Three-ocean interactions and climate variability: a review and perspective

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Abstract

Interactions among the Pacific, Atlantic and Indian Oceans through ocean–atmosphere coupling can initiate and/or modulate climate variability. The Pacific Ocean is home to ENSO which affects other oceans through atmospheric bridges and the oceanic Indonesian throughflow (ITF). A warm Indian Ocean can produce atmospheric Kelvin waves that propagate eastward and increase equatorial easterly wind anomalies in the western Pacific and thus cool eastern Pacific sea surface temperature (SST). A positive Indian Ocean dipole establishes a southwestward pressure gradient force in the ITF region which increases the ITF transport and decreases ocean heat content in the western Pacific and may cool eastern Pacific SST. The Indian Ocean can also influence the Atlantic by atmospheric bridge and the oceanic Agulhas leakage south of Africa. Midlatitude North Atlantic SSTs may affect Pacific climate variability: (1) The Atlantic multidecadal oscillation (AMO) influences North Pacific variability; (2) The warm AMO phase increases the occurrence of central Pacific (CP)-type El Niño; (3) The warm AMO phase helps induce anomalous cyclonic circulation in the tropical western North Pacific; and (4) A cold midlatitude North Atlantic Ocean in the summer may initiate an El Niño in subsequent year via the East Atlantic/West Russia teleconnection. A warm tropical North Atlantic in the spring can induce a CP-type La Niña in the subsequent winter, via two pathways of the tropical eastern North and South Pacific. Finally, the Atlantic Niño (Niña) in the summer, through the Walker circulation and ocean dynamics, helps induce an eastern Pacific-type La Niña (El Niño) in the subsequent winter. The Atlantic Niño can also warm the tropical western Indian Ocean and weaken Indian monsoon rainfall.

1 Introduction

Climate can be thought of as the average weather conditions that we experience, or is the statistics of weather over long periods of time. Climate varies naturally on a whole range of timescales and these variations can have profound impacts on the Earth, and thus climate is of vital importance for both our economy and society. Because of its importance, climate has been intensively studied and focused during the past 20–30 years. The past 20–30-year climate studies have reached a consensus that ocean–atmosphere interaction especially over the tropical oceans can form various climate phenomena or climate variability. The word "climate" normally indicates ocean–atmosphere interaction or is a synonym of ocean–atmosphere interaction.

El Niño-Southern Oscillation (ENSO) is the first climate phenomenon which was hypothesized to be a result of the interaction between the ocean and atmosphere (Bjerknes 1969). Given a warm sea surface temperature (SST) anomaly in the equatorial eastern Pacific, the east–west SST gradient in the tropical Pacific is reduced. The atmospheric response is a reduction of the east–west gradient of sea level pressure (SLP), and consequently resulting in a weakening of the Walker circulation and the easterly trade winds. The weakening of the easterly trade winds causes warm water in the western Pacific warm pool to move eastward, thus enhancing the initial warm SST anomaly in the equatorial eastern Pacific. This Bjerknes feedback continues to amplify and enhance initial SST perturbations into a large-scale warming in the tropical eastern Pacific—an El Niño event. In addition to the Bjerknes feedback, other ocean–atmosphere feedback processes were also proposed and hypothesized after the 1990s. For example, a positive wind-evaporation-SST (WES) feedback has been proposed for causing inter-tropical convergence zone (ITCZ) variability (Xie and Philander 1994) and producing tropical Atlantic north–south

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variability (Chang et al. 1997). Philander et al. (1996) introduced a positive SST-cloud-shortwave radiation feedback for climate variations in the tropical southeast Pacific. Wang and Enfield (2003) suggested a positive SST-cloud-longwave radiation feedback in the region of the Western Hemisphere warm pool.

Although these local ocean–atmosphere feedback processes can form and explain climate phenomena, they are not the whole story of climate. In particular, inter-ocean interactions can initiate and/or modulate climate variations over the global oceans (In this paper, inter-ocean, inter-basin, and three-ocean interactions are alternately used). The study of inter-ocean interactions is in the infancy stage, but it has been demonstrated that inter-ocean interactions are very important for global climate. The purpose of the present paper is to review past research on the influences of three-ocean interactions (the Pacific, Atlantic and Indian Oceans) on climate and to provide a perspective on this topic. The paper only reviews three-ocean interactions on timescales longer than seasonal timescales. The variations on shorter timescales such as intraseasonal timescales can also link one ocean basin to others; however, this review does not discuss these. The review does not include the polar oceans and the Southern Ocean either. Inter-ocean interactions are emerging as a hot research topic in the climate community. The present review is a timely one in this topic, complementary but distinct from a recent independent review by Cai et al. (2019). Cai et al. (2019) only focused on the interactions among the three tropical oceans by atmospheric bridges. The present review also includes three-ocean interaction processes operated through the midlatitudes and the oceanic pathways. Additionally, this review is comprehensive in Atlantic influences on the Pacific and Indian Oceans.

The paper is organized as follows. Section 2 summarizes the previous studies of the influence of the Pacific Ocean on Indian and Atlantic climate. Section 3 discusses the influence of the Indian Ocean on Pacific and Atlantic climate. Section 4 provides an overview of the influence of the Atlantic Ocean on Pacific and Indian climate. Finally, a summary and future work are given in Sect. 5.

2 Influence of the Pacific Ocean on Indian and Atlantic climate

Among the three oceans, the Pacific Ocean is the largest one in size and also hosts the largest climate signal of ENSO in Earth's climate of modern time. Because the Pacific is home to ENSO, it has been extensively studied and its climate variability and influence on other ocean basins have been relatively well understood and emphasized during the past. The influences of the Pacific Ocean on other ocean basins

can be operated through both the atmospheric bridges and oceanic pathways.

2.1 Atmospheric bridges

2.1.1 Walker and Hadley circulations

From the name, the atmospheric bridges indicate atmospheric processes or patterns that can connect and link two or more sites far away. One example of the atmospheric bridges is the Walker circulation and the Hadley circulation, which can link the Pacific Ocean to other ocean basins. The occurrence of an El Niño event in the Pacific redistributes atmospheric heating sources and then changes atmospheric convection and circulation. El Niño can also induce tropospheric warming that produces atmospheric and SST changes far away from the Pacific (Chiang and Sobel 2002; Chiang and Lintner 2005). Klein et al. (1999) presented a schematic diagram of the anomalous Walker and Hadley circulation in association with El Niño events by using some data measured by satellite and ship (Fig. 1). By analyzing atmospheric reanalysis product, Wang (2002a, b) further confirmed the ENSO-altered Walker and Hadley circulations similar to Fig. 1. Due to the heating redistribution caused by El Niño, global atmospheric circulation is altered: The equatorial central and eastern Pacific is characterized by anomalous ascent, whereas the equatorial Atlantic and the equatorial Indo-western Pacific are featured with anomalous descent. As a result, when an El Niño occurs in the tropical Pacific, the Walker circulations in the Pacific and Atlantic are weakened. Meridionally, the regional Hadley circulations are also changed in association with El Niño. Figure 1 showed that anomalous ascent in the equatorial Pacific is associated with anomalous subsidence in the subtropical Pacific, and anomalous decent in the equatorial Indian and Atlantic is corresponded to anomalous ascent in the subtropical Indian and Atlantic. As a result, the regional Hadley circulation is strengthened in the eastern Pacific, but the regional Hadley circulations are weakened in the Atlantic and Indo-western Pacific basins.

