1

Variability of the Oceans

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1.1 Introduction

The oceans have a huge capability to store, release, and transport heat, water, and various chemical species on timescales from seasons to centuries. Their transports affect global energy, water, and biogeochemical cycles and are crucial elements of Earth’s climate system. Ocean variability, as represented, for example, by sea surface temperature (SST) variations, can result in anomalous diabatic heating or cooling of the overlying atmosphere, which can in turn alter atmospheric circulation in such a way as to feedback on ocean thermal and current structures to modify the original SST variations. Ocean–atmosphere interactions in one ocean basin can also influence remote regions via interbasin teleconnections that can trigger responses having both local and far-field impacts. This chapter highlights the defining aspects of the climate in individual ocean basins, including mean states, seasonal cycles, interannual-to-interdecadal variability, and interactions with other basins. Key components of the global and tropical ocean observing system are also described.

1.2 Pacific Ocean

The Pacific Ocean extends from the Arctic Ocean in the north to the Southern Ocean in the south and is bounded by Asia and Australia in the west and the Americas in the east (Figure 1.1). Major features include wind-driven circulations in the subtropical gyres of the North and South Pacific and the subpolar gyre in the North Pacific. The clockwise circulation of the North Pacific subtropical gyre consists of the strong western boundary Kuroshio Current, the westerly driven North Pacific Current, the eastern boundary California Current, and the trade wind–driven North Equatorial Current (NEC). The clockwise circulation of the North Pacific subpolar gyre is seen north of 50°N, which includes the eastward-flowing North Pacific Current, the poleward and westward flowing Alaska and Aleutian Currents, and the southward flowing Oyashio Current. In the South Pacific, the subtropical gyre consists of the East Australian Current to the west, the Peru Current to the east, the eastward South Equatorial Current (SEC) to the north, and the Antarctic Circumpolar Current (ACC) to the south. The complex system of equatorial currents is further mentioned in other chapters.
1.2.1 Seasonal Cycle

The Pacific Ocean exhibits prominent large-scale variations in SST on timescales ranging from seasonal to interdecadal as well as long-term trends. One outstanding example of SST variability on seasonal timescales is the seasonal cycle of the cold tongue in the tropical eastern Pacific. Here, the annual cycle of SST is stronger than in any other part of the tropics (Kessler et al., 1998). The local seasonal variations are dominated by the annual harmonic with values higher than their annual mean from January through June and lower from July through December, even though the sun crosses the equator twice a year (e.g., Li and Philander, 1996; Yu and Mechoso, 1999). In contrast, in the tropical western Pacific, SSTs are dominated by semiannual components, and temperatures warm up and cool down twice a year. Another major feature of the SSTs in the tropical eastern Pacific is a pronounced interhemispheric asymmetry, with cooler temperatures in the Southern Hemisphere than in the Northern Hemisphere (Philander et al., 1996; McPhaden et al., 2008). Ocean–atmosphere coupling processes are particularly critical for the variability of the cold tongue–ITCZ complex in the region, where the thermocline depth is shallow and ocean advection has a strong influence on SSTs.
1.2.2 Interannual Variability

El Niño and the Southern Oscillation. On interannual timescales, El Niño and its opposite phase La Niña represent the most prominent mode of Pacific Ocean variability (Figure 1.2a). El Niño and La Niña are warming and cooling events, respectively, which occur every two to seven years in the tropical eastern-to-central Pacific (Rasmusson and Carpenter, 1982). El Niño events typically begin in boreal spring, develop in summer and fall, peak in winter, and decay in the following spring. These ocean variation events are accompanied with seesaw fluctuations in sea-level pressures over the southern Pacific and Indian Oceans, which are referred to as the Southern Oscillation (SO; Bjerknes, 1969). The coupling between the oceanic (i.e., El Niño and La Niña) and atmospheric (i.e., the SO) anomalies and the positive feedback between them enables the El Niño/Southern Oscillation (ENSO) phenomenon to grow to strong intensities to profoundly impact climate worldwide.

Since Bjerknes (1969) first recognized the coupled nature of ENSO, extensive effort has been expended by the research community to improve its description, understanding, and prediction. This effort resulted in novel theories that succeed in explaining many features of ENSO. (Chapter 3 reviews outstanding aspects of these theories.) The effort also resulted in the development of numerical climate models that can simulate and predict ENSO events from months ahead with useful skill.

