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A Global Survey of Ocean–Atmosphere Interaction and Climate Variability

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The interaction of the ocean and atmosphere plays an important role in shaping the climate and its variations. This chapter reviews the current state of knowledge of air–sea interaction and climate variations over the global ocean. The largest source of climate variability in the instrumental record is El Niño–Southern Oscillation (ENSO), which extends its reach globally through the ability of the atmosphere to bridge ocean basins. The growth of ENSO owes its existence to a positive ocean–atmosphere feedback mechanism (originally envisioned by J. Bjerknes) that involves the interaction of ocean dynamics, atmospheric convection, and winds in the equatorial Pacific. The Bjerknes feedback and the resultant equatorial zonal mode of climate variability are a common feature to all three tropical oceans despite differences in dimension, geometry and mean climate. In addition to this zonal mode, the tropics also support a meridional mode, whose growth is due to a thermodynamic feedback mechanism involving the interaction of the cross–equatorial gradient of properties such as sea surface temperature and displacements of the seasonal intertropical convergence zone. This meridional mode is observed in the tropical Atlantic, with some evidence of its existence in the Pacific and Indian Oceans. In the extratropics, in contrast, the sources of climate variability are more distributed. Much of climate variability may be explained by the presence of white noise due to synoptic weather disturbances whose impact on climate at longer timescales is due to the integrating effect of the ocean's ability to store and release heat. Still, there is some evidence of a more active role for the mid–latitude ocean in climate variability, especially near major ocean currents/fronts. Finally, various atmospheric and oceanic bridges that link different ocean basins are discussed, along with their implications for paleoclimate changes and the current global warming.

Earth's Climate: The Ocean-Atmosphere Interaction
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1. INTRODUCTION

The Earth's climate is determined by many complex physical, chemical and biological interactions among the ocean, atmosphere, land and ice/snow, subject to solar and tectonic forcing. The present volume addresses the key physical interactions of the ocean and atmosphere, which affect climate variability on all timescales. The role of the ocean in climate variability results partly from its large capacity to store and distribute heat. While the ocean is 4000 m deep on average, the upper 10 m has the same mass as the entire atmosphere and the upper 4 m has a similar capacity to store heat. The importance of the ocean for climate is evident in the comparison of diurnal and annual ranges of air temperature over land and ocean. For example, air temperature varies more than 50°C in Beijing within a year but only about 10°C at the same latitude off the coast of northern California.

Solar radiation is the ultimate driving force for all motions in the ocean and atmosphere and gives rise to the pronounced and regular diurnal and seasonal cycles throughout much of the world. Even the minute variations in the intensity and distribution of solar radiation due to orbital changes are implicated in the ice age cycles as well as other features of the Earth's paleoclimate. On the other hand, climate does not just display repeating and regular cycles of solar radiation, but also displays variability that is not correlated with solar radiation. The most famous among such climate variations is the atmosphere's Southern Oscillation [*Walker*, 1924] and its oceanic counterpart El Niño.

Historically, the term El Niño refers to extended episodes of anomalous warming of the ocean off the coast of Peru, while the term Southern Oscillation refers to a sea level pressure (SLP) swing between Darwin, Australia and the island of Tahiti in the central tropical Pacific. *Bjerknes* [1969] recognized that El Niño and the Southern Oscillation (ENSO) are in fact just two different aspects of the same phenomenon, and demonstrated a remarkable correlation between Darwin atmospheric pressure and water temperature off Peru, two locations separated by the vast span of the Pacific Ocean. He further realized that ocean–atmosphere interaction is at the heart of the ENSO phenomenon, and described how an initial change in the ocean could affect the atmosphere in such a manner that the altered atmospheric conditions would in turn induce oceanic changes that reinforce the initial change. For example, if at some initial time sea surface temperatures (SSTs) in the equatorial eastern Pacific are anomalously warm, then the east–west gradient in SST will be reduced. The atmosphere will respond by reducing the east–west gradient of SLP, and consequently relaxing the strength of the easterly trade winds. The relaxation of the easterly winds in turn causes an eastward surge of warm water along the equator, positively reinforcing the initial warm SST anomalies. Thus, this positive ocean–atmosphere feedback of Bjerknes amplifies small initial perturbations into large observable amplitudes.

A further step forward was taken as a result of the intense warm episode of the 1982–83 El Niño, which was not recognized until it was well developed, surprising the oceanographic and meteorological communities. Soon after this surprise, the ten–year (1985–94) international TOGA (Tropical Ocean and Global Atmosphere) Program was launched building on earlier efforts such as the Equatorial Pacific Ocean Climate Studies (EPOCS) Program. One of TOGA's goals was "to study the feasibility of modeling the coupled ocean–atmosphere system for the purpose of predicting its variations on timescales of months to years". The 10–year TOGA program greatly advanced the understanding, simulation, and prediction of the coupled system, as summarized in a series of review articles published in the special volume of the *Journal of Geophysical Research (Oceans)* in June 1998.

A great legacy of TOGA has been the Tropical Atmosphere–Ocean (TAO) and TRITON array [*Hayes et al*., 1991; *McPhaden et al*., 1998], which provides real–time assessments of the thermal structure, currents, and surface meteorology of the tropical Pacific. Beginning in the 1970s and 1980s with outgoing longwave radiation [*Xie and Arkin*, 1996] and SST [*Reynolds and Smith*, 1994], satellite–based observations have played an important role in air–sea interaction and climate research. Since the 1990s, several new space–borne microwave sensors have allowed all–weather observations of SST, rainfall, surface wind and sea surface height over the global ocean. These instruments collectively provide an unprecedented level of detail over the ocean, which has initiated a wide array of air–sea interaction research as reviewed by *Xie* [2004a]. In addition to modern instruments and data archival, the painstaking compilation of global datasets based on historical ship observations [e.g., *Levitus*, 1982; *Woodruff et al*., 1987], and global products using dynamical models to assimilate observations [*Kalnay et al*., 1996; *Gibson et al*., 1997; *Carton et al*., 2000] have all aided the rapid progress in describing, understanding and simulating the climate and its variations.

In 1969, the year of publication of Bjerknes' seminal work, Manabe and Bryan published the results from the first coupled ocean–atmosphere general circulation model. While using a simple sector configuration partitioned between land and sea and a coarse numerical resolution, they demonstrated that such a coupled model could produce a climate not far away from observations. Since then, the coupled ocean–atmosphere models have become an important tool for understanding the climate system and predicting its changes, with ever increasing sophistication and realism [*Meehl*, 1992]. Today there are a multitude of climate models developed independently and they sometimes display such diverse behavior that model intercomparison [e.g., *Mechoso et al*., 1995; *Davey et al*., 2002] is regularly conducted to determine the causes of their differences.

The years since 1969 have seen a succession of conceptual advances. Notable among these are the discovery of mechanisms for the full cycle of ENSO, the mechanisms behind the strong annual cycle in the tropics [e.g., *Mitchell and Wallace*, 1992; *Xie*, 1994], and the identification of thermodynamic exchanges as an important ocean–atmosphere feedback mechanism. While in the Bjerknes feedback mechanism the role of regulating SST is played by ocean dynamics, recent studies have identified important additional roles for surface heat flux. A positive feedback due to interaction of surface wind, evaporation and SST has been proposed for the northward displacement of the intertropical convergence zone over the Atlantic and eastern Pacific [*Xie and Philander*, 1994] and the meridional gradient mode of tropical Atlantic variability [*Chang et al*., 1997]. *Philander et al.* [1996] proposed a positive feedback between SST and low–level stratus clouds that shield incoming solar radiation over the cold water region of the southeast tropical Pacific. More recently, *Wang and Enfield* [2003] suggested a positive feedback between SST and downward longwave radiation of deep convective clouds over the Western Hemisphere warm pool.

In the extratropics, the atmosphere is stably stratified and the direct effect of SST changes is limited to a shallow atmospheric boundary layer. This limitation, along with high levels of weather noise, has made it difficult to make a robust identification of mechanisms by which the ocean can induce positive feedback in the coupled system (except near narrow ocean fronts). Indeed, it now appears that much of mid–latitude SST variability can be explained by the null hypothesis of *Hasselmann* [1976], in which the ocean mixed layer integrates white weather noise in time to yield a red spectrum without much feedback to the atmosphere. On the other hand, ocean–atmosphere interaction in the extratropics has been argued to play a role for mid–latitude climate variability, although it may be relatively weak compared to that in the tropics [e.g., *Latif and Barnett*, 1996; *Kushnir et al*., 2002; *Czaja et al*., 2003].

This chapter provides an overview of global climate vari ations, and is a general introduction to the subjects discussed in the papers that follow in this volume. Sections 2 and 3 review our understanding of ocean–atmosphere interaction in the tropics and extratropics, respectively. Section 4 considers interaction among different ocean basins and between the tropics and extratropics. Finally, Section 5 discusses remaining issues and future challenges.