These Walker and Hadley circulation changes provide a connection of ENSO events with other oceans far away from the equatorial central and eastern Pacific Ocean. The Walker and Hadley circulation changes are associated with the anomalous ascent and descent which result in changes of atmospheric wind, humidity, cloud cover and so on. These changes in turn influence surface heat fluxes, wind and ocean circulation, resulting in SST variations in other oceans. Therefore, when an El Niño event is developed and matured, a cold North Pacific Ocean, a cold tropical western North Pacific (WNP) Ocean, a basin-side warm Indian Ocean (IOB) and a warm tropical North Atlantic Ocean occur simultaneously or in the following spring (Fig. 2).
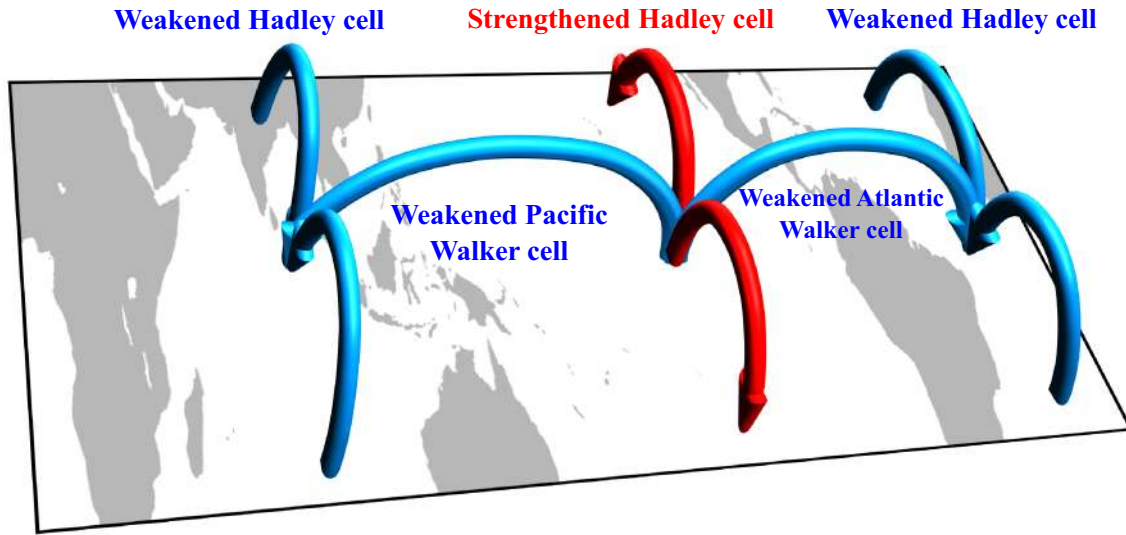
Fig. 1 Schematic diagram of the changed Walker and Hadley circulations associated with an El Niño event. When an El Niño occurs, both the Pacific and Atlantic Walker circulations are weakened. Addition-ally, the regional Hadley circulation is strengthened over the eastern Pacific but weakened over the Atlantic and Indo-western Pacific
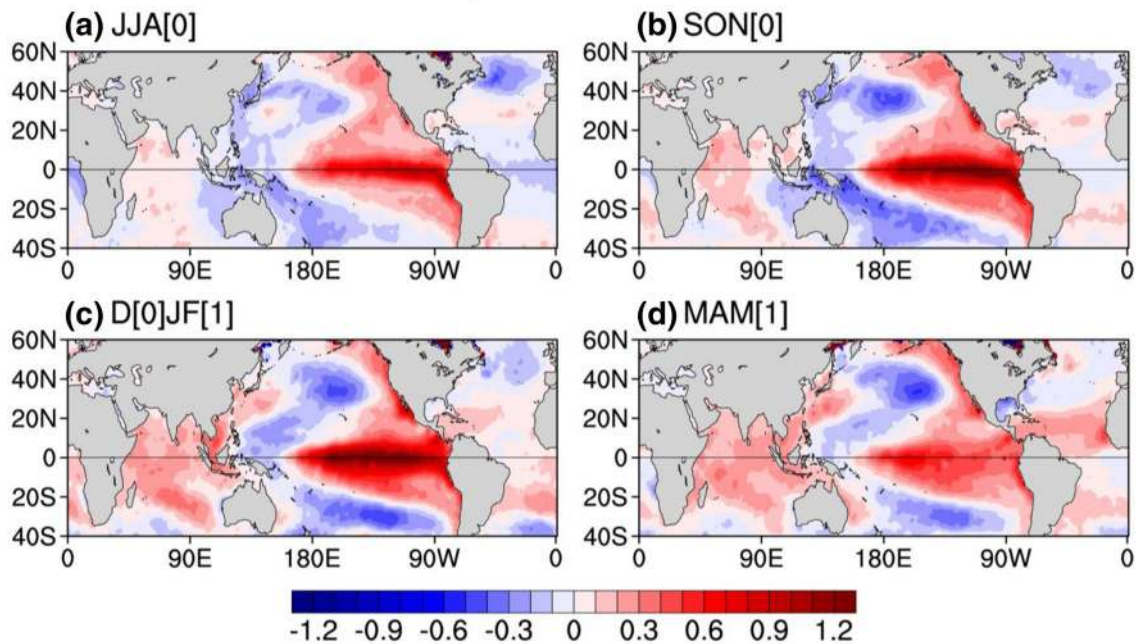
Fig. 2 Regressions of DJF Niño3.4 index onto global SST anomalies. Shown are the regressions (°C per °C) of D[0]JF[1] Nino3.4 index onto the SST anomalies during **a** JJA[0], **b** SON[0], **c** D[0]JF[1], and **d** MAM[1]. "0" indicates the El Niño year and "1" is the El Niño's subsequent year

Because the ocean takes time to respond El Niño's influences, the maximum remote influence of El Niño on other ocean basins usually delays one season as shown in Fig. 2d. However, El Niño can also affect other ocean basins in early time such as during the developing phase of the El Niño year. An example is the influence on the Indian Ocean dipole (IOD) during the El Niño developing year, with the cold SST anomalies in the eastern Indian Ocean and warm

Pacific-North American (PNA) Pattern
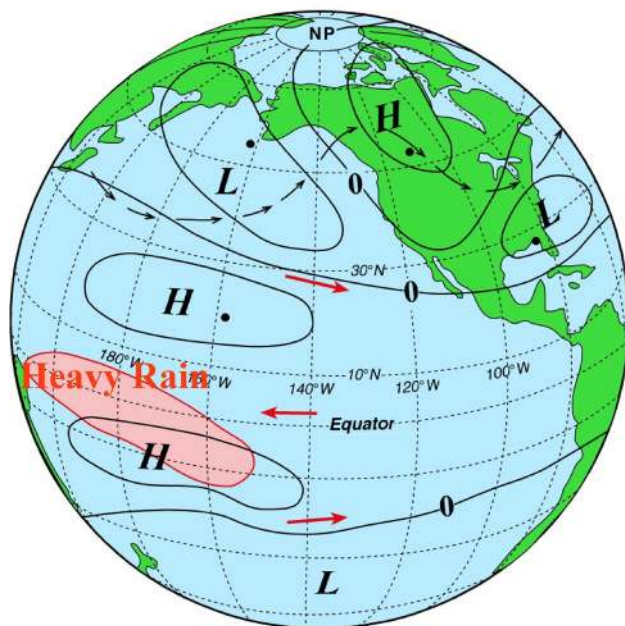
Fig. 3 Schematic diagram of middle and upper tropospheric geopotential height anomalies associated with the Pacific North American (PNA) pattern. The red shading represents rainfall associated with El Niño. "H" and "L" represent high and low geopotential height anomalies, respectively. The black and dashed line with arrows indicates the subtropical jet stream

SST anomalies in the western Indian Ocean (Fig. 2b). The eastern cold part of the IOD is short-lived, lasting only a few months. The cold SST anomalies in the eastern Indian Ocean quickly disappear after the fall of the El Niño developing year. Then the basin-side warming of the IOB quickly develops in the winter and peaks in the subsequent spring (Fig. 2c and d).

2.1.2 Pacific–North American pattern

In addition to the Walker and Hadley circulations, ENSO can also remotely affect remote places via other atmospheric teleconnections. An example of the atmospheric teleconnections to other middle-high latitudes is the Pacific-North American (PNA) pattern (Wallace and Gutzler 1981). This PNA teleconnection pattern is often generated by a heating source in the tropical Pacific (Fig. 3; Horel and Wallace 1981). An El Niño produces positive rainfall anomalies in the equatorial central Pacific which heat the atmosphere. In response to the heating, the PNA wave train is produced: The middle and upper troposphere is featured with a high in the tropical North Pacific, a low in the North Pacific, a high in northwestern America and a low in the southeastern United States and the Gulf of Mexico. The PNA pattern is in a barotropic structure, so the SLP anomalies have the same sign as those in the troposphere. The low SLP anomalies in the North Pacific enhance the westerly winds in the North Pacific which increase surface heat flux from the ocean and southward oceanic advection associated with Ekman transport (Alexander et al. 2002). Both of these effects are to cool the North Pacific Ocean, explaining the cold North Pacific Ocean associated with El Niño shown in Fig. 2. Similarly, the low SLP anomalies over the southeastern United States and the Gulf of Mexico correspond to a weakening of the easterly trade winds and thus a decrease of the latent heat flux (Enfield and Mayer 1997). The decrease of heat flux loss from the ocean, along with local processes such as the SST-cloud-longwave radiation feedback of Wang and Enfield (2003), results in the warming in the tropical North Atlantic during the El Niño's subsequent spring as shown in Fig. 2d.

2.1.3 Pacific–South American pattern

The Pacific–South American (PSA) pattern is the Southern Hemisphere counterpart of the PNA (Mo 2000). El Niño events, with atmospheric heating and convection in the equatorial central Pacific, can also excite a PSA pattern of Rossby wave train that emanates from the tropical Pacific and propagates to the high-latitude Southern Hemisphere (Hoskins and Karoly 1981; Rasmusson and Mo 1993). The PSA pattern manifests as low pressure anomalies off western New Zealand, high pressure anomalies over the Amundsen Sea low, and low pressure anomalies off South America. The low pressure anomalies of the PSA pattern in South America alter atmospheric circulation and cause rainfall anomalies over South America. It is also found that the PSA modulates the westerlies in the South Pacific which in turn affect rainfall in New Zealand, air temperature and sea ice in the Antarctic Peninsula, and blocking events in the high-latitudes of the South Pacific (Kwok and Comiso 2002; Renwick 2002). In summary, El Niño events can excite both the PNA and PSA teleconnection patterns that remotely affect the mid-high latitudes of both hemispheres.