In more recent times it has been widely recognized that ENSO events since the 1990s have been noticeably different from those in earlier decades (e.g., Kao and Yu, 2009; Lee and McPhaden, 2010; Yu et al., 2012a; Capotondi et al., 2015). The earlier reported or
“canonical” events (Rasmusson and Carpenter, 1982) starting from the South American coast are now referred to as the eastern Pacific (EP) El Niño (Kao and Yu, 2009; Figure 1.2a). El Niño in recent decades has tended to develop more frequently in the tropical central Pacific near the International Dateline (Figure 1.2b). These events are referred to as a Central Pacific (CP) El Niño (Kao and Yu, 2009), Dateline El Niño (Larkin and Harrison, 2005), El Niño Modoki (Ashok et al., 2007), or Warmpool El Niño (Kug et al., 2009). The EP and CP types of ENSO produce different teleconnections through the atmosphere and impact global climate in different ways (e.g., Larkin and Harrison, 2005; Ashok et al., 2007; Yu et al., 2012b). The change in ENSO type may be due to variations of the background state changes in the tropical Pacific, which are either caused by anthropogenic warming (Yeh et al., 2009; Kim and Yu, 2012) or integral part of decadal/multidecadal variability related to Pacific Decadal Oscillation or Atlantic Multidecadal Oscillation (McPhaden et al., 2011; Yu et al., 2015a). Random climate fluctuations have also been suggested as a plausible reason for changes in ENSO type on decadal timescales (Newman et al., 2011).

The oceanic anomalies associated with El Niño disturb the Walker circulations in the atmosphere to induce warming in the tropical Indian Ocean and the tropical North Atlantic Ocean about three to six months after its peak (e.g., Lau and Nath, 1996; Klein et al., 1999; Alexander et al., 2002; Cai et al., 2019). They also excite atmospheric wave trains that propagate into middle and high latitudes of both hemispheres to remotely influence precipitation and temperature over continents especially North and South America (e.g., Ropelewski and Halpert, 1986; Karoly, 1989; Mo, 2000; Yeh et al., 2018), as well as the Arctic polar vortex configuration (e.g., Sassi et al., 2004; García-Herrera et al., 2006; Manzini et al., 2006), Southern Ocean SSTs (e.g., Ciasto and Thompson, 2008; Yeo and Kim, 2015), Antarctic sea ice concentrations (e.g., Liu et al., 2004; Stammerjohn et al., 2008; Yan and Li, 2008), and Antarctic surface air temperatures (e.g., Kwok and Comiso, 2002; Ding et al., 2011; Schneider et al., 2012). The precise aspects of these impacts can be different between the EP and CP types of El Niño.

The Pacific Meridional Mode (PMM). The PMM is a leading mode of interannual variability in the northeastern Pacific and is characterized by covariability in SST and surface wind (Chiang and Vimont, 2004; Figure 1.3). Wind fluctuations associated with extratropical atmospheric variability, particularly those linked with the North Pacific Oscillation (NPO; Walker and Bliss, 1932; Rogers, 1981; Linkin and Nigam, 2008), induce SST anomalies in the subtropical Pacific via surface evaporation. The SST anomalies then feedback on the atmosphere to modify the winds via convection, which tends to produce the strongest wind anomalies to the southwest of the subtropical SST anomalies (Xie and Philander, 1994), where new SST anomalies can be formed through anomalies in evaporation. The atmosphere then continues to respond to the new SST anomalies by producing wind anomalies further southwestward. Through this wind-evaporation-SST (WES) feedback (Xie and Philander, 1994), the SST anomalies initially induced by the extratropical atmosphere can extend southwestward from near Baja, California toward the tropical central Pacific to form the spatial pattern of the PMM (see Chapter 3).

The PMM tends to reach its strongest intensity in boreal spring. A strong association has been found between a spring PMM index and a subsequent winter ENSO index.
About 70 percent of El Niño events occurring between 1958 and 2000 were preceded by SST and surface wind anomalies resembling the PMM. The PMM was suggested to trigger ENSO events either by exciting equatorial ocean waves that propagate toward the eastern Pacific when the PMM wind anomalies arrive at the equator (e.g., Alexander et al., 2010), or by directly increasing the ocean heat content in the equatorial Pacific via modulations in the strength of the trade winds (Anderson, 2004). Some studies consider the PMM particularly important in triggering the CP types of ENSO (e.g., Yu et al., 2010; Yu et al., 2017; Yu and Fang, 2018; Yang et al., 2018). The PMM and its associated subtropical Pacific coupling can be a key source of complexity in ENSO evolution (Yu and Fang, 2018).