2. TROPICAL CLIMATE VARIABILITY

2.1. Mean State

Easterly trade winds prevail in tropical oceans. This prevalence of the easterlies is the surface wind response to deep atmospheric convection and heavy rainfall taking place in the intertropical convergence zone (ITCZ). The ITCZ is readily recognizable as a bright band of clouds in satellite images. Condensational heating in the ITCZ lowers the local SLP and causes surface wind to converge onto it. The Coriolis force acting on these converging meridional winds induces an easterly component, leading to the southeast and northeast trade winds in the Southern and Northern Hemispheres, respectively (Figure 1).

The southeast trades penetrate slightly north of the equator over the eastern Pacific and Atlantic. These easterlies drive surface water away from the equator because the Coriolis parameter changes sign there, forcing cold deep water to upwell. This upwelled water keeps the eastern equatorial Pacific and Atlantic Oceans cold. Along the equator, the easterly wind stresses acting upon the ocean are to first order balanced by a zonal tilt of the thermocline that shoals in the east (Figure 1). The shoaling of the thermocline in the east brings cool thermocline water to the sea surface, reinforcing the cooling effect of the equatorial upwelling. The deepening of the thermocline in the western Pacific and Atlantic, on the other hand, prevents cold thermocline water from being upwelled to the sea surface, keeping SST warm there. This thermal contrast between the warm western and cold eastern ocean establishes a westward pressure gradient in the lower atmosphere, enhancing the easterly winds on the equator that act to reinforce the sea surface cooling and shoaling thermocline in the eastern ocean. This ocean–atmospheric interaction, in maintaining climatological mean states, is similar to the positive feedback mechanism *Bjerknes* [1969] proposed for ENSO [*Neelin and Dijkstra*, 1995; *Sun and Liu*, 1996].

The tongue of cold SSTs prevents deep atmospheric convection from forming on the equator. This cold tongue displaces atmospheric convection on both sides of the equator in the western Pacific. In the eastern Pacific and Atlantic, by contrast, the atmospheric ITCZ is peculiarly displaced north of the equator. As well as its role in developing the east–west contrast, air–sea interaction also plays a key role in maintaining this climatic asymmetry in the north–south direction. One consequence of the displaced ITCZ is a local reduction of surface wind speed (Figure 1). Winds are weak in the ITCZ

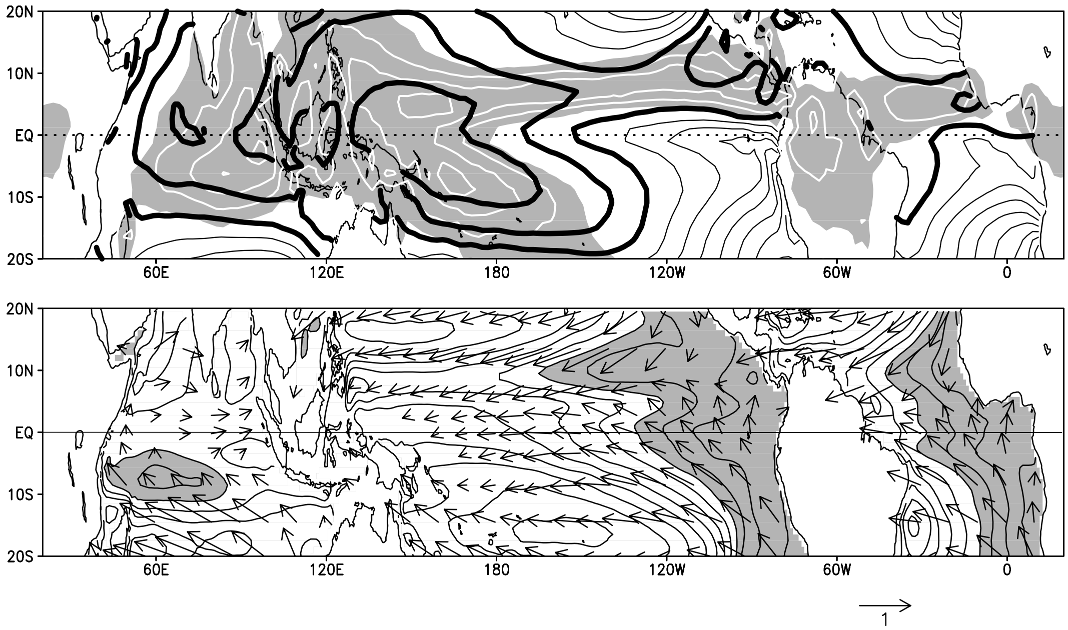

Figure 1. Annual-mean climatology: (upper) SST (black contours at 1°C intervals; contours of SST greater than 27°C thickened) and precipitation (white contours at 2 mm day^{-1}; shade >4 mm day^{-1}), based on the *Reynolds and Smith* [1994] and CMAP [*Xie and Arkin*, 1996] products, respectively; (lower) surface wind stress vectors (N m^{-2}) and the 20°C isothermal depth (contours at 20 m intervals; shade < 100 m), both based on the Simple Ocean Data Assimilation [*Carton et al.*, 2000].

since the Coriolis force, acting upon the southerly cross–equatorial flow, accelerates the easterly trades south and decelerates them north of the equator. In addition to this feedback resulting from the interaction of surface wind, evaporation and SST, the extensive low–level cloud deck that tends to form in the eastern tropical ocean south of the equator is also important in creating the thermal contrast between high SSTs north and low SSTs south of the equator. These positive feedbacks between the ocean and atmosphere help the coupled system to break the equatorial symmetry set by the distribution of annual–mean solar radiation, and to amplify asymmetric perturbations induced by coastline orientation and other aspects of continental geometry. See *Xie* [2004b] for a review of the recent progress in studying this northward–displaced ITCZ.

In the far eastern Pacific and Atlantic where zonal winds are weak, the strong southerly winds displace the upwelling zone south of the equator as indicated by the southward displacement of the meridional minimum of SST. This southerly–induced upwelling helps keep the eastern Pacific and Atlantic cold and, through the Bjerknes feedback mechanism, plays an important role in maintaining the east–west SST gradients and equatorial easterlies. A strong SST front (associated with tropical instability waves) forms north of the equator during boreal summer to fall, with much weaker SST gradients to the south. The southerly winds experience strong modification as they cross this equatorial front, and the associated adjustment has been the focus of a recent observational campaign [*Raymond et al*., 2004; *Small et al*., 2004] which further explores the dynamics associated with this process.

2.2. Seasonal Cycle

Solar radiation is dominated by an annual cycle with a spatial structure that is roughly anti-symmetric about the equator. In response, the seasons in the Northern Hemisphere are opposite to those in the Southern Hemisphere. Over the oceans in the Northern Hemisphere, the seasonal maximum (minimum) in SST generally take place in September (March), three months after the summer (winter) solstice, a lag due to the large thermal inertia of the upper ocean.

This local waxing and waning of solar radiation is a reliable predictor of seasons over most of the world except on the equator. While the annual harmonic of solar radiation reaches a minimum near the equator, a pronounced annual cycle in SST is observed in the eastern Pacific [*Horel*, 1982] and Atlantic, both locations where the climatological–mean thermocline is shallow. For example, the annual range of SST near the Galapagos Islands is about 6°C, greater than most of the tropical oceans (Figure 2). *Mitchell and Wallace* [1992] note that there is a westward co-propagation of seasonal SST and zonal wind anomalies along the equator in the eastern

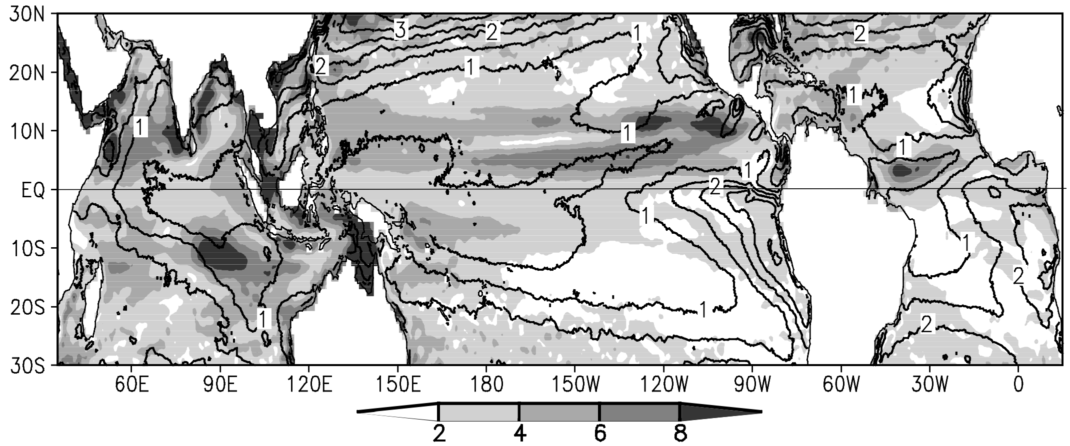

Figure 2. Root-mean-square variance of seasonal variations: SST (contours at 0.5°C intervals) and sea surface height (shade in cm), based on TRMM observations and a combined Topex/POSEIDON and ERS product.