2.1.4 Pacific–Japan pattern

The tropical-extratropical teleconnection also includes the Pacific–Japan (PJ) pattern that features a meridional alternating negative and positive anomaly structure in lower tropospheric circulation and rainfall anomalies in the WNP, and provides a link between the tropics and extratropics in the WNP (Nitta 1986, 1987; Kosaka and Nakamura 2006). The positive PJ pattern is featured with high, low and high SLP anomalies in the tropical, middle and extratropical WNP, respectively, as shown in Fig. 4a (also see Huang and Sun 1992; Kubota et al. 2015). The corresponding rainfall distribution is negative and positive rainfall anomalies in the tropical and middle WNP

(a) Sea level pressure
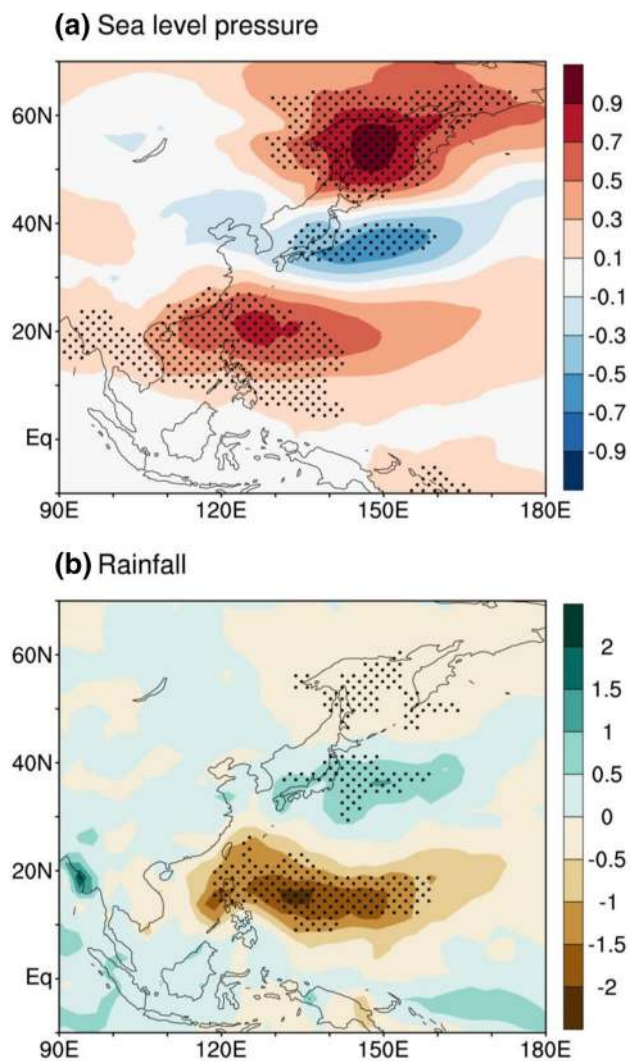
(b) Rainfall

Fig. 4 Structure of the Pacific-Japan (PJ) pattern. Show are **a** SLP (hPa) and **b** rainfall (mm/day) anomalies regressed against the leading principal component of JJA seasonal-mean zonal wind velocity at 850 hPa multiplied by the Coriolis parameter over the region (10°–55°N, 100°–160°E) from 1979 to 2014. Stippling indicates > the 95% confidence level, based on the Student's *t* test. The figure is replotted by using the data in Xie et al. (2016). The data were provided by Dr. Yu Kosaka

(Fig. 4b). Two mechanisms have been hypothesized to explain the PJ pattern's existence: (1) an internal mode of the atmosphere (Kosaka and Nakamura 2010; Hirota and Takahashi 2012) and (2) an SST-forced mode (Nitta 1987). The tropical lobe of the PJ pattern is an anticyclone in the tropical WNP, often observed during the peak and decayed phases of El Niño. The mechanisms, based on local SST-related feedback and remoted SST forcing from the Indian Ocean to explain the anticyclone in the tropical WNP, may thus work for the PJ pattern (e.g., Srinivas et al. 2018).

2.2 Oceanic pathways

2.2.1 Great ocean conveyor belt

The "Great Conveyor Belt" of Broecker (1987, 1991) emphasizes the interconnectedness of all oceans and the heat redistribution by ocean circulation. The global conveyor belt begins on the surface of the North Atlantic Ocean and is driven by ocean circulation called the thermohaline circulation in the Atlantic or the Atlantic meridional overturning circulation (AMOC) (Fig. 5). In the North Atlantic Ocean, the Gulf Stream transports the surface warm and saline water northward. As the surface warm and saline water moves northward, the ocean heats the atmosphere in the cold northern latitudes. This loss of heat to the atmosphere makes the water cooler and denser over the North Atlantic Ocean, causing it to sink to the bottom of the ocean. In the deep ocean of the Atlantic, water flows southward and enters the Southern Ocean and then reaches the Indian Ocean and the North Pacific Ocean at depth. The cold bottom water upwells to the surface in the Indian Ocean and the North Pacific Ocean. The upwelled surface water in the Pacific Ocean moves to the Indian Ocean through the Indonesian throughflow (ITF), the South China Sea throughflow (SCSTF) and the Tasman leakage (TL), and joins the upwelled water in the Indian Ocean and eventually returns to the North Atlantic Ocean via the Agulhas leakage (AL) where the cycle begins again to encircle the globe. A complete circuit of the global ocean conveyor belt takes a thousand or more years. However, the ITF, the SCSTF, the TL and the AL take a shorter time for the connection and transport between the Pacific Ocean and the Indian Ocean and between the Indian Ocean and the Atlantic Ocean.

2.2.2 Indonesian throughflow

The ITF transports the warm and fresh water in the western Pacific Ocean to the Indian Ocean. Many oceanic measurements were implemented at different channels of the ITF region during the past (e.g., Gordon et al. 2008; Sprintall et al. 2009). As reviewed by Feng et al. (2018), the estimate of the ITF transport is about 15 Sv (1 Sv = 10^6 m^3/s) although this number may be slightly higher than that from previous measurements (e.g., Gordon et al. 2008). Unlike the ITF, long-time direct measurements of the SCSTF do not exist. However, numerical models indicate that the SCSTF is important to the ITF and is probably the main reason for the ITF to form subsurface maximum velocity and to have seasonal variability (Tozuka et al. 2007, 2009; Wang et al. 2011a).

Atmospheric wind is an important factor to determine the strength of the ITF transport. Since the surface wind in the tropical Pacific and Indian Oceans shows different variations

Fig. 5 Global oceans linked by ocean circulation of "Great Conveyor Belt" named by Broecker (1987). The global conveyor belt begins on the surface of the North Atlantic Ocean and is driven by the Atlantic meridional overturning circulation (AMOC). The red color indicates surface warm water and blue color means deep cold water. See the main text for the Indonesian throughflow (ITF), the South China Sea throughflow (SCSTF), the Agulhas leakage (AL), and the Tasman leakage (TL). ACC stands for the Antarctic circumpolar current

Global Oceans Linked by Ocean Circulation
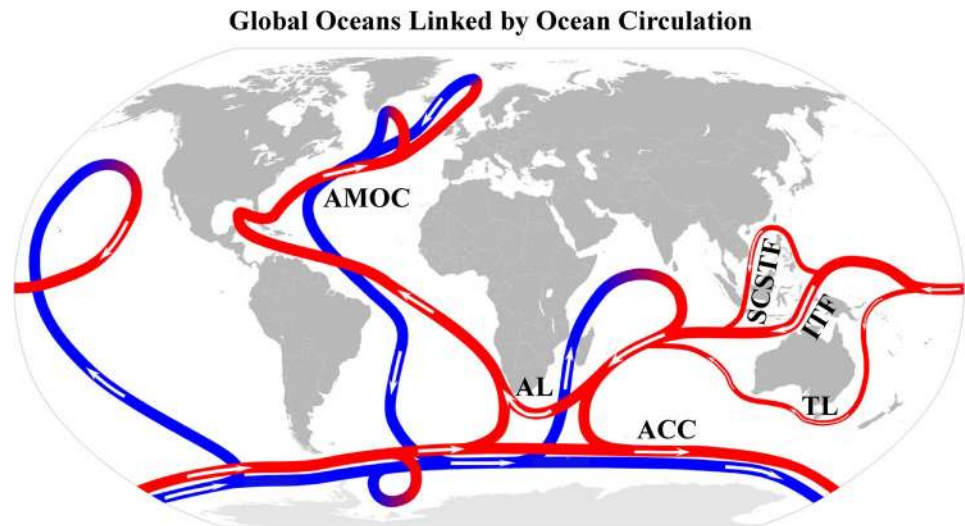
on different timescales, the ITF also experiences different variations on different timescales. Direct observations showed that on interannual timescales, ENSO largely affects the ITF transport. Because La Niña (El Niño) is associated with a strengthening (weakening) of the easterly trade winds in the Pacific, the ITF is strong (weak) during La Niña (El Niño) with an amplitude of about 5 Sv (e.g., Meyers 1996). Kelvin and Rossby waves, generated by remote zonal winds along the equator of both the Indian and Pacific Oceans, change seal level and thermocline (Clarke and Liu 1994; Wijffels and Meyers 2004). The ITF transport thus varies with the sea level variations induced by ENSO events. However, the ITF transport actually lags ENSO by 9 months due to the slow oceanic processes (England and Huang 2005). For the Indian Ocean's influence, a positive IOD event is associated with cold water and low sea level in the eastern Indian Ocean which enhances the ITF transport (Potemra and Schneider 2007; Sprintall and Révelard 2014; Liu et al. 2015). Because the positive IOD normally occurs during the fall of the El Niño developing year, the effects of the IOD and El Niño on the ITF transport actually offset each other. However, the IOD is short-lived and lasts only a few months during the fall of the El Niño developing year, so El Niño's influence on the ITF should dominate.

