difference between P. Delgada, Azores and Reykyavik, Iceland SLP difference between Lisbon, Portugal and Stykkisholmur, Iceland SLP difference between Gibraltar and Reykyavik, Iceland SLP difference from gridded data: [65°N–60°N, 20°W–15°W] – [40°N–35°N, 30°W–25°W] 1 <sup>st</sup> EOF of SLP over box [80°–20°N, 90°W–40°E]	Rogers (1984) Hurrell (1995) Jones et al. (1997) Luterbacher et al. (1999) Hurrell and van Loon (1997)
Northern Annular Mode (NAM)/Arctic Oscillation (AO)	1 <sup>st</sup> EOF of wintertime (NDJFMA) SLP north of 20°N	Thompson and Wallace (1998)
Southern Annular Mode (SAM)/Antarctic Oscillation (AAO)	1 <sup>st</sup> EOF of 850 hPa geopotential height south of 20°S Normalized, zonal mean SLP difference between 40°S and 65°S	Thompson and Wallace (2000) Gong and Wang (1999)
Tropical Atlantic Variability (TAV)	1 <sup>st</sup> EOF of tropical Atlantic SST (30°N–30°S) SSTA difference between tropical N. and S. Atlantic: [24°N–0°, 50°W–20°W] – [0°–24°S, 40°W–10°W] SSTA difference between tropical N. and S. Atlantic: [22°N–6°N, 80°W–15°W] – [2°N–22°S, 35°W–10°E]	Hastenrath (1990) Hastenrath and Greischar (1993) Enfield (1996)
Indian Ocean Dipole (IOD)/Indian Ocean Zonal Mode (IOZM) Zonal Wind Index (ZWI)	SSTA difference between western and eastern Indian Ocean: [10°N–10°S, 50°E–70°E] – [0°–10°S, 90°E–110°E] Surface zonal wind component averaged over box: [4°N–4°S, 60°E–90°E] 850 hPa zonal wind component averaged over box: [5°N–5°S, 60°E–90°E]	Saji et al. (1999); Webster et al. (1999) Hastenrath (2000) Vuille et al. (2005)

tropics are weaker than for ENSO but stronger in the northern hemisphere extratropics (e.g., Mantua et al., 1997; Figure 22.5). Zhang et al (1997) used a different approach to define the PDO by decomposing Pacific SST variability into a high-frequency component ascribed to ENSO and a low-frequency component associated with the PDO. Subtracting the high-frequency component from the Pacific SST field and considering the leading EOF from the residual variance also yields a PDO-like signal (Zhang et al., 1997). While the PDO focuses primarily on the North Pacific region, subsequent analyses

showed that interdecadal variability exhibits similar spatial signatures in SST, SLP and low-level winds over the Pacific Ocean in the southern hemisphere (Garreaud and Battisti, 1999). The Pan-Pacific nature of this hemispheric mode commonly referred to as ENSO-like decadal variability or Interdecadal Pacific Oscillation (IPO) suggests that it is rooted in the tropics (Garreaud and Battisti, 1999). Unlike the PDO this mode is best expressed in the tropical Pacific, largely symmetric about the equator with expressions in both hemispheres, and appears to be dominated by decadal periods of



**Figure 22.5** SST (color shaded) and SLP (contoured) regressed upon (a) the PDO index and (b) the CTI for the period of record 1900–1992. Contour interval is 1 mb, with additional contours drawn for  $\pm 0.25$  and 0.50 mb. Positive (negative) contours are dashed (solid). Reproduced from Mantua et al. (1997) with permission from the American Meteorological Society.

20–30 years. The PDO on the other hand tends to have a strong component varying at multidecadal periods of 50–70 years and is best expressed in the North Pacific north of 20°N.