Pacific and suggest that the SST annual cycle results from its interaction with the atmosphere. The northward displacement of the annual ITCZ implies that southerlies prevail in the eastern equatorial Pacific throughout the year, reaching a maximum in September and a minimum in March. A weakening (strengthening) of the cross–equatorial southerlies in response to the anti-symmetric solar forcing acts to decrease (increase) upwelling, vertical mixing, and surface evaporation, all helping to warm (cool) the surface ocean in a cascade of ocean–atmosphere interactions [*Xie*, 1994; *Chang*, 1996; *Nigam and Chao*, 1996].

In the east the annual cycle of equatorial SST is influenced by zonal variations in thermocline depth, but not its temporal variations. In fact, the peak amplitude of the annual cycle of SST is located on or slightly south of the equator while at this latitude the annual harmonic of sea surface height (thermocline depth varies proportionally to sea surface height, with a proportionality factor of roughly 200)—reaches a minimum (Figure 2). The annual harmonic of SST decreases westward along the equator as the thermocline becomes deeper and ocean mixed layer temperature is less sensitive to changes in winds through upwelling, vertical mixing, and/or surface heat flux. Similarly, the annual harmonic of SST reaches a minimum to the north at which latitude the annual harmonic of sea surface height reaches its tropical maximum. Thus, the annual harmonic of thermocline depth variability is largely decoupled from SST, in sharp contrast to their strong coupling on interannual timescales.

The collocation of a variance maximum in sea surface height and a variance minimum in SST along the position of the mean ITCZ is particularly clear over the tropical Atlantic presumably because of weaker interannual variability there than over the Pacific. Unlike the Pacific, there is a marked maximum in sea surface height variance near the equator in the Gulf of Guinea where the thermocline depth shoals by as much as 40 m in July–August relative to March [*Houghton*, 1983]. Thus, thermocline feedback probably plays a more important role in the annual cycle of equatorial SST in the Atlantic than in the Pacific.

Flanked by major continents, the tropical Atlantic is also more strongly influenced by continental monsoons. As an example of this continental influence, *Mitchell and Wallace* [1992] note a strong asymmetry between the seasonal cooling and warming in equatorial Atlantic SST, with the former taking just 3 months and the latter the rest of the year. The abrupt equatorial cooling is initiated by the rapid onset of the West African monsoon and the intense cross–equatorial southerlies [*Okumura and Xie*, 2004].

Among the tropical oceans, the strongest influences of the continental monsoon are found over the tropical North Indian Ocean, where the seasonal winds reverse direction, from southwesterly in summer to northeasterly in winter. This seasonal cycle in wind speed, which peaks twice a year, forces a distinctive semi-annual cycle in SST there. This semi-annual SST cycle is particularly pronounced in the western Arabian Sea and the South China Sea where coastal upwelling induced by the southwest monsoon causes a mid-summer cooling.

Seasonal SST variance reaches a meridional minimum in the Indo–Pacific warm pool region where the thermocline is deep. Equatorial Indian Ocean SST displays a weak annual cycle, which is due to cloud–induced solar forcing [*McPhaden*, 1982]. Twice a year in spring and fall, an intense eastward jet forms in the surface equatorial Indian Ocean. These *Wyrtki* [1973] jets are part of a resonant response of the Indian Ocean to a semi-annual cycle in equatorial zonal winds. *Jensen* [1993] and *Han et al.* [1999] show that the basin mode of the second baroclinic mode is in resonance with semi-annual forcing at the zonal size of the Indian Ocean. *Philander and*

Pacanowski [1986] note a similar resonance at the semi-annual frequency for the tropical Atlantic, where a pronounced semi-annual cycle in thermocline depth is observed in the east.

2.3. Regional Views

This and the next subsections discuss climate variability in tropical oceans from regional and comparative perspectives.

2.3.1. Tropical Pacific. The vast tropical Pacific Ocean hosts ENSO whose influence spans the globe. An extensive literature exists on interactions between the atmosphere and ocean that give rise to ENSO. Several articles review this literature during the past one and a half decades [*Philander*, 1990; *McCreary and Anderson*, 1991; *Battisti and Sarachik*, 1995; *Neelin et al*., 1998]. The latest review of this subject is provided by *Wang and Picaut* [this volume] who emphasize the progress made after the TOGA era.

The starting point for most of the current literature is a consensus that the Bjerknes positive ocean–atmosphere feedback mechanism, involving the interaction of thermocline adjustment, upwelling, SST, atmospheric convection and winds, is central to the development of the interannual swings of ENSO. There also is a consensus that wind–induced mass exchange between the equatorial and off–equatorial oceans is important in the transition phase, thus possibly determining ENSO's interannual timescales. However, the reader will find differing views regarding the mechanisms for this mass exchange, leading to different negative feedback mechanisms. Among the views being promoted are a wave reflection process at the western boundary (delayed oscillator), a discharge/recharge process due to Sverdrup transport (recharge oscillator), an eastward surge that depends on a wind–forced Kelvin wave from the western Pacific (western Pacific oscillator), and a process that relies on zonal redistribution of heat due to anomalous zonal advection (advective–reflective oscillator). See *Wang and Picaut* [this volume] for a detailed discussion of these ENSO oscillator models.

In addition to its interpretation as a self-sustained mode, ENSO may be viewed as a stable mode whose phase is regulated by stochastic forcing such as that provided by the Madden–Julian Oscillation of the tropical atmosphere (see references listed in *Lengaingne et al.* [this volume] and *Wang and Picaut* [this volume]). This view does not necessarily contradict the view of ENSO as a self-sustained oscillation. After an El Niño reaches its mature phase, negative feedbacks, such as those invoked in aforementioned ENSO oscillators, are still required to terminate its growth. The question of whether ENSO is best viewed as a self-sustained oscillation or as a stable mode triggered by random forcing is not settled yet.

Finally, examination of past records of ENSO frequencies and amplitudes makes clear that there is substantial decadal and interdecadal modulation of ENSO. A rich variety of mechanisms have been proposed, including both mechanisms confined to the tropics and mechanisms that rely on tropical–extratropical exchanges and interactions. *Wang and Picaut* [this volume] discuss both the mechanisms for ENSO and rapidly evolving subject of ENSO low–frequency modulation.

2.3.2. Tropical Atlantic. Research on climate variability in the tropical Atlantic sector has expanded tremendously in recent years due to the recognition of scientifically interesting climate phenomena linked to interactions between the ocean and atmosphere as well as the need for improved prediction capabilities. In comparison with the Pacific Ocean, the Atlantic Ocean circulation has a stronger mean northward component at upper and intermediate depths to compensate for the presence of southward transport at deep levels, i.e., a strong meridional overturning circulation. The climate of the tropical Atlantic is primarily seasonal, but also varies on longer timescales ranging through decadal and beyond. Variability of this climate causes massive disruptions of populations as well as changes to the environment. There is now strong evidence that a significant part of this variability is the result of, or is modified by, local air–sea interaction within the tropical Atlantic sector itself [*Carton et al.,* 1996; *Chang et al.,* 1997].

Like the Pacific, the tropical Atlantic is subject to a Bjerknes-type feedback mechanism [*Zebiak,* 1993]. The resulting Atlantic Niños resemble their Pacific counterparts in that they involve the disappearance of the cold tongue of water along the equator (during boreal summer or fall), a surge of warm tropical water eastward and then southward along the southern coast of Africa, an anomalous reversal of direction of the equatorial trade winds, and shifts of atmospheric convection towards the anomalous warm water in the east. However, they are weaker and more frequent than the Pacific El Niños.

While the Atlantic Niño is most pronounced in the eastern half of the basin, striking variability also occurs in the west. The eastern Nordeste region of Brazil lies at the southern edge of the range of latitudes spanned by the seasonal migration of the ITCZ and receives most of its annual rainfall during March and April. Small shifts in the latitude of convection to the north or south during these months leads to droughts or floods in this sensitive region. Furthermore, it has been known since the 1970s that anomalous disturbances in the latitude of the ITCZ result from anomalous changes in the cross–equatorial SST gradient. Frequently these patterns of anomalous SST resemble a "dipole", a term used to refer to this phenomenon. Furthermore, as reviewed in *Xie and Carton* [this volume], perturbations in the latitudinal position of the ITCZ cause anomalous changes in surface fluxes of heat into the ocean that tend to reinforce the original anomaly. Alternative flux–based feedback mechanisms may also involve low clouds, which influence net surface radiation.