An example of Pacific ENSO's influence on the Indian Ocean via the ITF is the SST warming off the west coast of Australia which is named as the Ningaloo Niño by Feng et al. (2013). A Ningaloo Niño often occurs during a La Niña event of the Pacific when the easterly winds in the equatorial western Pacific strengthen the ITF and then drive a stronger southward Leeuwin current. The stronger southward oceanic current together with atmospheric teleconnection and local ocean–atmosphere processes lead the Ningaloo Niño to the peak phase (Kataoka et al. 2013; Marshall et al. 2015; Tozuka et al. 2014). Zhang and Han (2018) recently showed

that the Ningaloo Niño can in turn actively affect the tropical Pacific. In other words, a positive feedback loop between ENSO and the Ningaloo Niño may exist in association with inter-ocean interaction processes.

The wind variations in the tropical Pacific also drive ITF decadal transport. The decadal variations of the tropical Pacific and Indian Oceans are dominated by the Pacific decadal oscillation (PDO) because decadal variability in the Indian Ocean is weak (Han et al. 2014). Thus, the PDO plays an important role in decadal variations of the ITF transport (Zhuang et al. 2013). Like El Niño and La Niña events, the warm (cold) phase of the PDO corresponds to weaker (stronger) than the normal easterly trade winds. Therefore, the warm (cold) phase of the PDO weakens (strengthens) the ITF transport (Wainwright et al. 2008; Liu et al. 2010, 2015; Feng et al. 2011; Lee et al. 2015).

On longer or global warming timescales, the estimated ITF transport is based on numerical models because of the lack of long-term ITF observations. Feng et al. (2017) showed a weakening of the modeled ITF by 32% during the 21st century under global warming scenario, associated with reductions in net deep ocean upwelling in the tropical and South Pacific. The 32% reduction of the ITF in the model simulations is similar to the reduced percentage of the AMOC projected by CMIP5 climate models (Cheng et al. 2013; Reintges et al. 2017). The similar reductions of the ITF and the AMOC in the future model simulations may be an accident, but also may indicate that the ITF is linked with the AMOC on longer timescales.

2.2.3 Tasman leakage

In additional to the ITF, the Tasman leakage (TL) also connects the Pacific Ocean and the Indian Ocean (e.g., Rintoul and Sokolov 2001; van Sebille et al. 2014). The TL

is defined as all water which passes south of Australia on route from the subtropical Pacific Ocean to the subtropical Indian Ocean. The TL is fed by the east Australia current—the western boundary current of the southern Pacific Ocean. This southward flowing current bifurcates at approximately 35°S. Most of the water then flows eastward along the Tasman front with some water continuing southward, mostly in the form of eddies and filaments (Suthers et al. 2011). The TL is the part of this southward flowing water that flows around Tasmania and reaches the Indian Ocean. Compared to the ITF, the TL is far less studied and understood. Measurements and monitoring of the TL are difficult because the TL involves oceanic eddies and filaments. A recent study showed that the TL may carry as much as 10 Sv of Pacific water to the Indian Ocean (Qu et al. 2019). TL variations on various timescales and their relationships with climate variability need to be studied.

3 Influence of the Indian Ocean on Pacific and Atlantic climate

Not long ago, the tropical Indian Ocean was viewed as a non-interactive ocean in terms of ocean–atmosphere interaction and was often treated as a slaved ocean—ocean dynamics is unimportant. Recent studies, however, showed that this view and treatment are not true. Ocean–atmosphere feedback processes do operate in the Indian Ocean, and the thermocline and ocean current variations play important roles in the heat distribution of the upper layer ocean. Due to ocean–atmosphere interaction and ENSO's influence, the tropical Indian Ocean possesses two large-scale climate modes of the IOB and IOD. The variations of the Indian Ocean associated with these two modes can in turn affect Pacific and Atlantic climate via the atmospheric bridges and oceanic pathways.

3.1 Influence via the atmosphere

When an El Niño occurs, the equatorial eastern and central Pacific SST is warm. As shown and discussed in the last section, the warm Pacific SST alters the Walker circulation over the globe, resulting in anomalous descent motion in the equatorial Indian Ocean. The descent motion redistributes the surface wind in the Indian Ocean. The El Niño-related wind in the eastern Indian Ocean forces westward-propagating oceanic Rossby waves which warm the Indian Ocean. This ocean wave dynamics along with other processes and feedbacks such as the WES feedback cause a basin-wide warming of the IOB in the El Niño's subsequent spring (Fig. 2c, d). Figure 6a confirms that the positive and persistent correlation between the Nino3.4 and IOB indices occurs when the Nino3.4 index leads the
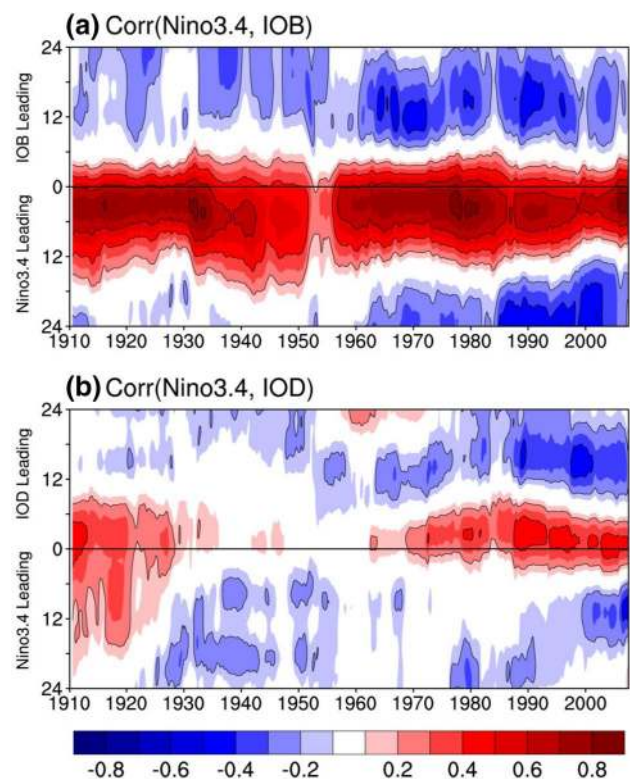
Fig. 6 The lead-lag correlations between ENSO and Indian climate. Shown are the 21-year sliding correlations **a** between the Nino3.4 and IOB indices and **b** between the Nino3.4 and IOD indices. The y-axis represents the lag and lead time in months. The ERSST data from 1900 to 2017 are used. The linear trends from 1900 to 2017 are removed before the calculation of correlations. The correlation is shaded when it exceeds the 95% confidence level, based on the Student's t-test