A Southern Hemispheric analogue of the PMM develops in the southeastern Pacific and is termed the southern PMM. This is also characterized by covariability in SST and trade wind anomalies extending from the Peruvian Coast toward the equatorial central Pacific. The southern PMM can influence the deep tropics through connection with cold tongue

![Figure 1.3 SST (contours) and surface wind (vectors) anomalies regressed onto the PMM index using HadISST and NCEP-NCAR reanalysis data during period 1958–2014. The PMM index comes from www.aos.wisc.edu/~dvimont/MModes/Home.html. (Figure created by author)
ocean dynamics to influence the development of an EP ENSO (Zhang et al., 2014; You and Furtado, 2017).

Besides influencing the ENSO onset, the PMM can also modulate the occurrence of western Pacific typhoons and eastern Pacific hurricanes with associated high precipitation occurrences in East Asia and South America (e.g., Li et al., 2011; Zhang et al., 2016, 2017; Murakami et al., 2017).

The Pacific Warm Blob (PWB). The 2013–2015 PWB was an extreme event of interannual SST variability of the extratropical North Pacific that persisted for an unusually long time and produced significant impacts on regional climate and ecosystem (Bond et al., 2015; Figure 1.4). The event developed in the Northeastern Pacific Ocean in mid-2013 as a persistent near-surface warming that lasted through 2015. It was reported that the event caused a dramatic species range shift in the Gulf of Alaska during the summer and fall of 2014 (Medred, 2014), delayed the onset of upwelling off California during the 2015 summer (Peterson et al., 2015; Zaba and Rudnick, 2016), and altered spring temperatures in the Pacific Northwest by enhancing warm air transport into the region (Bond et al., 2015). The PWB event peaked during winter 2014 and summer 2015 when SST anomalies in the warming region reached as high as 2–3°C and penetrated as deep as 180 m below the ocean surface (Bond et al., 2015; Hu et al., 2017). During the following winter, the warm water patch propagated from the Gulf of Alaska toward the coastal regions resulting in an arc-shaped warming off the North American coast (Amaya et al., 2016; Di Lorenzo and Mantua, 2016; Gentemann et al., 2017). Recently, Myers et al., (2018) on the basis of an observation analysis suggested that a positive cloud-surface temperature feedback was key to the extreme intensity of the oceanic heatwave off Baja, California associated with the PWB.

The PWB event was accompanied by an unusually persistent ridge in the atmospheric flow over the Northeastern Pacific (Seager et al., 2014; Swain et al., 2014; Bond et al., 2015; Hartmann 2015; Seager et al., 2015; Amaya et al., 2016; Di Lorenzo and Mantua 2016; Hu et al., 2017). This anomalous high-pressure system induced clockwise surface wind anomalies that reduced local surface evaporation and weakened cold ocean advection in the region (Bond et al., 2015). This anomalous high-pressure system was linked to

Figure 1.4 SST anomalies during the Pacific warm blob (PWB) event December 2014 to February 2015.

several atmospheric teleconnection patterns, including the Pacific North American pattern (PNA; Wallace and Gultzer, 1981), the NPO, and the Tropical Northern Hemisphere pattern (TNH; Mo and Livezey, 1986). Liang et al. (2017) found that this pattern was particularly important to producing the PWB.

1.2.3 Interdecadal Variability

The Pacific Decadal Oscillation (PDO). The main mode of SST variability in the Pacific Ocean on decadal or interdecadal timescales is the PDO (Mantua et al., 1997; Zhang et al., 1997), or the closely related Interdecadal Pacific Oscillation (England et al., 2014). The PDO is characterized by a horseshoe pattern of SST anomalies in the tropical eastern Pacific together with SST anomalies with opposite polarity in the central North Pacific with dominant periodicities in the 50–70 year and bidecadal year bands (Minobe, 1999). During the period of reliable instrumental records, the PDO shifted its phase twice: from a negative to a positive phase around 1976–1977 and back to a negative phase around 1999–2000. Around the time of these shifts, significant changes occurred in the World Ocean. The PDO was initially considered to be a physical mode of climate variability that results from atmosphere–ocean coupling within the North Pacific (e.g., Latif and Barnet 1994; Alexander and Deser, 1995) or extratropical–tropical interactions (Gu and Philander, 1997). More recent views consider it not a single physical mode but a combination of several processes operating on various timescales (Newman et al., 2016). Ocean circulation patterns associated with the PDO involve changes in the subtropical cells (McCreary and Lu, 1994) and affect equatorial upwelling on decadal time scales (McPhaden and Zhang, 2002, 2004).