Because of the magnitude of ENSO and its close proximity to the Atlantic basin, the phase of ENSO has a direct impact on Amazonian convection, as well as winds and SSTs in the tropical Atlantic, and most particularly in the northern half of the basin [e.g., *Curtis and Hastenrath*, 1995; *Enfield and Mayer*, 1997]. Several atmospheric bridge mechanisms have been suggested, including the effect of ENSO on temperatures throughout the troposphere, through its effect on the anomalous Walker circulation in the Atlantic sector and through more mid-latitude routes. These issues, as well as the ongoing debate regarding the connection between the tropical Atlantic and the northern mid-latitude basin are reviewed in *Xie and Carton* [this volume].

2.3.3. Tropical Indian Ocean. ENSO exerts a strong influence on the tropical Indian Ocean, causing a basin–wide warming following a Pacific El Niño event. The tropical Indian Ocean has in the past been viewed as uninterestingly non-interactive by part of the atmospheric/climate modeling community, and was often modeled as a slab mixed layer. Recent studies, however, paint a picture of a more dynamic Indian Ocean with variability in thermocline depth and ocean currents that alter the transport of heat. The Bjerknes feedback mechanism operates in this basin as well, but only during boreal summer and fall when the equatorial winds are weakly easterly and winds off the coast of Indonesia favor upwelling. When a strong Indian Ocean dipole develops with anomalous easterlies on the equator, the fall eastward Wyrtki jet often disappears as part of a dynamic response that reinforces the cooling in the east [*Saji et al*., 1999]. As a result of this disappearance, the Wyrtki jet is much more variable in fall than in spring.

Unlike the Pacific and Atlantic Oceans, the thermocline is flat and deep on the equator in the Indian Ocean, but is shallow in a dome south of the equator (Figure 1). This thermocline dome results from a Sverdrup–type ocean response to basin–wide positive wind curl between the equatorial westerlies and southeast trades to the south. The shallow thermocline and presence of upwelling allow subsurface variability to affect SST in this thermocline dome. Large–amplitude ocean Rossby waves are excited by the curl resulting from zonal wind anomalies associated with ENSO and the Indian Ocean dipole mode. As these waves propagate westward, they induce large SST anomalies in the South Indian Ocean dome, which in turn induce changes in atmospheric convection and winds [*Xie et al.*, 2002]. This coupling of oceanic Rossby waves with the atmosphere is quite strong because the thermocline dome resides within the meridional band encompassed by the annual migrations of the ITCZ. This off–equatorial thermocline dome and the associated maximum in thermocline feedback are unique to the South Indian Ocean. The fact that these Rossby waves take a few months to cross the basin and affect convection and cyclone development in the west may be exploited for useful climate prediction. Elsewhere in this volume, Yamagata et al. and Annamalai and Murtugudde review Indian Ocean variability.

Another important area in which progress is being made is in understanding the Indian Ocean's considerable effect on atmospheric variability. In the Bay of Bengal large SST fluctuations, with peak–to–peak values exceeding 2°C [*Sengupta et al*., 2001; *Vecchi and Harrison*, 2002], occur in conjunction with the break and active cycle of the Indian summer monsoon. The even greater subseasonal SST anomalies in the summer western Arabian Sea due to the instability of the Somali Current and the Great Whirl induce significant anomalies in atmospheric stability, surface wind speed, and wind curl [*Vecchi et al*., 2004]. The Findlater wind jet (the southwesterly monsoonal wind off the coast of Somalia and Arabia), for example, is found to slow down as it passes over cold filaments in the ocean, a result of stabilization of the near–surface atmosphere and decoupling of the boundary layer from the faster moving winds aloft. The tropical South Indian Ocean is another region of large subseasonal SST variability in boreal winter and spring when the Indian Ocean ITCZ is roughly collocated with this region of open–ocean upwelling [*Saji et al*., 2004].

2.4. A Comparative View

2.4.1. Equatorial zonal mode. The tropical Pacific and Atlantic share many common features in their climatology, including the northward–displaced ITCZ, the prevailing easterly trades, the associated eastward shoaling of the thermocline, and an eastern cold tongue along the equator in the latter half of the year. Not surprisingly, both oceans feature an equatorial zonal mode of interannual variability that tends to be phase locked to the cold season. As shown in Figure 3, the equator of both oceans stands out as a meridional maximum in interannual SST variance (except in the far western Pacific).

Philander et al. [1984] show that the vanishing Coriolis effect near the equator renders the Bjerknes feedback positive, allowing for unstable growth of coupled ocean–atmospheric disturbances when the presence of a shallow thermocline allows tight coupling between thermocline variations and SST. The Bjerknes feedback becomes negative off the equator because of the changes in the phase of SLP, wind stress and ocean upwelling introduced by the effects of earth rotation.

Yet, despite these common features, the zonal mode in the Atlantic (Atlantic Niño) is considerably weaker in amplitude, occurs more frequently, and has a shorter duration when it does occur than the corresponding mode in the Pacific (El

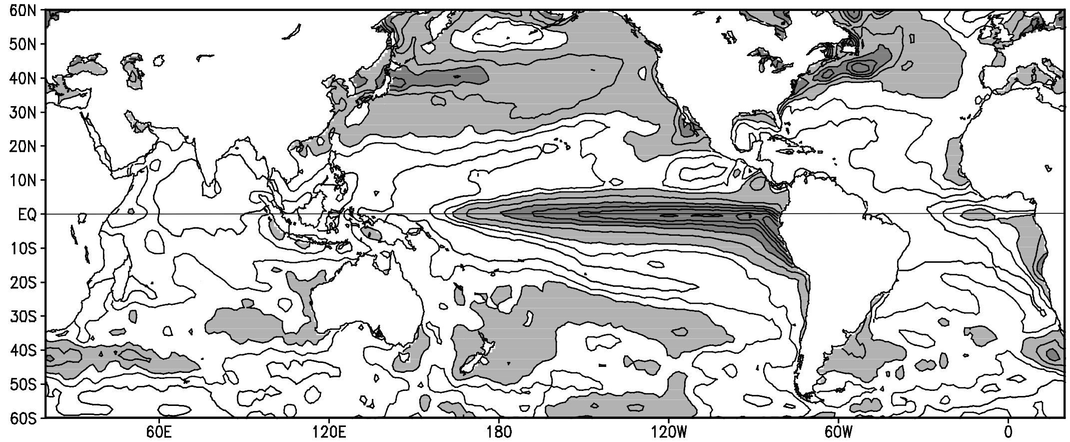

Figure 3. Root-mean-square variance of interannual SST anomalies, based on the Reynolds and Smith [1994] dataset for 1982-2000. Contour intervals are 0.1°C (0.2°C) for values smaller (greater) than 0.6°C. Light (dark) shading denotes values greater than 0.6°C (1.0°C).

Niño). The Atlantic Niño is generally limited to a brief window of June–September when upwelling is normally strong and the thermocline is shallow in the east. The causes of these differences may be found from the results of stability analysis of linear coupled models. These generally give a dispersion relation in which the growth rate of the equatorial mode vanishes at both the long and short wave limits and peaks at a zonal wavelength close to the basin size of the Pacific, much longer than the modest width of the Atlantic Ocean [*Hirst,* 1986]. Drawing much the same conclusion, *Zebiak* [1993] has argued that the Atlantic Niño is in reality the least–damped of a variety of decaying modes, and thus is easy to be excited in response to external forcing.

The annual–mean climate of the equatorial Indian Ocean is very different from that of either the Pacific or Atlantic. In response to the weak westerlies that prevail on annual average, the thermocline is nearly flat along the equator. Because of this deep and flat thermocline and lack of equatorial upwelling (under the mean westerlies), the thermocline feedback mechanism described above is very weak in the equatorial Indian Ocean. This realization of the weakness has led many scientists to model the Indian Ocean as a slab mixed layer that passively responds to remote forcing by ENSO [e.g., *Lau and Nath*, 2000]. Recent studies, as reviewed by Yamagata et al. and Annamalai and Murtugudde in this volume, however, show that during boreal summer and fall, strong anomalies of SST cause east–west SST gradients to occasionally develop and these gradients and the anomalies of convection and zonal winds which may be arranged in such a way as to support positive Bjerknes feedback (akin to that observed in the Pacific and Atlantic).

The eastern Indian Ocean is part of the Indo–Pacific warm pool that hosts a major convection center of the global atmosphere. Because of the intense convection normally occurring here, a unit change in SST induces a large response in atmospheric convection and winds. In fact, the interannual variance of precipitation over the eastern equatorial Indian Ocean is as large as that over the equatorial Pacific despite the much smaller variance of SST [*Saji and Yamagata*, 2003]. In linear coupled models, the growth rate of a coupled mode is dependent on the ocean's response to the atmosphere and the atmosphere's response to the ocean. In the eastern equatorial Pacific and Atlantic, the ocean's response to wind changes, involving thermocline feedback, is very strong because of the prevailing upwelling and shallow thermocline, but the atmosphere's response to SST changes is weak, because the mean SST is low and mean convection is weak. In the equatorial Indian Ocean, in contrast, the atmosphere's response is strong and as a result, a strong growth of an equatorial zonal mode is possible during the seasons when the upwelling favorable winds switch on a thermocline feedback (albeit weak).