IOB about one season. Figure 6a also shows a significant but negative correlation with the IOB leading about one year, suggesting that the IOB may affect Pacific ENSO. Many studies have investigated the influence of the IOB on the Pacific (e.g., Watanabe and Jin 2002; Wu and Kirtman 2004; Annamalai et al. 2005; Kug et al. 2006; Xie et al. 2009, 2016; Du et al. 2009; Wu et al. 2009a, b, 2017; Li et al. 2017). The warming in the Indian Ocean can produce atmospheric Kelvin waves that propagate eastward and thus induce or increase equatorial easterly wind anomalies in the western Pacific. In fact, the correlation between the IOB index and zonal wind anomalies in the Nino5 region of the equatorial western Pacific (120°–140°E, 5°S–5°N) does show a negative correlation with the IOB leading about 2–3 months after the 1950s (Fig. 7a) [The ENSO indices in the western Pacific were defined in Wang et al. (1999)]. This whole process about the effect of the Indian Ocean's influence on the Pacific is called "Indian Ocean capacitor effect" (Xie et al. 2009), which is operated through the atmosphere. Xie et al. (2016) and Li et al.
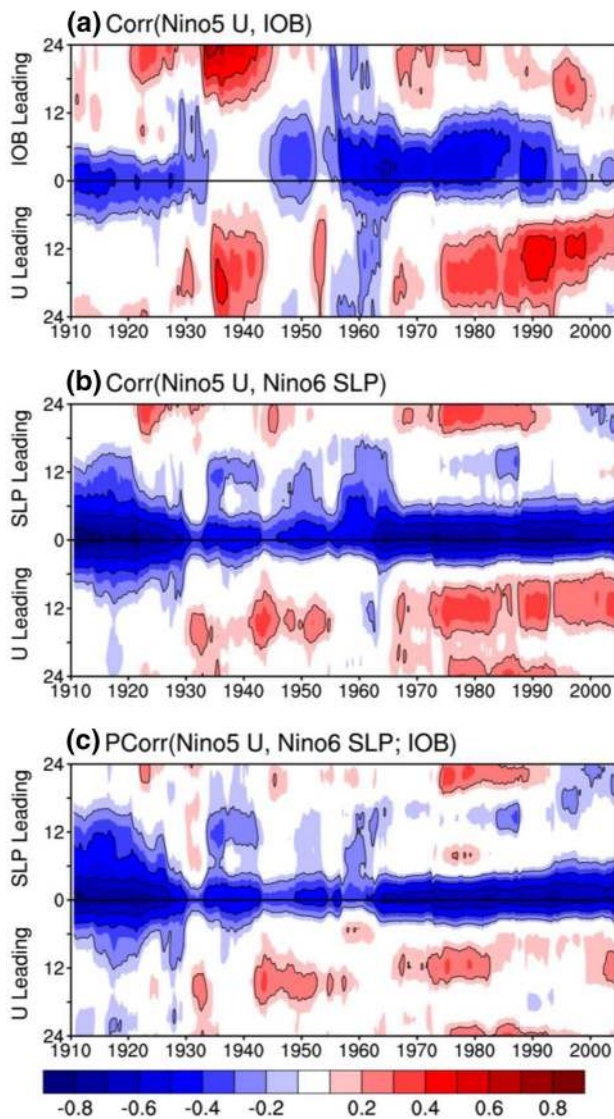
Fig. 7 The lead-lag correlations of Nino5 wind with the IOB and Nino6 SLP. Shown are the 21-year sliding correlations **a** between the IOB index and Nino5 surface zonal wind anomalies, **b** between the Nino6 SLP and Nino5 surface zonal wind anomalies, and **c** partial correlation between the Nino6 SLP and Nino5 surface zonal wind anomalies by removing the effect of the IOB. The Nino5 region is defined the equatorial western Pacific over 120°–140°E, 5°S–5°N and the Nino6 region is in the off-equatorial western Pacific over 140°–160°E, 8°–16°N (Wang et al. 1999). The y-axis represents the lag and lead time in months. The ERSST and 20th century reanalysis from 1900 to 2014 are used. The linear trends from 1900 to 2014 are removed before the calculation of correlations. The correlation is shaded when it exceeds the 95% confidence level, based on the Student's t-test

et al. 1999; Wang and Weisberg 2000). When an El Niño event occurs, the positive SST anomalies in the equatorial eastern Pacific produce a pair of anomalous cyclones in the off-equatorial region and westerly wind anomalies in the equatorial region (Gill 1980). Associated with the pair of anomalous cyclones is the off-equatorial wind stress curl which shallows the thermocline by Ekman pumping and produces westward propagating oceanic Rossby waves. These oceanic processes raise the thermocline in the Nino6 region of the off-equatorial western Pacific (140°–160°E, 8°–16°N), consistent with the large observed variation in the thermocline depth there. Although the mean thermocline in the western Pacific is relatively deep compared with that in the eastern Pacific, the shallow off-equatorial thermocline variation can initiate the cold SST anomalies in the Nino6 region (Wang et al. 1999). The cold SST anomalies result in the high Nino6 SLP anomalies which produce the Nino5 equatorial easterly wind anomalies. Later on, Wang et al. (2000) suggested that the local WES feedback further develops and maintains the cold SST anomalies and anomalous anticyclone in the WNP. Figure 7b shows a high correlation between the Nino6 SLP anomalies and the Nino5 zonal wind anomalies at zero lag. Partial correlation by removing the effect of the IOB in Fig. 7c shows a similar correlation pattern with that in Fig. 7b in spite of a little bit reduction in correlation amplitude. This suggests that the equatorial easterly wind anomalies in the western Pacific are mostly initiated and induced by local processes associated with ENSO event. The Indian Ocean capacitor effect may maintain and persist the equatorial easterly wind anomalies and anticyclone in the WNP during the decayed phase of El Niño.

In addition to the IOB, the Indian Ocean also hosts the IOD mode (e.g., Saji et al. 1999; Webster et al. 1999) which is caused by either ENSO or local ocean–atmosphere interaction. The 21-year sliding correlation between the IOD and Niño3.4 indices in Fig. 6b shows a positive correlation with the IOD leading about 3 months after the 1970s. This reflects the fact that IOD events occur and peak in the fall of the ENSO developing year and but ENSO events mature in the following winter. After the 1950s, a negative correlation with the IOD leading more than 1 year suggests that IOD events may cause ENSO events in the tropical Pacific with a time delay. Indeed, Izumo et al. (2010, 2014) showed that the negative (positive) phase of the IOD can cause El Niño (La Niña). During a negative IOD event, the southeast Indian Ocean experiences warm SST anomalies which peak in the fall. The warming of the southeast Indian Ocean produces the easterly wind anomalies in the western Pacific. These easterly anomalies help accumulate warm water in the western Pacific and serve as a precondition for the occurrence of El Niño. After the fall, the eastern pole of the IOD quickly collapses. The quick demise of the IOD anomaly in the winter then induces a sudden collapse of the easterly
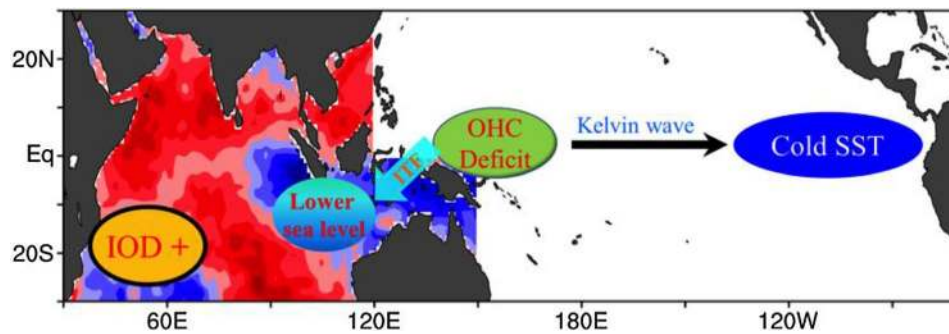
(2017) provided detailed reviews of the Indian Ocean capacitor effect.

However, it is worthy of noting that the Nino5 wind anomalies during ENSO are initiated by and associated with oceanic processes (Weisberg and Wang 1997; Wang

Fig. 8 Schematic diagram of Indian Ocean's influence on the Pacific through the ITF. A positive IOD event in the Indian Ocean is associated with lower sea level in the eastern Indian Ocean. The lower sea level in the eastern Indian Ocean increases the ITF transport which decreases the ocean heat content (OHC) in the western Pacific Ocean. The OHC deficit in the western Pacific Ocean propagates eastward as an upwelling Kelvin wave, terminating El Niño or initiating La Niña in the tropical Pacific

wind anomalies over the Pacific Ocean, which leads to the development of El Niño. The reverse is true for the positive phase of the IOD.

The influence of the Indian Ocean on Atlantic climate has not been well studied. By performing a set of numerical model experiments with an atmospheric general circulation model, Bader and Latif (2003) showed that the warming of the tropical Indian Ocean may have contributed to the strengthening of the North Atlantic oscillation and may also have played an important role in driving the observed decadal drying trend over the West Sahel. However, they did not provide the detailed mechanisms for these remote influences of the Indian Ocean warming. In other words, it remains unclear how the Indian Ocean affects Atlantic climate although the Indian Ocean-induced atmospheric Rossby waves and/or atmospheric circulations may be candidates.

3.2 Influence via the oceans

The influence of the Indian Ocean on Pacific climate can also operate through the ITF. Based on observational data, Yuan et al. (2013) proposed that a positive IOD event might help cool the central and eastern Pacific Ocean 1 year later. Given a positive IOD event, sea level in the eastern Indian Ocean is lower than the normal (Fig. 8). The lower sea level in the eastern Indian Ocean establishes a southwestward pressure gradient force in the ITF region which increases the ITF transport from the western Pacific Ocean to the Indian Ocean. Because of the increased ITF transport, the ocean heat content (OHC) in the upper layer of the western Pacific Ocean is decreased. Yuan et al. (2013) argued that the OHC deficit in the western Pacific Ocean propagates eastward as an upwelling oceanic Kelvin wave, terminating El Niño or initiating La Niña in the tropical eastern Pacific. This hypothesis provides a new idea for the Indian Ocean to affect Pacific climate via the ITF. However, the detailed processes and mechanisms

need to be further tested and studied. In particular, the upwelling oceanic Kelvin wave is not necessarily induced by and associated with the ITF-induced OHC deficit. As shown previously (Wang et al. 1999; Boulanger and Menkes 2001; Boulanger et al. 2003), the upwelling oceanic Kelvin wave can be forced by easterly wind anomalies in the equatorial western Pacific which are locally induced by the anticyclone in the WNP and/or are enhanced by the IOB (see Fig. 7).