### 3.2 Theory and mechanisms

The origins of Pacific decadal variability are unresolved and a topic of ongoing research and debate. Some theories favor the importance of stochastic atmospheric forcing and argue that the PDO is essentially a reddened response to both atmospheric noise and ENSO, resulting in more decadal variability than either one of the two forcings produce

on their own. Other hypotheses stress the importance of atmospheric teleconnections to the North Pacific (e.g., Graham, 1994; Trenberth, 1990), mid-latitude ocean–atmosphere interactions (e.g., Latif and Barnett 1994, 1996), tropical–extratropical interactions (e.g., Gu and Philander, 1997) or oceanic teleconnections to mid-latitudes (e.g., Wu et al., 2003). It has also been suggested that the PDO may not be a dynamical mode *sensu stricto*, but may arise from the superposition of SSTA with different dynamical origins. In this scenario the PDO is not governed by a single physical process that defines a climate mode, akin to ENSO, but results from several different processes.

### 3.3 PDO impacts

The most notable impacts of the PDO are found over North America and in particular over Alaska and northwestern Canada, where both temperature and precipitation strongly depend on the phase of the PDO (Mantua et al., 1997; Minobe 1997). During the positive phase the Aleutian low off the coast of Alaska is significantly deepened and the associated enhanced cyclonic flow of warm, moist air results in anomalously high precipitation along coastal and central Alaska and above-average temperatures over Alaska, Canada and the northwestern US. The PDO can also be identified in streamflow data from the west coast of North America with river runoff varying out-of-phase between Alaska and the northwestern US (Mantua et al., 1997). Most notably, salmon catch is also out-of-phase between Alaska and the northwestern US as the productivity of zoo- and phytoplankton related to upper-ocean changes in mixed-layer depth and temperature are driven by PDO the (Mantua et al., 1997). The mid-1970s phase change is also considered a major driver of anchovy decline and sardine rise off the coast of Peru, with major repercussions for the Peruvian fishing industry (e.g., Chavez et al., 2003).

### 3.4 PDO reconstructions

Given the lack of long meteorological records from the Pacific and the western Americas, proxy-based reconstructions are a valuable tool for extending the PDO back in time. Tree-ring chronologies from the US and Canada have been used to identify regime shifts between cold and warm Pacific states, similar to the one observed in 1976/1977, dating back several centuries (Biondi et al., 2001; D'Arrigo et al., 2001; Minobe, 1997) and to reconstruct the PDO for the past millennium (MacDonald and Case, 2005). These reconstructions suggest that the Pacific exhibited significant power in the 50–70 year frequency band throughout most of the past millennium with some weakening of the oscillation in AD 1200–1300 and AD 1500–1800 and that the PDO resided in its cool phase during much of the Medieval from approximately AD 1000–1300.

Corals are the other commonly used proxy for reconstructing Pacific decadal variability. Several reconstructions stem from the southern hemisphere SPCZ region and their signature is generally coherent with reconstructions from northern hemisphere sites, providing a powerful argument that tropical forcing may be an important factor for decadal variability in the Pacific Ocean (Linsley et al., 2000).