Thus, despite large differences in the annual mean winds and thermocline structure, the equatorial mode relying on the Bjerknes feedback mechanism turns out to be a feature common to all the three tropical oceans. The Pacific ENSO displays the largest amplitudes and longest timescales, with significant SST anomalies persisting for a year or more. The Atlantic Niño and the Indian Ocean dipole mode are generally weaker and less regular, with significant SST anomalies limited to a brief window of a few months when the upwelling and thermocline feedback reach their seasonal maximum.

2.4.2. Meridional mode. While thermocline feedback is important in regions of upwelling, Ekman downwelling prevails over the vast off–equatorial and subtropical oceans. In these downwelling regions, subsurface ocean variability is shielded from the sea surface, and other ocean–atmospheric feedback mechanisms involving surface heat flux becomes important. A positive feedback between surface wind, evaporation and SST (WES) favors an anti-symmetric mode that maximizes the cross–equatorial SST gradient [*Chang et al*., 1997]. Unlike the Bjerknes feedback that favors east–west oriented anomalies, this anti-symmetric mode involves air–sea interaction in the north–south direction, with a growth rate that peaks at zonal wavenumber zero and decreases with increasing wavenumber [*Xie et al*., 1999]. In the large Pacific basin, the Bjerknes feedback is strong and so ENSO dominates the climate variability. In the smaller Atlantic basin, the Bjerknes feedback weakens sufficiently that it is comparable to the WES feedback. As a result, the two modes co-exist in this basin without one dominating the other.

Very recently, *Chiang and Vimont* [2004] have carried out a careful analysis of tropical Pacific variability and report a meridional mode analogous to that in the tropical Atlantic. Like the Atlantic meridional mode, which is strongly influenced in the subtropics by local atmospheric variability such as that associated with the North Atlantic Oscillation (NAO), the Pacific meridional mode is linked to the North Pacific teleconnection pattern of the atmosphere. This Pacific meridional mode also has a zonal uniform structure. Further confirmation of the presence of this mode in the Pacific comes from some coupled models [*Yukimoto et al*., 2000; *Okajima et al*., 2003].

The easterly trades prevail on both sides of the equator/ITCZ in the Pacific and Atlantic, a necessary condition for the WES feedback to be positive. This condition is partially met in the South Indian Ocean where the southeast trades prevail year around. While the annual mean winds are weakly southwesterly in the North Indian Ocean, the northeasterlies prevail during boreal winter and spring, meeting the necessary condition for positive WES feedback. During these seasons, *Kawamura et al.* [2001] show that interannual anomalies of SST and surface winds also sometimes organize themselves into equatorial anti-symmetric patterns indicative of the WES feedback. They suggest that this meridional mode may affect the strength of the subsequent summer monsoon over South Asia, an idea that deserves further attention.

3. EXTRATROPICAL CLIMATE VARIABILITY

3.1. Mid-Latitudes

As we move towards mid-latitudes, the difficulty in identifying ocean–atmosphere interactions increases. The difficulty may be due to more complex meteorology in which local anomalies of SST and surface winds are less strongly linked, to oceanic conditions in which SSTs are cooler and mixed layers are deeper, to the longer timescales of oceanic response to atmospheric conditions, as well as to the momentum constraints of the larger Coriolis term. Still, despite this difficulty, there is considerable evidence of interactions. For example, in their analysis of surface air pressure in the North Pacific sector, *Trenberth and Hurrell* [1994] found a long period of elevated surface pressure in midbasin spanning the decades of the 1950s and 1960s, followed in the mid-1970s by a period of reduced surface pressure. The reduction intensified and caused an eastward shift of the Aleutian Low, bringing warm moist air to the west coast of North America and a southward shift of the mid-latitude storm track. See *Mantua and Hare* [2002] and *Miller et al.* [2003] for recent reviews of Pacific decadal variability and its effects on ocean ecosystems.

In his review of the literature, *Latif* [1998] divides mid-latitude interaction theories into three categories: (1) those occurring through interactions in both mid-latitudes and tropics, (2) those involving changes in the gyre circulation, and (3) those occurring in mid-latitudes and involving changes in the thermohaline circulation. Here we examine each of these.

The possibility of interactions of the first type involving both the tropics and mid-latitudes was proposed by *Gu and Philander* [1997]. In their simple conceptual model water with temperature anomalies of either sign is introduced into the oceanic mixed layer in the extratropical Pacific due to local meteorological conditions. This newly formed water is subducted in the thermocline and then follows Lagrangian pathways called subtropical cells (STCs) equatorward and generally westward [e.g., *McCreary and Lu*, 1994; *Liu and Philander*, 2001; *Schott et al*., this volume]. Depending on its geographic origin the water may eventually enter the equatorial thermocline, thus altering the stratification along the equator. According to this conceptual model, changes in stratification lead to changes in SST, causing changes in winds that in turn affect the properties of subducted extratropical water. Changes in the stratification of the equatorial Pacific are also surmised to affect the development of ENSO cycles [*Zebiak and Cane*, 1987; *Neelin et al*., 1994]. Thus, the apparent frequency of El Niño conditions in the 1990s may have resulted from a slow deepening of the thermocline. Some observational evidence to support these theories has been provided by examination of the meridional propagation of temperature anomalies by *Deser et al*. [1996] although the data analysis of *Schneider et al.* [1999] does not find any significant link between the North Pacific and the equator through the subduction of SST anomalies.

Kleeman et al. [1999] proposed an alternative role for STCs in climate, arguing that wind–induced changes in the strength of these shallow overturning cells play a key role in generating equatorial SST anomalies by modulating the amount of cold thermocline water advected into the tropics. This mechanism is supported by a study of *McPhaden and Zhang* [2002] who observed a slowdown of the Pacific STCs from the mid-1970s to the late 1990s associated with a decrease in STC transport. Using an ocean general circulation model (OGCM), *Nonaka et al.* [2002] found that unlike El Niño in which SST anomalies are mostly induced by equatorial winds, off–equatorial winds cause STCs to vary in strength and are thus important for decadal SST variations in the equatorial Pacific. Related studies have examined the pathways of water entering the tropical thermocline in the Atlantic [*Zhang et al.,* 2003; *Schott et al*., this volume]. Current research on the influence of conditions in the tropical Atlantic Ocean on the mid-latitude atmosphere is reviewed in *Xie and Carton* [this volume].

A flurry of research has been provoked by the joint observational and coupled modeling analysis of *Latif and Barnett* [1996] who found a decadal (~25 year) mode of variability involving the Aleutian Low and the subtropical oceanic gyre. They argue along the lines of the second interaction category. According to their mechanism an intensification of the subtropical gyre will transport more warm water into the central North Pacific, which through interaction with the overlying atmosphere will lead to positive feedback and enhanced SSTs through reduction of meridional temperature gradients and enhancement of net heat flux from the atmosphere. It will also sow the seeds of its own demise by reducing the strength of the wind stress curl, thus ultimately reducing the strength of the subtropical gyre on timescales of a decade or so. The importance of oceanic advection in regulating heat storage, a key aspect of this mechanism, is examined by *Kelly and Dong* [this volume] who show that the relationship between meteorological forcing in the North Pacific and Atlantic Oceans results from changes in advection caused by changes in mid-latitude westerly winds.

Fluctuations in the oceanic thermohaline circulation may also play a role in the third category of coupled interactions. As discussed in Section 4.3, the North Atlantic is the only basin with a northern source of deep water. The rate of formation of this North Atlantic Deep Water appears to be connected to the salinity balance of the North Atlantic and Arctic basins [*Mann and Park*, 1996]. This connection apparently is the source of interdecadal cycles in several coupled models [*Delworth et al*., 1993; *Lohmann*, 2003] in which changes in the mid-latitude oceanic gyre influence the properties of water entering the Arctic basin and thus the rate of formation of deep water. Fluctuations in the supply of freshwater to the Arctic has been implicated in the Great Salinity anomalies [*Lazier*, 1988] as well as in decadal variations of wintertime SST and ice in the northwest Atlantic [*Deser and Blackmon*, 1993].

Internal variability of the atmosphere generally increases in amplitude toward the poles, and is well organized in space (but highly disorganized in time) in the form of patterns such as the Pacific–North American (PNA) pattern and the NAO. The null hypothesis of no ocean to atmosphere feedback, originally proposed by *Hasselmann* [1976], suggests that the appearance of low frequency signals in the ocean could simply reflect the ability of the ocean to integrate in time the effects of forcing by the variable atmosphere [*Frankignoul*, 1985]. The Hasselmann null hypothesis has recently been expanded upon by *Deser et al.* [2003] to account for the ability of the ocean to sequester thermal anomalies in the summer, only to reveal them again the next winter when surface cooling and enhanced mixing cause the mixed layer to deepen sufficiently. This reemergence mechanism may be generalized by including ocean heat transport changes that vary subsurface temperature throughout the year [e.g., *Schneider and Miller*, 2001; *Tomita et al*., 2002]. The stochastic model for mid-latitude climate variability has been further developed [*Barsugli and Battisti*, 1998; *Neelin and Weng*, 1999].