The Indian Ocean can influence the Atlantic Ocean by the westward Agulhas current and its leakage in the south of Africa. The Agulhas current, which is driven by wind stress curl between the southeast trade winds and the westerlies in the southern Indian Ocean, flows toward the southwest as a western boundary current (Lutjeharms 2006). When the Agulhas current reaches south Africa, it loops anticlockwise and returns back as the Agulhas retroflection (Gordon 2003). The Agulhas retroflection sheds rings, eddies and filaments to the west, which represents a leakage of the Indian Ocean water into the Atlantic Ocean (Gordon et al. 1992; Schouten et al. 2000; Boebel et al. 2003). Direct continuous measurements of the Agulhas leakage (AL) are very difficult because of its intermittent nature and variable location. On interannual timescales, the IOD may modulate the Agulhas current and its leakage. A negative (positive) IOD strengthens (weakens) the tropical and subtropical gyres of the Southern Indian Ocean, as observed in satellite data (Schouten et al. 2002; Palastanga et al. 2006). These gyre changes increase or decrease the sources of the Agulhas current, affecting the frequency of eddy shedding and thus the Indian Ocean's influence. On longer timescales, a trend of increasing AL is found to relate to a poleward shift in the westerly winds over the past three decades (Biastoch et al. 2009b; Rouault et al. 2009). This feature is projected to continue over the 21st century under global warming (Sen Gupta et al. 2009; Cai 2006). The study of the Agulhas current and its leakage is important since they may be linked to the AMOC and global

Global SST biases averaged CMIP5 models
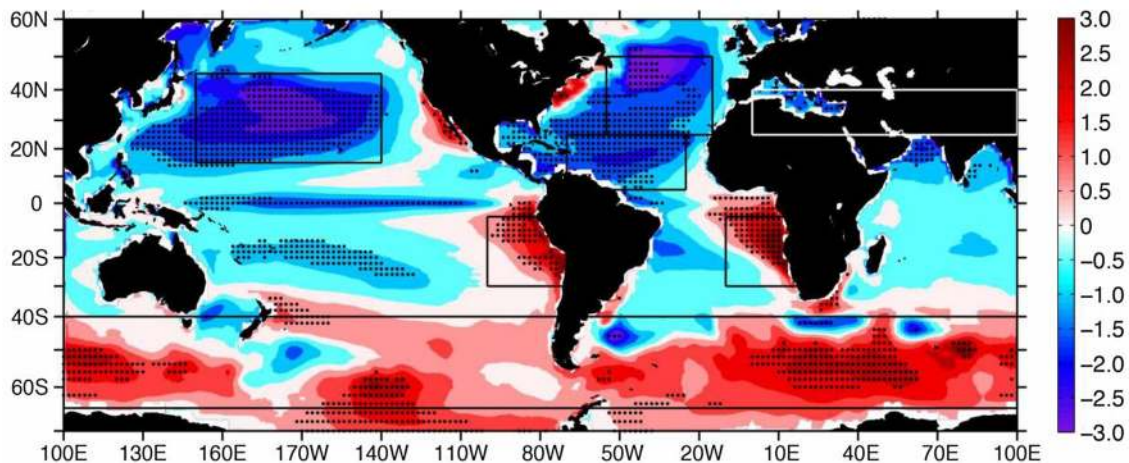
Fig. 9 Global SST bias averaged in 22 climate models. The annual-mean SST (°C) bias is calculated by the SST difference between the model SST and ERSST. The dots denote where at least 18 of 22 models (82%) have the same sign in the SST bias. The boxes represent the focused regions in Wang et al. (2014)

climate. For the detail of the Agulhas current and its leakage, see Biastoch et al. (2008, 2009a, b); Beal and Elipot (2016).

4 Influence of the Atlantic Ocean on Pacific and Indian climate

In spite of its small size compared to the Pacific Ocean, the Atlantic Ocean shows many important climate phenomena on various timescales. These include the Atlantic Niño, the tropical Atlantic meridional mode, the Atlantic warm pool (AWP), the Atlantic multidecadal oscillation (AMO) and the AMOC. These climate phenomena manifest in SST variations in different regions of the Atlantic Ocean, affecting Pacific and Atlantic climate. Their influences are thus summarized based on the regions of the Atlantic. The influences are presented in the order of the midlatitude North Atlantic, the tropical North Atlantic and the equatorial Atlantic.

4.1 The midlatitude North Atlantic

4.1.1 Influence on North Pacific climate

The AMO represents the basin-wide SST variations in the North Atlantic north of the equator, which seems to be related to AMOC variability (e.g., Delworth and Mann 2000; Knight et al. 2005). Zhang and Delworth (2007) have demonstrated that the AMO in the North Atlantic can affect North Pacific variability such as the PDO. The influence of the AMO on Pacific variations is also evidenced in

the North Atlantic waterhosing experiments (Stouffer et al. 2006). The waterhosing experiment with a 1 Sv freshwater forcing in the high-latitude North Atlantic showed that the AMOC quickly slows down, resulting in a cooling of SST in the North Atlantic and a southward displacement of the ITCZ due to a change of the oceanic heat transport (e.g., Wu et al. 2008). With the AMOC slow down or shut down, the Pacific also shows a robust response with the cooling of the North Pacific Ocean, the intensification of the westerly winds, and the deepening of the wintertime Aleutian low (e.g., Timmermann et al. 2007).

The influence of the North Atlantic on the North Pacific is also demonstrated in a study of assessing global model SST biases (Wang et al. 2014). In the Northern Hemisphere, almost all of climate models participating in the Coupled Model Intercomparison Project phase 5 (CMIP5) display cold SST biases in both the North Atlantic and the North Pacific (Fig. 9). CMIP5 models show that the simulated AMOC is highly related to the cold biases in the North Atlantic and North Pacific. Models with a weak AMOC show a cold SST bias in the North Atlantic because of the weak ocean heat transport, which is in agreement with the model SST response to the North Atlantic freshwater forcing (Zhang and Delworth 2005; Wu et al. 2008). The cold SST bias in the North Atlantic affects the cold SST bias in the North Pacific through the mechanism of the atmospheric "annular" mode, as demonstrated in Wang et al. (2014). In other words, the North Atlantic SST bias in Fig. 9, which is related to the weak simulated AMOC, can remotely cause the North Pacific SST bias.

4.1.2 AMO's impact on CP-type El Niño

Yu et al. (2015) used observational data and numerical model experiments to demonstrate that a warm phase of the AMO can teleconnect to the subtropical Pacific region and then cause the central Pacific (CP)-type El Niño events. It is shown that the positive North Atlantic SST anomalies during the warm phase of the AMO intensify the Pacific subtropical high which is associated with stronger background easterly trade winds. Because of the enhanced WES feedback mechanism (Xie and Philander 1994), the ocean–atmosphere coupling is strengthened in the subtropical Pacific. The ocean–atmosphere interaction processes ultimately result in interannual SST variability in the equatorial central Pacific. It is thus concluded that the warm phase of the AMO in the North Atlantic is favorable for the occurrence of the CP-type El Niño events in the tropical Pacific.

The study of Yu et al. (2015) explains the shift to more frequent occurrences of CP-type El Niño events in the late 1990s. The AMO transitions from the cold to warm phases around 1995, which is consistent with the increased occurrence of CP-type El Niño events in the recent decades. This study provides an alternative explanation to Yeh et al.'s (2009) argument that global warming increases the CP-type El Niño events by using CMIP3 climate model outputs. However, Yeh et al. (2009)'s conclusion cannot be obtained or confirmed by CMIP5 climate models (Power et al. 2013; Xu et al. 2017). Additionally, the argument of global warming cannot explain why the increase of the CP-type El Niño events occurs only after the late 1990s because global warming takes place before the 1990s. All of these indicate that global warming may not be a major cause of the recent increase of the CP-type El Niño events.

4.1.3 AMO's impact on the tropical WNP

The AMO can remotely affect the multidecadal variability of atmospheric circulation in the tropical WNP (Sun et al. 2017). The warm phase of the AMO induces local anomalous ascent in the North Atlantic, with anomalous divergence in the upper troposphere. The upper level divergence flows toward the North Pacific, resulting in anomalous subsidence and an anomalous high in the North Pacific. The anomalous high weakens the Aleutian low and the associated easterly wind anomalies decrease the mean westerly wind in the subtropical North Pacific. The decreased wind speed decreases the latent heat flux loss from the ocean, thus warms subtropical Pacific SST. The atmospheric surface wind response to the SST warming is to converge toward the warm SST center from the tropics, which produces an anomalous low pressure and cyclonic circulation in the tropical WNP. At the same time, the SST-cloud-longwave radiation positive feedback, proposed by Wang and Enfield (2003), also helps to develop and enhance warm SST anomalies associated with the anomalous cyclonic circulation in the tropical WNP.