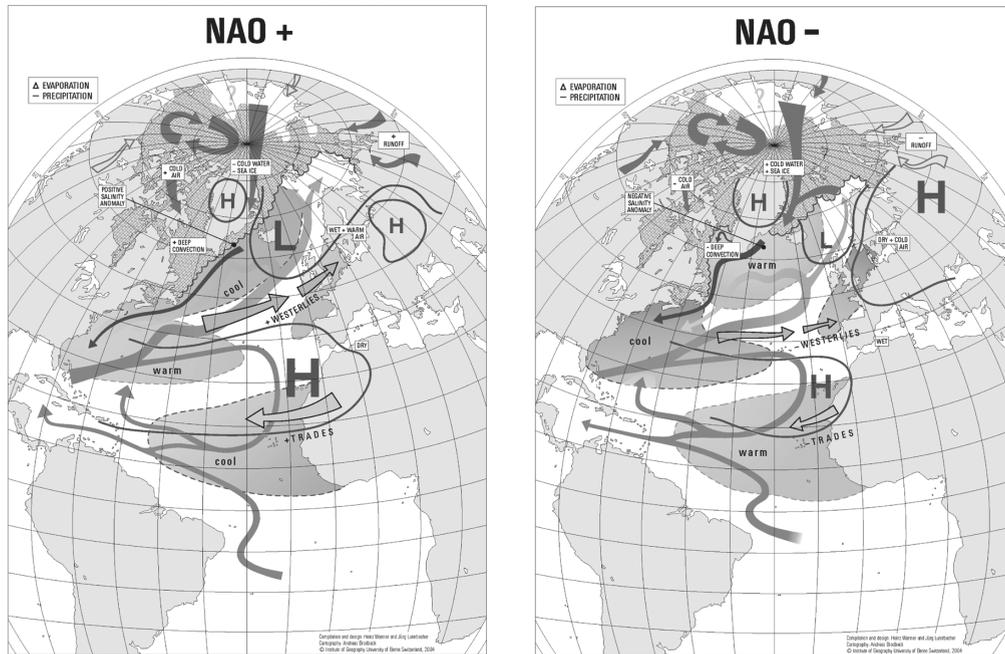
A comparison of the different PDV reconstructions yields correlations significant at 95 percent, yet the common variance is rather low (MacDonald and Case, 2005). The reconstructed major step changes associated with regime shifts in PDV, however, appear to be robust and accurately identified in most PDV reconstructions. In summary, proxy-based reconstructions suggest that PDV does not appear to be a twentieth century phenomenon but has operated in similar ways over (at least) the past several centuries and that the 1976/1977 regime shift is not extraordinary in the longer-term perspective. Proxy reconstructions also support the notion that Pacific (multi)-decadal variability appears to have two preferred time scales of variability: at ~20 years and at ~50–70 years.

## 4 THE NORTH ATLANTIC OSCILLATION (NAO)

### 4.1 NAO dynamics

The NAO was first identified in a series of studies by Sir Gilbert Walker, when searching for global predictors for Indian monsoon rainfall (Walker and Bliss, 1932). Walker noticed that time series of winter SLP and surface air temperature (SAT) at widely dispersed stations in eastern North America and Europe were strongly correlated with each other. He hypothesized that these strong correlations are a reflection of a preferred mode of planetary-scale fluctuations that he referred to as the NAO. Today the NAO is commonly defined as an oscillation of atmospheric mass between the Arctic and the subtropical Atlantic (Hurrell, 1995). It is arguably the most important mode of climate variability over the North Atlantic Ocean and affects weather and climate over eastern North America, the Atlantic and Europe. The NAO is usually defined by the gradient in SLP between stations in Iceland and the Azores, Portugal or Gibraltar, but EOF-based definitions or definitions based on gridded SLP data averaged over the Icelandic low and Azores high pressure regions are also common (Table 22.2). When SLP is lower than normal near Iceland, it tends to be higher than normal near the Azores and vice versa. During the positive phase of the NAO the meridional pressure gradient over the North Atlantic is large because the Icelandic low-pressure center and the high-pressure center at the Azores are both enhanced, while during the negative NAO phase both centers are weakened (Figure 22.6). The most extreme case of a negative NAO occurs when the NAO actually reverses sign; that is, the actual SLP is higher near Iceland than near the Azores.

The westerly winds over the North Atlantic, in geostrophic balance with the meridional pressure gradient, are enhanced during the positive NAO phase, whereas when the NAO is low, the westerlies are weaker than normal.



**Figure 22.6** The two states of the NAO. Surfaces mark SSTs and sea-ice extension, arrows show ocean currents, atmospheric circulation and rivers, blue and red lines indicate near-surface SLP and white rectangles describe characteristic climate conditions or important processes. Reproduced from Wanner et al. (2001), with permission from Springer Science and Business Media.

There are some striking spatial asymmetries between the positive and negative NAO phase. The two pressure centers are displaced eastward by  $\sim 30^\circ$  longitude in the positive relative to the negative regime and the northern center shows northeastward extension of SLP anomalies during positive NAO regime months (Hurrell et al., 2003).