High levels of atmospheric internal variability pose a challenge for detecting SST's influence on the atmosphere [*Kushnir et al*., 2002]. On basin scale, SST and wind speed tend to be negatively correlated in the extratropics, especially in winter, a correlation consistent with the Hasselmann null hypothesis. Traditional climatic datasets, however, do not adequately resolve major ocean currents and fronts that are only a few hundred kilometers wide. Yet it is near these regions that ocean currents are expect to play an important role in SST variability. Indeed, recent high–resolution satellite observations reveal robust patterns of air–sea interaction near major ocean currents like the Kuroshio, the Gulf Stream, and circumpolar currents. Surface wind speeds tend to increase over positive SST anomalies, a positive correlation indicative of ocean–to–atmospheric feedback. *Xie* [2004a] and *Chelton et al.* [2004] discuss these new results from satellite observations and their climatic implications. Oceanic fronts along the extensions of major western boundary currents are often regions of extratropical cyclogenesis. *Nakamura et al.* [this volume] discuss possible interactions of oceanic fronts and atmospheric storm tracks.

3.2. High Latitudes

The leading mode of Northern Hemisphere sea level pressure variability based on an empirical orthogonal function analysis features a seesaw between the polar cap and the mid-latitudes, with a corresponding seesaw in zonal–mean zonal wind between the subtropical and subpolar belts. This northern annular mode, or Arctic Oscillation (AO), is most pronounced in boreal winter and is linked to variability in the upper troposphere and stratosphere [*Thompson and Wallace*, 2000]. In the winter stratosphere, the northern annular mode is nearly zonally symmetric and associated with changes in the strength of the polar vortex. Such an annular mode with strong coupling between the surface and stratosphere in local winter is also observed in the Southern Hemisphere where the zonal symmetry is even stronger than the northern annular mode. The northern annular mode exhibits considerable variation in the zonal direction in the troposphere, with the largest loading over the North Atlantic where it resembles the NAO. This similarity gives rise to alternative interpretations of the same phenomena.

In its positive phase, SLP associated with the northern annular mode drops over the polar cap, and increases over the Azores high in the North Atlantic and the Aleutian low in the North Pacific. More storms reach northern Europe, resulting in wetter conditions; surface air temperature increases over the mid- and high-latitude Eurasia and most of North America and dryer conditions and air temperature decreases over northeastern Canada. SST anomalies generally follow a pattern of negative (positive) values where prevailing winds intensify (weaken). The meteorological conditions associated with the positive phase of the northern annular mode feature positive SST anomalies in the mid-latitude North Pacific as the surface westerly jet weakens and a tripole SST pattern in the North Atlantic (with negative SST anomalies in the subpolar region as the surface westerlies intensify). The lead/lag relationship of this index of meteorological conditions to anomalous SSTs in the North Atlantic suggests the possibility that this meteorological phenomenon is significantly influenced by exchanges with the underlying ocean [*Czaja and Frankignoul*, 2002]. However, as recently reviewed by *Czaja et al.* [2003], the role of the ocean–atmosphere coupling in the northern annular mode may be relatively weak. The annular modes seem to be the internal mode of the atmosphere resulting from the interaction of subpolar jets and storm tracks. They appear as the dominant mode of atmospheric variability in atmospheric models even when forced just with climatological SST. The issue of the importance of oceanic feedback remains one of active debate.

There is some evidence that stratospheric anomalies lead tropospheric changes in time and this lead may be exploited for prediction of subseasonal variability in the troposphere [*Baldwin and Dunkerton*, 2001]. Since anthropogenic climate change due to the observed increase in carbon dioxide and the depletion of ozone is likely to be stronger and more robust in the stratosphere, the annular modes may be an important mechanism by which the first sign of anthropogenic climate change in the stratosphere reaches the earth surface. A recent AGU monograph [*Hurrel et al*., 2003] is devoted to studies of the NAO/AO.

Finally, we note that in the Southern Ocean, co-varying ocean–atmosphere–cryosphere signals of 4–5 year timescale have been observed propagating slowly eastward, taking 8–10 years to encircle the pole [*White and Peterson*, 1996; *Jacobs and Mitchell*, 1996]. In this Antarctic circumpolar wave, warm (cold) SST anomalies are associated with poleward (equatorward) meridional surface wind anomalies, suggesting that ocean–atmosphere interaction may be at work. Recently, *White et al.* [2002] suggested this high–latitude wave may influence tropical climate.

4. INTER-BASIN INFLUENCES AND INTERACTIONS

Climate in the tropics and extratropics is intimately linked, and from one ocean basin to another. This section discusses the mechanisms for these linkages, including atmospheric bridges within the tropical belt and to the extratropics, and ocean circulations that link the global oceans together. As discussed in Section 3.2, the northern annular mode may provide a mechanism for inter-basin interaction between the North Pacific and North Atlantic, and north branches of the Antarctic circumpolar wave may provide a linkage between the Southern Ocean and tropical oceans.

4.1. Interactions Among Tropical Oceans

It has been noted that SST anomalies among the tropical Pacific, Atlantic, and Indian Oceans are related to one another [e.g., *Hastenrath et al*., 1987; *Kiladis and Diaz*, 1989; *Tourre and White*, 1995; *Lanzante*, 1996], mainly reflecting the dominant forcing by the Pacific ENSO. The global nature of ENSO is shown in Figure 4, which displays the correlation between the Niño3 SST anomalies during November–December–January (NDJ) and global SST anomalies during the following February–March–April (FMA). These seasons are chosen because ENSO peaks during boreal winter while its influence on other ocean basins normally peaks 1–2 seasons later [e.g., *Alexander et al*., 2002]. Outside the Pacific, significant warming is found over the tropical North Atlantic and the entire tropical Indian Ocean. Figures 5a and 5b compare SST anomalies in the tropical North Atlantic and Indian Ocean with those in the Niño3 region, showing high correlations among these time series. Figure 4 does not reflect the Indian Ocean dipole mode that sometimes co-occurs with ENSO and usually peaks in September–November.

The influence of ENSO on other tropical oceans is transmitted through the "atmospheric bridge" of atmospheric circulation changes. Based on a correlation analysis of satellite and ship observations, *Klein et al.* [1999] provide a schematic of the Walker and Hadley circulations that accompany El Niño events (Figure 6). The adjustment in these circulations is confirmed by the direct circulation analyses of the NCEP-NCAR reanalysis fields [*Wang*, 2002, 2004]. During the warm phase of ENSO, convective activity in the equatorial western Pacific shifts eastward. This shift in convection leads to an altered Walker circulation, with anomalous ascent over the equatorial central and eastern Pacific, and anomalous descent over the equatorial Atlantic and the equatorial Indo–western Pacific region. Thus, the Hadley circulation strengthens over the eastern Pacific but weakens over the Atlantic and Indo–western Pacific sectors. These anomalous Walker and Hadley circulations result in variations in surface wind speed, humidity, and cloud cover that in turn influence surface heat fluxes and SST over the tropical Indian and Atlantic Oceans. Elsewhere in this volume, Kawamura et al. examine long–term changes in the anomalous Walker circulation linking the Indo–Pacific Oceans; Xie and Carton discuss the tropical Atlantic response to ENSO; and Su et al. study a mysterious lack of correlation between tropical mean temperature and precipitation during ENSO.

The tropical Indian Ocean is likely to affect tropical Pacific variability. On intraseasonal timescales, the atmospheric Madden–Julian Oscillation develops over the Indian Ocean and propagates eastward to the Pacific, which may affect the Pacific El Niño [*Takayabu et al.*, 1999; *Lengaigne et al.*, this volume]. On interannual timescales, *Watanabe and Jin* [2003] suggest that El Niño–induced Indian Ocean warming helps amplify the Philippine Sea anomalous anticyclone that forms due to Pacific Ocean–atmosphere processes [*Wang et al.*, 1999; *Wang et al.*, 2000], a result consistent with atmospheric GCM hindcasts [*Lau and Nath*, 2000]. The Indian Ocean dipole mode is associated with strong cold SST anomalies and suppresses convection in the eastern equatorial Indian Ocean, which may affect the Pacific through the attendant surface wind anomalies, a hypothesis that received some support from *Behera and Yamagata*'s [2003] correlation analysis.

Several studies suggest global propagating waves [e.g., *Barnett*, 1985; *Yasunari*, 1985; *White et al.*, 2003]. In particular, *White et al.* [2003] found that these waves are composed primarily of global zonal wave numbers 1 and 2, traveling eastward. Based on uncoupled and coupled GCM simulations, *Latif and Barnett* [1995] concluded that most of the variability in the tropical Atlantic and Indian Oceans associated with the interannual global wave is forced by Pacific SST anomalies via changed atmospheric circulations, with local air–sea interactions acting as an amplifier of the Pacific–induced signal.