Anomalous cyclonic and anticyclonic circulations in the tropical WNP are important for climate variability in Asia because they can change moisture transport from the ocean to southeastern Asia. The AMO-induced cyclonic circulation in the tropical WNP provides a mechanism for explaining the AMO's linkage with and impact on Asian climate. However, Sun et al. (2017) did not mention and discuss whether the cold phase of the AMO is able to produce an anomalous anticyclonic circulation in the tropical WNP.

4.1.4 Initiating a Pacific El Niño event

Sections 4.1.1–4.1.3 just review the influences of the North Atlantic on the Pacific via processes that extend westward. The remote influences of the North Atlantic Ocean on the Pacific can also be through a process or mechanism extending eastward across Eurasia and Asia. The influence is seen by the fact that the correlation between the summer North Atlantic SST anomalies and ENSO index is negatively significant when the North Atlantic SST anomalies lead the ENSO index by 5–13 months (Wang et al. 2009b, 2011b). This implies that a cold North Atlantic Ocean in the summer is possibly followed by the onset of an El Niño event in the subsequent spring in the tropical Pacific. The mechanism for the linkage of the North Atlantic SST with El Niño is through atmospheric teleconnection pattern which is similar to the East Atlantic/West Russia teleconnection (Barnston and Livezey 1987). The North Atlantic SST anomalies produce a wave train that is characterized by a cyclone in the North Atlantic, an anticyclone near England, and a cyclone near the Caspian Sea, and an anticyclone in Lake Baikal (Fig. 10). The anticyclone near Lake Baikal strengthens the continental northerlies and enhances the East-Asian winter monsoon. These changes in the winter are followed by anomalous cyclone in the off-equatorial western Pacific during the subsequent spring. The cyclone induces westerly wind anomalies in the equatorial western Pacific which can help to initiate a Pacific El Niño event. That is, an El Niño event is likely to be initiated during the spring after a cold North Atlantic Ocean occurs in the preceding summer.

Atmospheric general circulation model experiments were performed to confirm the data-derived influence of the North Atlantic SST on the Pacific (Wang et al. 2013). The model experiments showed that the cold SST anomalies in the North Atlantic produce a Rossby wave-like teleconnection pattern which strengthens the Siberian High in the winter and subsequent spring. The stronger Siberian High increases the continental northerlies over East Asia and is associated with the anomalous cyclone over the WNP, which is favorable for the occurrence of the westerly wind bursts. The stronger northerlies over East Asia can also induce frequent

East Atlantic/West Russia pattern
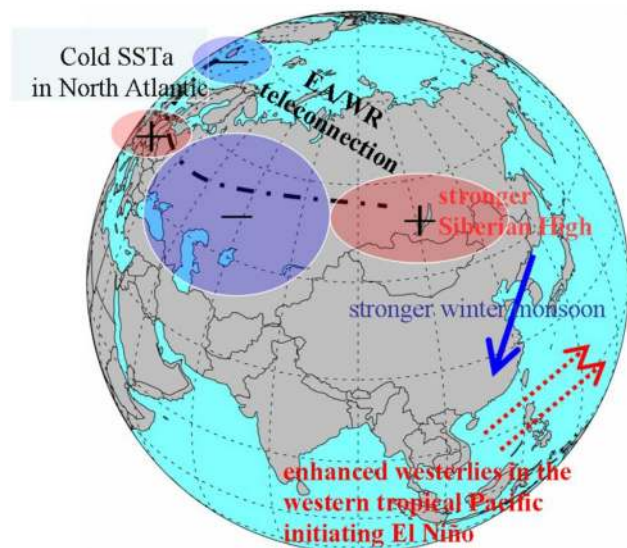
Fig. 10 Schematic diagram of the East Atlantic/West Russia (EA/WR) teleconnection pattern. A cold North Atlantic Ocean corresponds to the EA/WR teleconnection pattern, featuring two anticyclonic circulations (denoted by "+" sign) near England and Lake Baikal, and two cyclonic circulations (denoted by "−" sign) over the North Atlantic and near the Caspian Sea

cold surges, which tend to produce the equatorial westerly wind anomalies in the western Pacific. The westerly wind bursts and the westerly wind variations help initiate an El Niño event in the tropical Pacific. However, the results of the atmospheric general circulation model need to be confirmed and further studied, especially using coupled ocean–atmosphere-land models.

4.2 The tropical North Atlantic

The influence of the tropical North Atlantic on the tropical Pacific may operate via two routes of: (1) the tropical northeast Pacific (Rong et al. 2010; Ham et al. 2013; Ding et al. 2017; Wang et al. 2017) and (2) the tropical southeast Pacific (Wang et al. 2006, 2010, 2014). The earliest study by Rong et al. (2010) showed that tropical North Atlantic SST anomalies can produce an anomalous anticyclone over the tropical WNP which is an important component of El Niño events and a bridge linking El Niño events with Asian climate. Then, several recent studies further demonstrated that the positive SST anomalies in the tropical North Atlantic in the spring can induce a CP-type Pacific La Niña in the subsequent winter. Based on Gill's (1980) atmosphere, warm SST anomalies in the tropical North Atlantic during the boreal spring induce a pair of low-level anomalous cyclones in the eastern Pacific as a result of atmospheric Rossby wave response. The northern cyclone is located over

the tropical northeast Pacific. In the west of this cyclone, the ocean becomes cooling due to the cold meridional advection and the increased latent heat flux. The cool SST is further developed, and produces a low-level anticyclonic flow over the tropical WNP during the following months. This flow generates equatorial easterly wind anomalies over the western and central Pacific that cool the equatorial Pacific and may trigger a CP-type La Niña event in the following winter.

The second route is through an inter-hemispheric influence on the tropical southeast Pacific (Wang et al. 2006, 2010, 2014). Seasonally, the AWP is the heating source for the Hadley circulation during the boreal summer in the Western Hemisphere. The AWP in the summer induces ascending motion which crosses the equator in the upper troposphere and flows toward and descends over the tropical southeast Pacific. The air flows equatorward in the lower troposphere in the tropical southeast Pacific, resulting in the southeasterly wind anomalies. Interannually, when the AWP is anomalously large in the summer/fall, the regional Hadley circulation is strengthened. This anomalous meridional circulation and the associated subsidence strengthen the South Pacific subtropical anticyclone, and in turn lead to an enhancement of low cloud, a strengthening of the southeasterly trade winds and thus a decrease of the SST. This indicates that a large AWP can lead to a cold tropical Pacific Ocean or a CP-type La Niña event via the Hadley circulation change.

4.3 The equatorial Atlantic

The Bjerknes feedback also operates in the equatorial Atlantic and forms the Atlantic Niño, as in the equatorial Pacific for the El Niño. The Pacific El Niño can influence the Atlantic Niño, but depends on the pre-conditioning of the Atlantic (Chang et al. 2006). The Atlantic Niño also affects the Pacific El Niño via the change of the Atlantic Walker circulation (Wang 2006). The Pacific El Niño does not simultaneously correlate with the Atlantic Niño because of the slow oceanic adjustment processes and intervening of local processes. However, anomalous warming or cooling of the two equatorial oceans can alter the Walker circulation and influence each other. In other words, warming or cooling in the equatorial Pacific and Atlantic forms an inter-basin SST gradient variability that produces surface zonal wind anomalies over equatorial South America and over some regions of both the ocean basins in association with the anomalous Walker circulation. These, along with local oceanic and atmospheric processes, further enhance equatorial warming or cooling and thus reinforce the inter-Pacific-Atlantic SST gradient. Through the inter-basin SST gradient which is associated with the overlying Walker circulation, the Atlantic Niño is capable of affecting the equatorial Pacific.

The influence of the Atlantic Niño on ENSO variability via the Walker circulation was then investigated by many researchers (e.g., Jansen et al. 2009; Frauen and Dommenget 2012; Rodríguez-Fonseca et al. 2009; Ding et al. 2012; Kucharski et al. 2011, 2015; Polo et al. 2015). The Atlantic Niño peaks in the summer which can alter the tropical atmospheric circulation favoring the development of Pacific La Niña in the following winter. The mechanism involves the Atlantic Walker circulation change induced by the Atlantic Niño, featured with anomalous ascent over the Atlantic and anomalous descent over the central Pacific. As a result, easterly surface wind anomalies in the central Pacific pile up warm water in the western Pacific, and shallow the thermocline and cool SST in the eastern Pacific. Because of slow oceanic adjustment processes, an Atlantic Niño in the summer is able to produce an EP-type La Niña in the tropical Pacific in the subsequent winter. Similarly, an Atlantic Niña can subsequently induce an EP-type Pacific El Niño.