The NAO index is usually calculated for winter months only, the so-called active season, as impacts are most pronounced during winter. Unlike ENSO the NAO has a continuum of possible states rather than a finite set of regimes and bimodality in the different NAOI series is not determinable (Hurrell et al., 2003). The NAO is characterized by a large amount of within-season variance in the atmospheric circulation of the North Atlantic. Most winters are not dominated by any particular regime; rather, the

atmospheric circulation anomalies in one month might resemble the positive index phase of the NAO, while in another month they resemble the negative index phase or some other pattern altogether (Hurrell et al., 2003).

Much of the atmospheric circulation variability in the form of the NAO arises from internal, nonlinear dynamics of the extratropical atmosphere. Interactions between the time-mean flow and synoptic-time scale transient eddies give rise to a fundamental time scale for NAO fluctuations of about ten days. Such intrinsic atmospheric variability exhibits fairly little temporal coherence on longer time scales; therefore month-to-month and interannual changes in the phase and amplitude of the NAO are largely unpredictable. It remains debatable, however, to which extent the forced, anomalous extratropical SST field

feeds back to affect the atmosphere, but most evidence seems to suggest that this effect is quite small compared to internal atmospheric variability (Hurrell et al., 2003, 2006). Some studies suggest that the variability of the NAO is merely the response to chaotic high-frequency forcing by the atmosphere on the surface of the North Atlantic, and therefore fundamentally an atmospheric phenomena. Indeed, the spectrum of the NAO is almost white. The fact that the NAO does not vary on any preferred time scale is consistent with the notion that much of the atmospheric circulation variability in the form of the NAO arises from processes internal to the atmosphere. Accordingly the predictive power of the NAO is weak. It is estimated that only ~10 percent of mean winter NAO variance can successfully be predicted one year in advance (Hurrell and van Loon, 1997). Nevertheless, the interaction between the ocean and atmosphere could be important for understanding the details of the observed amplitude of the NAO and its longer-term temporal evolution and a better understanding of North Atlantic ocean–atmosphere interactions could potentially increase the prospects for meaningful predictability (Hurrell et al., 2003, 2006).

#### **4.2 NAO impacts**

The most significant impacts of the NAO are felt in Europe during winter, associated with the shift in storm tracks over the North Atlantic and the advection of heat by the anomalous mean flow (e.g., Wanner et al., 2001). During the positive NAO phase the more northern jet axis and the increased number of Atlantic storms (Dickson et al., 2000) leads to wet and warm winter climate in Scandinavia, while cool and dry conditions dominate in southern Europe and northern North Africa (e.g., Hurrell, 1995; Hurrell and van Loon, 1997; Wanner et al., 2001). A reversed but weaker temperature impact can be observed along the North American coast with positive anomalies

over the Saragasso Sea and negative anomalies over Greenland and the Labrador Sea. Almost opposite conditions usually prevail during the negative phase of the NAO. The change in location and strength of the storm tracks also significantly affect sea level and sea-ice extent in the Norwegian Sea and the Barent Sea (e.g., Dickson et al., 2000). Economic impacts of the NAO are felt primarily in Europe and include impacts on offshore oil drilling and hydropower production through changes in stream flow and crop yields.

#### **4.3 NAO reconstructions**

The NAO has been the target of several proxy-based reconstructions, using tree-ring information from Europe, North Africa and North America, ice-core data from Greenland, documentary data from Europe and Spanish sailors, speleothems from Europe and ocean model and driftwood records or a combination of multiple proxies (e.g., Appenzeller et al., 1998; Cook et al., 2002; Luterbacher et al., 1999; Trouet et al., 2009). In general the multiproxy reconstructions have yielded more robust results as they avoid the problem of relying on a single site, which raises issues of site sensitivity, signal reproducibility and stationarity of the NAO impacts (e.g., Schmutz et al., 2000). Using a diversified and complementary proxy data set which is sensitive to different climate parameters (e.g., temperature and precipitation) can improve the reconstruction skill (Schmutz et al., 2000). Most reconstructions tend to accurately capture the low-frequency variability but the performance is generally rather poor for individual extremes, especially negative index years. NAO reconstructions are particularly useful to put the unusual trend toward positive polarity of the NAO at the end of the twentieth century in a longer-term perspective. It appears as if the twentieth century NAO variability is somewhat unusual, but that comparable periods of persistent positive NAO phases may have occurred in the past,

especially during the Medieval Climate Anomaly (Trouet et al., 2009).