4.2. Atmospheric Bridge to the Extratropics

Changes in tropical convection excite planetary waves that bring about climatic anomalies around the world and thus ENSO's influence extends far beyond the tropics [*Alexander et al.*, this volume]. As a result, SST correlations with ENSO exceed 0.6 in the extratropical North and South Pacific (Figure 4). The North Pacific SST anomaly time series in Figure 5c is best correlated with Niño3 SST anomaly at a three–month lag. However, the nature of the atmospheric response changes with latitude. In contrast to the dominant baroclinic structure

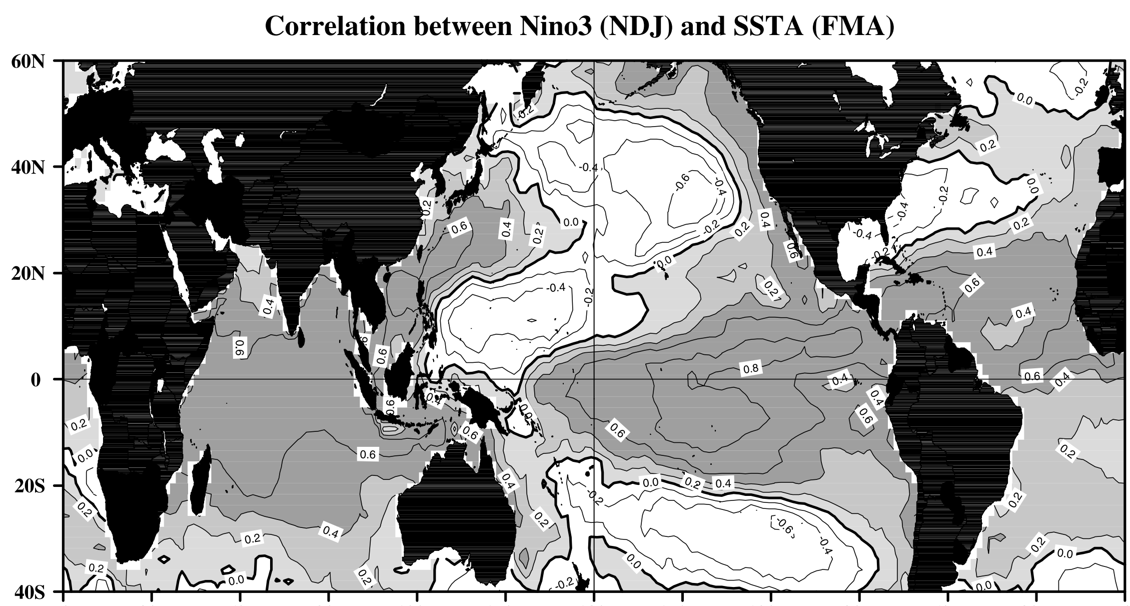

Figure 4. Correlation between the Niño3 (5°S–5°N, 150°W–90°W) SST anomalies during November–December–January (NDJ) and global SST anomalies during the following February–March–April (FMA). The calculation is based on the NCEP SST data from 1950–1999.

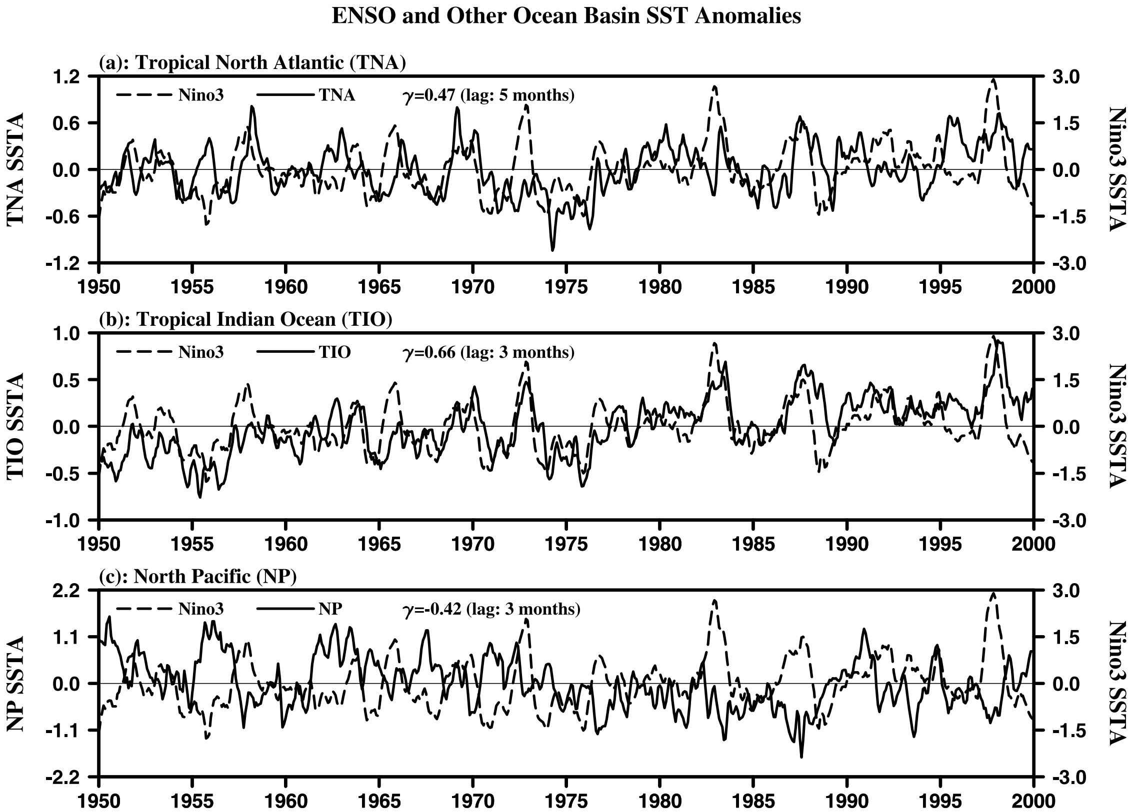

Figure 5. Comparisons of the Niño3 (5°S–5°N, 150°W–90°W) SST anomalies with (a) the SST anomalies in the tropical North Atlantic (5°N–25°N, 55°W–15°W), (b) the SST anomalies in the tropical Indian Ocean (10°S–10°N, 50°E–100°E), and (c) the SST anomalies in the North Pacific (35°N–45°N, 160°E–160°W). All of the time series are three-month running means. The γ represents correlation coefficient.

in the tropics, in the extratropics these waves are barotropic and most pronounced in winter in the Northern Hemisphere [*Trenberth et al*., 1998]. An example of this atmospheric bridge to the extratropics is the PNA pattern of *Wallace and Gutzler* [1981] and *Horel and Wallace* [1981]. This PNA teleconnection pattern is often associated with ENSO in the tropics, with alternating positive and negative geopotential height anomalies that emanate from the tropical Pacific, pass the North Pacific, curve eastward to northwestern America and then equatorward to reach southeastern United States and the Gulf of Mexico (Figure 7). SLP anomalies are generally of the same sign as geopotential anomalies in the middle and upper troposphere, deepening the Aleutian low during an El Niño. The associated surface wind changes lower SST in the central North Pacific (Figure 4) by increasing surface heat flux from the ocean and Ekman advection [*Alexander et al*., 2002]. The SLP and wind changes over the tropical North Atlantic contribute to the warming of the tropical North Atlantic and the Western Hemisphere warm pool [*Enfield and Mayer*, 1997; *Wang and Enfield*, 2003]. *Alexander et al.* [this volume] discuss ENSO's teleconnection in summer, which is much less studied than that in winter.

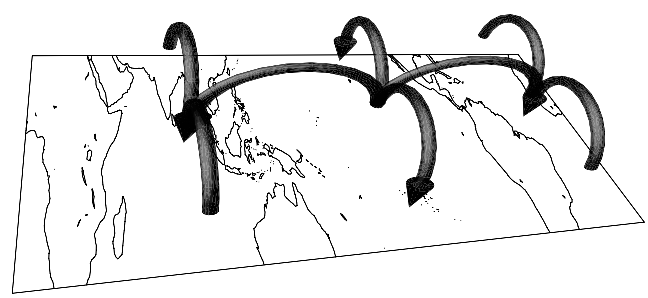

Figure 6. Schematic diagram of the anomalous Walker and Hadley circulation during an El Niño event [*Klein et al.*, 1999].