1.2.4 Long-Term Trend

The Pacific Ocean has experienced an overall warming trend since the 1880s. This feature has been primarily attributed to the increased atmospheric concentrations of greenhouse gases that cause ocean temperatures to rise as they are absorbed by the ocean (IPCC, 2013). The largest warming trends (around 1.4–2.0°C per century) occur off the eastern Asia where the northward flowing Kuroshio Current and southward flowing Oyashio Current converge. Accurate assessment of long-term trends is, however, inevitably limited by the effects of imperfect spatial and temporal sampling and inhomogeneous measurement practices, thereby inducing considerable uncertainty (Rayner et al., 2003). The uncertainty of the long-term trends is especially more evident in the tropical Pacific, giving rise to a pronounced discrepancy over the central and eastern tropical Pacific among various observational or reconstructed datasets. During the twentieth century, for example, one widely accepted SST reconstruction from the National Oceanic and Atmospheric Administration (NOAA; Smith et al., 2008) shows a robust warming of approximately 0.4–1.0°C per century over the entire tropical Pacific. In contrast, another reconstruction from the Hadley Center (Rayner et al., 2003) reveals a weak cooling in the central to eastern tropical
Pacific (approximately 0–0.4°C per century) with warming in the west, i.e., a La Niña–like pattern (e.g., Vecchi et al., 2008; Deser et al., 2010).

In reference to trends over decadal or interdecadal timescales, a phase shift of the PDO that generates a horseshoe pattern with opposite SST anomalies between the tropical eastern Pacific and central North Pacific has been identified as a key component for the global mean SST changes since it overwhelms the background warming trends (e.g., Trenberth, 2015; Meehl et al., 2016). Regarding this, modeling studies have attributed the latest slowdown period, or “hiatus” period in the rise in global mean SST between the late 1990s and the early 2010s to a negative phase of the PDO, which started around 1999–2000. This negative PDO phase manifests a decadal intensification of the Pacific trade winds and surface cooling in the central to eastern tropical Pacific, with a pattern that tends to mitigate the rise in global mean SST by cooling the global atmosphere (e.g., Kosaka and Xie, 2013; England et al., 2014; Xie and Kosaka, 2017).

1.3 Atlantic Ocean

The Atlantic Ocean is connected with the Arctic Ocean and the Nordic Seas to the North, and the Southern Ocean to the south. Its most important wind-driven circulations are the subtropical gyres in the North and South Atlantic and the subpolar gyre in the north (Figure 1.1).

The north Atlantic subtropical gyre is formed by the strong western boundary Gulf Stream and North Atlantic currents, and the eastern boundary Canary Current Upwelling System. The NEC closes the gyre north of the equator. North of 40N, the North Atlantic Current (NAC) conforms to the southern edge of the subpolar gyre, which is closed by the western East Greenland and Labrador currents. The two northern gyres are connected through the NAC (Talley et al., 2011). The south Atlantic subtropical gyre’s major currents are the western boundary Brazil Current, the Benguela current in the eastern boundary upwelling system, the SEC to the north, and the ACC to the south (Talley et al., 2011). When the SEC reaches the South American coast, part of it diverges northward into the North Brazil Current. The complex system of equatorial currents is further discussed in Chapter 3.

The North Atlantic Ocean is one of the sources of deep-water formation in the global oceans, creating an overturning circulation that increases the transport of water and heat from the tropical regions to the north Atlantic sector, resulting in a cross-equatorial northward transport or heat from the South Atlantic Ocean.

1.3.1 Seasonal Cycle

The seasonal cycle is the dominant component in the variability of the tropical Atlantic Ocean. This important variability component is highly coupled with the atmosphere and affected by the shape of surrounding continents (Li and Philander, 1997; Okumura and Xie, 2004). An outstanding aspect of the seasonal cycle of the tropical Atlantic is the
development of a tongue of cold SSTs from April to July, followed by a slow warming over the rest of the year (Okumura and Xie, 2004). SSTs in the equatorial Atlantic band are highest in boreal spring, when the solar irradiation is maximum in the tropics and the trade winds are weakest. Also, in this season, the thermocline is deeper in the east and the ITCZ is located at the equator above the band of maximum SSTs. Starting in April, the trade winds intensify, and a positive zonal pressure gradient develops along the equator, while SSTs in the eastern Atlantic cool down and the thermocline shoals. The cooling is maximum in boreal summer, when a well-developed cold tongue appears south of the equator. As the tropical Atlantic cools down, the oceanic ITCZ moves north following the maximum SST band. The formation of the cold tongue is tightly related to the onset and development of the West African Monsoon (WAM): From boreal spring, solar irradiation heats the African continent north of the equator, establishing a meridional gradient of surface pressure along the zonally oriented coast with the equatorial Atlantic that produces an enhancement of the ocean-to-continent southerly winds. These winds induce ocean upwelling and cool the eastern equatorial Atlantic Ocean, which, in turn, further intensifies the southerly winds and the WAM (see Chapter 7).