#### 4.4 Recent trends in the NAO

The NAO exhibited a sustained upward trend from 1960s to the early 1990s (Figure 22.3), causing a flurry of papers about potential anthropogenic impacts on the NAO (see Wanner et al., 2001). This trend also spurred discussion as to whether both the observed changes in precipitation due to the more northerly storm tracks, and increasing temperatures during the second half of the twentieth century in Europe, could be attributed to the unusual behavior of the NAO rather than direct greenhouse gas forcing. However the NAO has trended downward recently while winter-time hemispheric warming over Europe has continued throughout the entire period. Therefore the NAO may have contributed to hemispheric and regional warming for multiyear periods, but the global warming trend over the last 30 years is not related to the NAO.

Multimodel simulations with coupled ocean-atmosphere general circulation models (GCMs) suggest that increased greenhouse gas forcing in the twenty-first century will lead to higher SLP over the Mediterranean and a decrease in SLP over the Arctic Ocean, effectively pushing the NAO into a more positive mode behavior (Kuzmina et al., 2005; Osborn, 2004). While there is general consensus between models on this large-scale pattern, there is little agreement regarding the magnitude of this change and even less concerning the detailed regional-scale changes in winter atmospheric circulation (Osborn, 2004; Stephenson et al., 2006). In general, the detection of changes in the NAO is difficult because indices based on SLP are noisy and not as well suited for climate change detection and attribution studies as air temperature or SST data. Furthermore, it is largely unknown if and how the NAO will respond to changes in the thermohaline circulation, which is predicted to weaken under

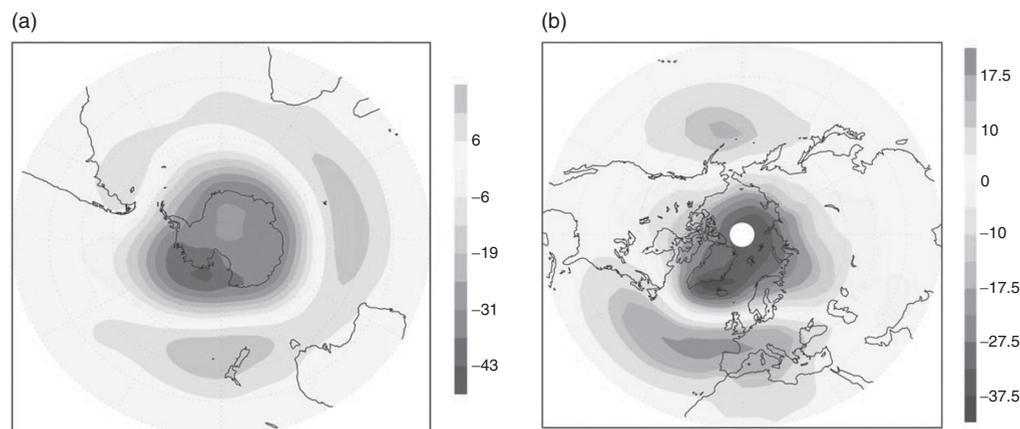
enhanced greenhouse gas emission scenarios (Meehl et al., 2007).

## 5 THE NORTHERN AND SOUTHERN ANNULAR MODE (NAM AND SAM)

### 5.1 The concept of annular modes

In both hemispheres, atmospheric variability at high latitudes is dominated by a seesaw in SLP between the poles and a surrounding zonal ring centered along 45°N and 45°S respectively (Gong and Wang, 1999; Thompson and Wallace, 1998, 2000). These modes are commonly referred to as Northern and Southern Annular Modes (NAM, SAM) or Arctic and Antarctic Oscillations (AO, AAO), respectively, and describe a large-scale oscillation of atmospheric mass between mid and high latitudes. They are generally defined as the first EOF of SLP north and south of 20° latitude respectively (Gong and Wang, 1999; Thompson and Wallace, 1998; Figure 22.7 and Table 22.2). The annular modes are robust patterns that dominate both intraseasonal and interannual variability, but they exhibit their largest amplitude during the northern hemisphere winter (Baldwin and Dunkerton, 1999; Thompson and Wallace, 2000). Owing to the differences in land-sea distribution and orographic effects, the annular shape of the NAM is less well defined than its southern hemisphere counterpart, with the outer ring separated into two separate centers of action over the Euro-Atlantic and Pacific sectors respectively, in which SLPs fluctuate in phase.