The atmospheric bridge may also operate the other way around, from the extratropics to the tropics [*Pierce et al.*, 2000]. On decadal timescales the largest anomalies of SST and ocean heat content occur in mid-latitudes instead of the tropics [*Giese and Carton*, 1999]. *Vimont et al.* [2003] sug-

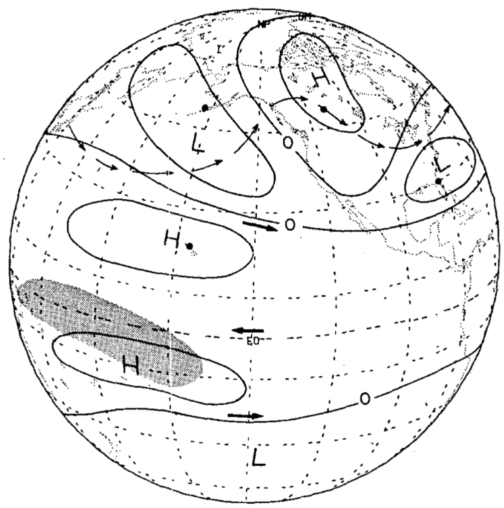

Figure 7. Schematic diagram of middle and upper tropospheric geopotential height anomalies associated with the Pacific North American (PNA) pattern during boreal winter of the Pacific El Niño year [*Horel and Wallace*, 1981].

gest a seasonal footprinting mechanism in which subtropical SST anomalies in winter persist into spring and summer, inducing a broad–scale atmospheric response in summer with significant wind anomalies near the equator that affect ENSO.

Over the North Atlantic, the Azores high in SLP is an important link between the tropics and mid-latitudes, with its variability affecting the northeast trades on one hand and the mid-latitude westerlies on the other. One possibility is that the Azores high is influenced by both the NAO and the Atlantic ITCZ. Indeed, some observational analyses suggest a pan–Atlantic SST pattern that features zonal bands of anomalies of alternating signs spanning from the South Atlantic to Greenland. *Xie and Carton* [this volume] and *Barreiro et al.* [this volume] discuss the interaction between the tropical and extratropical Atlantic.

4.3. Role of Ocean Circulation

Tropical and extratropical connections can also be through oceanic bridges. One of oceanic bridge mechanisms is through an oceanic wave signal transmitted between the mid-latitude and tropical oceans. Coastal Kelvin waves along the west coast of North America have been proposed as the mechanism for linking tropical and extratropical SST anomalies during ENSO and for longer timescales [e.g., *Enfield and Allan*, 1980; *Clarke and Lebedev*, 1999]. *Jacobs et al.* [1994] suggested that the 1982–83 El Niño could have decadal effects on the northwestern Pacific circulation, through mid–latitude Rossby waves reflected from equatorial Kelvin waves on the American coasts. *Lysne et al.* [1997] found a weak decadal signal in their search for another oceanic bridge driven by wave dynamics: anomalous temperature propagated by mid–latitude Rossby waves into the western boundary, then equatorward by coastal Kelvin waves and finally modifying equatorial SST via equatorial Kelvin waves. Recently, observational results of *Hasegawa and Hanawa* [2003] showed anticlockwise propagation of upper–ocean heat content anomalies around the northern tropical Pacific Ocean beginning with an eastward propagation along the equator, northward propagation at the eastern boundary, westward propagation at the subtropics, and finally southward propagation at the western boundary.

As discussed in Section 3.1, shallow overturning ocean circulation may link the tropics and subtropics. Water formed in the subtropics may enter the tropical thermocline within a few years after being subducted. Some of this water returns to the oceanic mixed layer near the equator as the result of entrainment and mixing. Very near the equator quite shallow tropical cells result from the local processes of Ekman divergence and near–equatorial subduction. In Eulerian zonal average these cells appear closed, but in fact involve complex time–dependent pathways.

The thermohaline circulation of the oceans, also known as the meridional overturning circulation, occurs on several vertical levels and at several timescales. Overturning of the lower 2 km of the ocean occurs on long centennial timescales (although the timescales may be much shorter in some regions). This deep and bottom water is formed only at the northern end of the North Atlantic and around Antarctica. Intermediate water masses, which appear at depths of 1 km or so, are formed in several locations. In the Atlantic intermediate water is formed notably by salty water exported from the Mediterranean Sea (Mediterranean water) and also by fresher water formed in the circumpolar current (Antarctic intermediate water).

Changes in the deeper circulation have been implicated in a number of climate anomalies appearing in the geologic record. For example, a shutdown of the deep meridional overturning circulation seems to have occurred at the end of the last ice age due to the release of meltwater from the Laurentide ice sheet some 13–11.5 kyr ago [*Broecker et al.*, 1989; *Ramstorf*, 2000, 2002]. The impact on the climate of high latitudes from this shutdown was prompt, dropping temperatures by more than 5°C. Much more modest freshwater anomalies have appeared at shallow levels in the North Atlantic due to anomalous sea ice export in 1969 and 1994 [*Lazier*, 1988]. A recent study by *Curry et al.* [2003] intriguingly suggests changes in the salinity of surface waters of the Atlantic during the past 30 years (increasing salinity in the tropics and decreases at high latitudes) that one might expect from a slowdown of the meridional overturning circulation.

5. DISCUSSION

For historical reasons the ocean and atmosphere have traditionally been studied separately, assuming that the state of the other fluid is a specified boundary condition. The development of geophysical fluid mechanics that began half a century ago is an attempt to understand the common dynamics shared by the ocean and atmosphere. This development has been a huge success—concepts like quasi-geostrophy, thermal wind, potential vorticity conservation, Kelvin and Rossby waves, and baroclinic instability have helped explain many phenomena in the ocean and atmosphere. The study of climate has triggered a new level of integration of the fields of meteorology and oceanography, demanding that we treat climate phenomenon as a coupled problem. This paradigm of ocean–atmosphere interaction has enabled us to unlock the mysteries of ENSO and other climate phenomena, leading to skillful predictions of ENSO [*Cane et al.*, 1986], its global impacts [*Latif et al.*, 1998], and certain aspects of tropical Atlantic variability [*Saravanan and Chang*, this volume]. Indeed, the success of TOGA has proved a premise that a better understanding of ocean–atmosphere interaction helps improve model simulation, and along with an adequate observing system leads eventually to useful climate prediction.

While great progress has been made, our understanding of tropical ocean–atmosphere interaction is still incomplete, especially in the Indian and Atlantic sectors. Our understanding of the extratropical ocean–atmosphere system is in an even less satisfactory state, limiting our ability to extend the success of seasonal forecasts into longer time leads. Most of the world's population lives on continents. Past research shows that atmospheric bridges play an important role in influencing continental climate. But such robust atmospheric bridges from the tropics may only apply to limited geographical areas, and are interfered by weather noise and other modes of climate variation. It remains unclear whether land processes have an active role to play in generating or maintaining climatic anomalies on interannual and longer timescales. *Webster et al.* [1998] gave a comprehensive review of studies of the structure and variability of the Asian–Australian monsoon system and its relationships with ENSO. Another area that is not covered in this volume but of great social and economic importance is the interaction of coastal/marginal seas with the overlying atmosphere. There, ocean bottom topography and land orography are generally important. Recent satellite studies reveal interesting features in coastal and marginal seas that involve interaction with the atmosphere [*Xie*, 2004a; *Chelton et al.*, 2004; *Hu and Liu*, 2003].

Anthropogenic climate changes also pose a great challenge for the climate research community as well as for the mankind in general. On the global scale, exchange of carbon dioxide and heat between the atmosphere and ocean has likely slowed down the rate of temperature rise (global warming) so that its effects are just beginning to be observed. On the regional scale, the temperature rise is unevenly distributed, with the greatest rise over continents but some cooling over the North Pacific and North Atlantic Oceans during the second half of the past century.

Human–induced climate change may have some further surprises in store. Some model projections suggest a slowdown of the North Atlantic meridional overturning circulation in a warmer climate due to increased sea surface evaporation, as well as increased rainfall in the mid/high latitudes as a result of a more intense hydrological cycle [*Manabe and Stouffer*, 1993]. This slow down of the meridional overturning circulation reduces the northward transport of warm surface water and may send the North Atlantic and Europe into a colder climate. Such abrupt climate changes are believed to have happened before as the climate warmed at the end of the last ice age. What causes such a rapid spread of North Atlantic cooling is unclear [*Broecker*, 2003]. Tropical air–sea interaction may have played an important role in transmitting this regional cooling into a global–scale event [*Dong and Sutton*, 2002; *Xie and Carton*, this volume]. Global warming may also affect ENSO variability. However, the relationship between ENSO and global warming is unknown [*Wang and Picaut*, this volume] and needs further examination.

Coupled ocean–atmosphere general circulation models are an important tool for predicting the evolving climate and projecting the impact of human activity. These models, however, suffer significant biases in their simulation of the current climate and its variations [e.g., *Mechoso et al.*, 1995; *Davey et al.*, 2002]. Such biases are certain to affect the model simulation of the climate response to increased greenhouse gases, casting doubts on the future climate projections produced by these models. These biases generally reflect a poor understanding of physical processes such as clouds, their response to SST changes and treatment in models. We anticipate that the improved understanding of the causes of tropical climate variability as represented by the papers in this volume will help isolate model deficiencies and thus contribute to improved predictive capability.

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