1.3.2 Interannual Variability

The Atlantic Ocean has been attracting increased recognition in recent decades as an important driver of the variability of the global climate system. This is apparent in the growing literature focused on the region and in the recent efforts aimed at building a more complete Atlantic observing system (see Foltz et al., 2019). Climate variability in the tropical Atlantic strongly influences the climate of the surrounding continents, affecting winds, precipitation, and temperature, and having important impacts on society through changes in hurricane activity and marine productivity, among other phenomena. Likewise, the Atlantic Ocean is strongly influenced by other components of the climate system. Moreover, there are intricate links between the interannual variability of the Atlantic Ocean and the other ocean basins, as well as with slow changes in the background state in which the interannual variability is embedded.

Early reviews (e.g., Marshall et al., 2001) described the variability of the north Atlantic climate as a combination of three main components: (1) Tropical Atlantic Variability (TAV), (2) the North Atlantic Oscillation (NAO), and (3) the Atlantic Meridional Overturning Circulation (AMOC). More recently, the multidecadal variability of North Atlantic SSTs, i.e., the Atlantic Multidecadal Variability (AMV), has also become a focal point for research. The following subsections introduce these components and their more important climate signatures. Other chapters in the book present more in-depth discussion of these topics.

The Atlantic Niño. The first mode of interannual variability in the tropical Atlantic is known as the Equatorial Mode, the Zonal Mode, or the Atlantic Niño. The pattern of SST anomalies associated with this mode shows maximum loadings in the eastern equatorial Atlantic that peak in boreal summer (Figure 1.5a). It has been established that the dynamics
of the Atlantic Niño are similar to the Pacific ENSO (Zebiak, 1993), in which the Bjerknes feedback (Bjerknes, 1969) is crucial. In an Atlantic Niño, warm SST anomalies in the eastern equatorial Atlantic extend to the south along the African coast, weaker trade winds develop in the western part of the basin, and the thermocline deepens in the east (Keenlyside and Latif, 2007; Deppenmeier et al., 2016). The opposite occurs for an Atlantic Niña. These changes occur through the propagation of equatorial Kelvin and Rossby waves (Polo et al., 2008). Nevertheless, some recent works have alluded to a thermodynamic mechanism triggered by stochastic atmospheric fluctuations as a controlling factor for the development of the Atlantic Niño (Nnamchi et al., 2015, 2016). The relative importance of dynamics versus thermodynamics in the generation of this mode is thus controversial at the present time, although the current consensus is that the dynamical mechanisms are dominant (Dippe et al., 2017; Jouanno et al., 2017).

The Atlantic Niño has direct impacts on the WAM system. The positive phase of the Atlantic Niño is associated with a southward migration of the ITCZ, delaying the onset of the WAM and increasing rainfall over the coast of Guinea (Sultan and Janicot, 2003; Okumura and Xie 2004; Rodríguez-Fonseca et al., 2011, 2015). In contrast, the southward migration of the ITCZ results in increased precipitation over Northeast Brazil (Mechoso and Lyons, 1988; Mechoso et al., 1990; Torralba et al., 2015). Outside the tropical Atlantic, it has been shown that the Atlantic Niño can alter the atmospheric circulation of the North Atlantic-European region in summer (Losada et al., 2012) and early winter (Haarsma and Hazeleger, 2007; García-Serrano et al., 2011). There are also impacts on the tropical Indian

![Figure 1.5 Dominant modes of Empirical Orthogonal Functions (EOFs) of variability in the tropical Atlantic Ocean. First (a) and second (b) EOFs of the monthly tropical Atlantic SSTs averaged between 30°S–30°N and 70°W–20°E. The modes were calculated from ERSSv3.b dataset (Smith et al., 2008) for the period 1854–2014, using a 13-year high pass filter. All year months were considered for the calculation of the EOF. The year-to-year standard deviation of the principal components shows the seasonality of the modes. (Figure created by author)