In both hemispheres the positive phase is defined as high pressure over mid latitudes and low pressure over the poles, strengthening the westerlies, while opposite conditions prevail during the negative phase. These pressure anomalies are apparent throughout the troposphere and lower stratosphere with a clear upward amplification of the signal. Indeed perturbations of the annular mode typically appear first in the stratosphere and



**Figure 22.7** First EOF of 850 hPa geopotential height anomalies poleward of 20° latitude regressed on SLP in southern hemisphere (a) and 850 hPa geopotential height in the northern hemisphere (b). Data courtesy of Todd Mitchell, JISAO, University of Washington.

then propagate downward into the troposphere (Baldwin and Dunkerton, 1999; Thompson and Solomon, 2002) and only GCMs with a realistic representation of the stratosphere are able to accurately simulate annular mode variability and trends (e.g., Shindell et al., 1999). Hence the annular modes can be interpreted as the surface signature of modulations in the strength of the polar vortex aloft (Thompson and Solomon, 2002; Thompson and Wallace, 1998).

The NAM is closely related to the NAO over the North Atlantic domain and debate continues over which paradigm (NAM or NAO) should be favored and whether the NAO is a truly unique mode or simply a regional expression of the NAM. Historical precedence clearly favors the NAO concept, but Wallace (2000) argued that Walker would probably have recognized the significant involvement of the entire Arctic basin in the mode that he labeled the NAO, had he had access to global SLP data. In this sense some scientists view the NAO as ‘a historical accident’ dictated by the station data availability. Wallace (2000) and Thompson et al. (2000) also suggested that the meridional structure of the NAM (and SAM) is presumed to be determined by processes that transcend the physical geography of any particular

hemisphere. However the NAO paradigm endures by virtue of the undeniable pre-eminence of the Atlantic sector in the SLP signature of this mode and the weakness of the SLP teleconnections between the Atlantic and Pacific sectors (Deser, 2000). Regardless of how it might be understood, the name NAO conveys the notion of a northern hemisphere, Atlantic-centric phenomenon, while the NAM portrays it as a more generic, planetary-scale phenomenon. The alternative name, ‘Arctic Oscillation (AO)’ is an attempt at a compromise that retains the flavor of Walker’s original label, while making more explicit the annular mode’s unique relation to the planetary geometry (Thompson et al., 2000; Wallace, 2000).

## 5.2 Impacts of annular modes

The two extremes of the NAM lead to dramatically different tropospheric circulation and changes in storm track location and surface cyclone activity. During the positive phase high-latitude winds are strong, warm air advection from oceans to the continents is enhanced, winters in northern Canada and Eurasia are unusually warm and northern hemisphere snow cover duration is significantly

reduced (Thompson and Wallace, 2001). Opposite conditions prevail when the NAM is negative and frequent blocking in the mid-tropospheric circulation over both Alaska and the North Atlantic lead to higher frequency of extreme cold events over North America, Europe, Siberia and East Asia, increasing the risk of frost damage and the frequency of occurrence of frozen precipitation (Thompson and Wallace, 2001).

The southern hemisphere counterpart of the NAM has a significant impact on climate of Antarctica and high-latitude southern hemisphere land masses. During the positive phase of the SAM Antarctica is unusually cold, except for the Antarctic Peninsula, where the strengthened westerlies enhance advection of relatively warm oceanic air (e.g., Thompson and Solomon, 2002). Significant impacts of the SAM on temperature, SLP and precipitation patterns are also observed over SE Australia, New Zealand, southern South America and South Africa (Gillett et al., 2006; Garreaud et al., 2009; Silvestri and Vera, 2003).

### **5.3 Recent trends in NAM and SAM**

The NAM has drifted into a positive phase over the past 30 years, peaking around 1990. It has since leveled off, but has remained mostly in its positive polarity (Figure 22.3). SLP is now almost always below normal over the polar region, hence the subpolar westerlies are almost always stronger than normal and high-latitude land areas are unusually warm. Tree-ring based reconstructions of the NAM suggest that the positive trend seen in the twentieth century may be unprecedented at least since 1650 (D'Arrigo et al., 2003); however, this reconstruction did not target the cold season, when the NAM pattern is most dominant. The current positive polarity also has ramifications for Arctic sea ice as advection of ice away from the Siberian coast, enhancing production of thin ice in the East Siberian and Laptev Seas, and export of sea ice through the Fram Strait, are both

enhanced during the positive polarity of the NAM (Rigor et al., 2002). Global warming experiments with GCMs suggest that greenhouse gas forcing will likely continue to maintain the NAM in its positive phase (e.g., Miller et al., 2006).

In the southern hemisphere, the SAM has shown a very similar upward trend, indicative of a strengthened polar vortex, and a drop in geopotential height over Antarctica (Figure 22.3). This trend has been attributed to anthropogenic emissions of ozone-depleting gases (Gillett and Thompson, 2003) as negative perturbations in the geopotential height and temperature over Antarctica, caused by stratospheric ozone depletion, propagate down into the troposphere with a 1–2 month lag. Subsequent studies by Shindell and Schmidt (2004) suggest that ozone changes are indeed the biggest contributor to the observed summertime intensification of the SAM but that increased greenhouse gas concentrations are likely responsible for part of the observed trends, at least in surface variables. Stratospheric ozone losses are expected to stabilize and eventually recover to preindustrial levels over the course of the twenty-first century. However, modeling studies suggest that the continued increase in GHG concentrations will further intensify the polar vortex throughout the twenty-first century and strengthen the westerlies, especially at upper levels (Miller et al., 2006). Whether this will lead to a poleward or an equatorward shift of the storm tracks, however, remains debated (Miller et al., 2006; Son et al., 2008).

## **6 OTHER MODES OF TROPICAL OCEAN-ATMOSPHERE INTERACTIONS**

### **6.1 Tropical Atlantic variability (TAV)**

Climate variability in the tropical Atlantic domain is primarily seasonal and has less

power in the interannual band than its Pacific counterpart, yet it nonetheless has tremendous societal relevance given the large impacts of tropical Atlantic variability (TAV) on precipitation in northeast Brazil and West Africa. The tropical Atlantic is subject to similar Bjerknes-type air–sea interactions and feedbacks as the Pacific (e.g., Zebiak, 1993), yet the amplitude of these Atlantic El Niño events, most common in the summer months June, July and August (JJA), is much smaller and their frequency is higher (Xie and Carton, 2004). Nonetheless changes in SST in the equatorial upwelling region in the Gulf of Guinea are closely linked to onset date and intensity of the West African monsoon (e.g., Folland et al., 1986; Lamb and Pepler, 1992). Equally important for rainfall predictions in the region are changes in the meridional SST gradients during boreal spring (see Table 22.2) which cause anomalous displacements of the ITCZ and can lead to flooding or drought conditions in northeastern Brazil (e.g., Enfield, 1996; Hastenrath, 1990; Hastenrath and Greischar, 1993; Moura and Shukla, 1981).

The tropical North Atlantic is subject to significant variability imposed by the tropical Pacific through ‘atmospheric bridge’ teleconnections (e.g., Curtis and Hastenrath, 1995; Enfield and Mayer, 1997). The warming of the tropical North Atlantic in boreal spring following an El Niño event and the resulting changes in meridional SST gradients significantly affects ITCZ position and trade wind strength and induces further air–sea interactions in the tropical Atlantic (Xie and Carton, 2004). Hence the relative roles of Atlantic variability caused purely by internal ocean–atmosphere interactions and passive variability imposed by SSTA in the tropical Pacific are still a matter of debate and continued research.

## **6.2 The Indian Ocean Dipole (IOD)**

Saji et al. (1999) and Webster et al. (1999) first put forth the idea of an independent

mode of coupled ocean–atmosphere variability in the Indian Ocean, characterized by an east–west dipole in SST with accompanying anomalies in precipitation and low-level winds. While the reality of this mode, and in particular its independence from ENSO, is still a matter of debate (as IOD and ENSO events often coincide and lead to a similar warming over the western Indian Ocean), it is now widely accepted as an important aspect of seasonal to interannual climate variability in the tropical Indian Ocean (e.g., Yamagata et al., 2004). The IOD, sometimes also referred to as Indian Ocean Zonal mode (IOZM), is characterized by an east–west reversal in sign of SST and commonly described as the difference between SSTA in the eastern and western Indian Ocean or by the strength of the equatorial low-level zonal wind (Table 22.2). IOD events are highly phase-locked to the annual cycle with SSTA starting to evolve in June and peaking around October, when zonal winds along the equator are best developed (Saji et al., 1999). Webster et al. (1999) suggested that dipole events initiate when easterly wind anomalies near Sumatra enhance upwelling and evaporative cooling and prevent intrusion of a warm equatorial current which in normal years accumulates warm water along the Sumatra coast. The equatorial easterly wind anomalies favor moisture convergence over the western Indian Ocean and in concert with warm SSTs lead to anomalous convection and precipitation. Observations of sea level and thermocline indicate a coupled oceanic seesaw mechanism (deepening of the thermocline in the west through downward-propagating Rossby waves and becoming more shallow in the east) during IOD events. The positive feedback from the ocean maintains anomalous easterly surface winds and hence can sustain warming for several months (e.g., Webster et al., 1999).

The IOD is of considerable societal relevance given its significant impacts on precipitation in many regions surrounding the Indian Ocean. Positive IOD events are often associated with flooding in East Africa and

drought in Indonesia (e.g., Birkett et al., 1999; Hastenrath, 2007; Vuille et al., 2005).

Fairly little is known about the past history of the IOD as the detection of this mode is still fairly recent. The Indian Ocean has shown a significant warming over the twentieth century, but the IOD shows no such clear trend (Figure 22.3) because it is based on an SST gradient, not absolute temperatures. Coral records from both sides of the Indian Ocean basin have been used to reconstruct IOD variability in the recent and more distant past (Abram et al., 2003, 2007; Kayanne et al., 2006) and the potential for multiproxy reconstructions based on stable isotopic archives has been demonstrated (Vuille et al., 2005).

## 7 CONCLUSIONS

Significant advances have been made in recent years in better understanding coupled ocean-atmosphere interactions and their impacts on climate variability. While the disciplines of meteorology and oceanography have a long tradition studying atmospheric and oceanographic processes, the recognition that the two systems exhibit multiple coupled modes is relatively new. This more integrated approach has helped to discover and better understand coupled ocean-atmosphere modes such as ENSO and their regional- to global-scale impacts. Climate research has also made significant advances in realizing how interconnected these modes are through teleconnections and the concept of 'atmospheric bridges'. Continued improvements in coupled models and data assimilation, however, will be needed to further advance climate predictions and to increase our confidence in model projections of future behavior of climate modes such as ENSO or the NAO in a world with increased greenhouse gas concentrations. As discussed in this chapter, climate variations associated with these modes are of significant societal relevance, yet their relationship with global

warming is often poorly understood. Model deficiencies to some extent reflect the limited understanding of physical processes in both the ocean and the atmosphere (e.g., in clouds). Continued climate monitoring and data analysis combined with improvements to the observational network (on land, ocean and in space) are one avenue for improving our understanding of such processes. In addition, continued efforts are needed to extend reconstructions of these modes back in time by using proxy data in locations that are most sensitive to ocean-atmosphere interactions. This will allow us to put current variations in climate in a longer-term context, which is particularly important given that our observational records often don't extend beyond the past ~150 years and therefore rarely capture a purely naturally forced, preindustrial climate signal.

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