The Atlantic Niño also influences the Indian Ocean, inducing a warming in the tropical western Indian Ocean and a weakening of Indian summer monsoon rainfall (Kucharski et al. 2008, 2009; Wang et al. 2009a). The warm equatorial Atlantic SST anomalies of the Atlantic Niño produce an atmospheric Kelvin wave which propagates eastward to the Indian Ocean in terms of Gill's (1980) physics. Associated with the Kelvin wave are surface easterly wind anomalies in the tropical western Indian Ocean. The easterly wind anomalies decrease evaporation and weaken the coastal upwelling, all of which can warm the SST in the tropical western Indian Ocean. The warming in the tropical western Indian Ocean, assisted with the Bjerknes feedback, may help to develop a positive IOD event in the Indian Ocean. However, the processes for an Atlantic Niño to induce a positive IOD need to be further studied and confirmed. Additionally, the Indian summer monsoon is associated with a mean southwesterly wind in the northern Indian Ocean. Because the Atlantic Niño induces the easterly wind anomalies over the Indian Ocean, the easterly wind anomalies reduce the mean southwesterly wind and thus can weaken the Indian summer monsoon. In short, an Atlantic Niño is able to induce a positive IOD event and can also weaken the Indian summer monsoon.

4.4 Long-term tropical Atlantic warming

During the past decades, the global tropical oceans are characterized by the warming in the Atlantic, the cooling in the eastern Pacific, and the warming in the Indo-western Pacific. This long-term warming-cooling trend pattern is attributed to the warming in the tropical Atlantic (Li et al. 2016). The processes of the tropical Atlantic warming-induced other ocean's cooling and warming have two steps. The first step is through the Gill's response and the WES feedback. Tropical

Atlantic warming generates anomalous atmospheric deep convection, with Kelvin wave propagating eastward and Rossby propagating westward, as Gill's model showed. The Kelvin wave-induced easterly wind anomalies over the Indian Ocean reduce surface wind speed and suppress evaporation, which warms the equatorial-northern Indian Ocean through the WES feedback. The response of the Pacific to the tropical Atlantic warming is Rossby wave-induced cooling in the off-equatorial eastern Pacific. This off-equatorial cooling propagates equatorward and westward via the WES mechanism (Xie 1999), resulting in the cooling in the central and eastern Pacific. The warming in the western Pacific and the cooling in the eastern Pacific form an SST west–east gradient over the tropical Pacific.

The second step involves the Bjerknes feedback with the Pacific Ocean dynamics after the SST west–east gradient over the tropical Pacific is established in the first step. The SST west–east gradient strengthens the easterly trade winds in the tropical Pacific which change oceanic circulation and thermocline and increase ocean upwelling in the equatorial eastern Pacific. These ocean dynamical processes further cool the cold tongue in the eastern Pacific and warm the warm pool in the western Pacific. These processes and the Bjerknes feedback ultimately produce the dipole-like SST trend pattern of the cooling in the eastern Pacific and the warming in the Atlantic and Indo-western Pacific during the past decades which is associated with the "global warming hiatus" in the literature.

In summary, the spatially and temporally varied SSTs in the Atlantic have rich influences on the Pacific and Indian Oceans. First, the AMO may cause North Pacific variability, the occurrence of CP-type El Niño and anomalous cyclonic circulation in the tropical WNP; and cold midlatitude North Atlantic SST anomalies may initiate an El Niño. Second, a warm tropical North Atlantic can induce a CP-type La Niña. Third, the Atlantic Niño (Niña) helps induce an EP-type La Niña (El Niño). The Atlantic Niño can also warm the tropical western Indian Ocean and weaken Indian monsoon rainfall. Finally, the tropical Atlantic warming during the past decades is attributed to the cooling in the tropical eastern Pacific and the warming in the Indo-western Pacific. These Atlantic influences look complicated, and some seem mutually contradictory. For example, enhancing CP-type El Niño events by the AMO warm phase, inducing a CP-type La Niña by the warm tropical North Atlantic, and initiating an El Niño by the cold midlatitude North Atlantic SST anomalies seem to be inconsistent. They seem to be inconsistent because the AMO warm phase is associated with the warming of the entire North Atlantic although the maximum SST anomalies are in the high latitudes. Ruprich-Robert et al. (2017) used numerical models to show that most of the simulated global-scale impacts of the AMO are driven by the tropical part of the AMO. However, we should keep
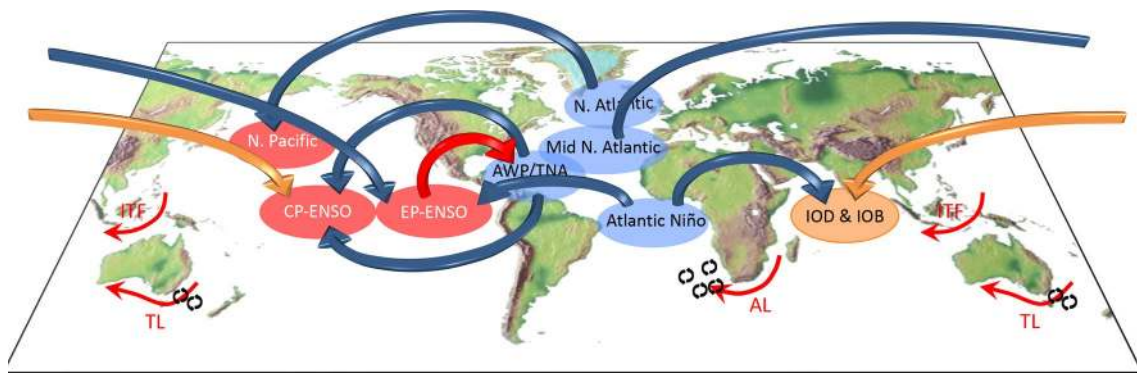
Fig. 11 Schematic diagram summarizing three-ocean interaction processes. Atmospheric bridge processes are represented by thick bending arrows. One-way bending arrow indicates one-way influenced process. Two-way bending arrow means that two oceans can affect each other. For example, ENSO can affect the IOD/IOB and the IOD/IOB can in turn influence ENSO events. The oceanic pathways of inter-ocean interactions are the Indonesian throughflow (ITF), the Agulhas leakage (AL), and the Tasman leakage (TL), which are drawn and shown by red arrows. AWP and TNA stand for the Atlantic warm pool and the tropical North Atlantic, respectively

in mind that the Atlantic influences summarized in this section are on different temporal and spatial scales. First, the AMO varies on multidecadal timescales, and is a basin-wide mode with the maximum warming in the high latitudes. The occurrence of CP-type El Niño is increased during the multidecadal period of the AMO warm phase. Second, the warm tropical North Atlantic occurs in the spring and its influence of inducing a CP-type La Niña is on interannual timescales. Third, the cold midlatitude North Atlantic SST anomalies are in the summer, and its influence of initiating an El Niño event is also on interannual timescales. If all these temporal and spatial differences are considered, it may not be surprising to see these different Atlantic influences. Nevertheless, the Atlantic influences on the other two oceans need to be studied systematically and integratedly.

5 Perspective and future work

Treating climate phenomenon as a coupled ocean–atmosphere problem has mostly begun since the 1980s with the original focus on ENSO. During the past several decades, the word "climate" was almost a synonym of ocean–atmosphere interaction. Indeed, the concept of ocean–atmosphere interaction has revealed and uncovered many puzzles of oceanic and atmospheric variations over the global oceans. However, focusing on ocean–atmosphere interaction is not the whole story of climate and is not enough for the study of climate. As reviewed in Sects. 2–4, interactions among the three oceans can play an important role in initiating and modulating climate phenomena over all the three oceans. Figure 11 presents inter-ocean interaction processes summarized and reviewed in this paper. The dominant processes involved three-ocean interactions are probably atmospheric bridges as shown in Fig. 11. However, the oceanic pathways

of the Indonesian throughflow (ITF), the Tasman leakage (TL) and the Agulhas leakage (AL) are also important for connecting climate variability over the three oceans.

The ultimate goal of climatic research, observational and modeling is to predict climate and serve the society. In spite of some progress and success in climate prediction, we have more work to do for achieving this goal. A good example is ENSO prediction which is regularly updated by IRI/CPC (at https://iri.columbia.edu/our-expertise/climate/forecasts/enso/current/). All of ENSO forecast models in the world in the spring of 2014 predicted that an El Niño was coming, but El Niño did not occur in 2014. In the spring of 2015, almost all models predicted no El Niño, yet a strong El Niño appeared in 2015. Sections 3–4 showed and discussed that both Atlantic and Indian Ocean variations can induce and/or modulate ENSO events, in addition to the local ocean–atmosphere feedback of Bjerknes type in the tropical Pacific. The deficient representation of inter-ocean interactions and the influence on ENSO in climate models may be one of the reasons for causing ENSO prediction uncertainty. Analyzing and studying inter-ocean interaction processes in these climate prediction models is a necessary step to improve the ENSO forecast.

Another issue is climate model biases. As shown in Fig. 9, most of coupled ocean–atmosphere models in the world suffer from large and common biases, with the average SST biases from − 3 °C to + 3 °C. Thus, the current climate models are far from meeting the requirements of our climate research and prediction, and need to be improved. As reviewed in Sect. 4, inter-ocean interaction processes can cause the global SST biases although local oceanic and atmospheric processes in individual oceans are also responsible for model biases. Improving climate models cannot be reduced to improved representation of regional processes. A better understanding of the global teleconnections or

inter-ocean interactions is needed for improving climate model performance.

Other questions and issues in the topic of inter-ocean interactions and climate that need to be addressed and answered are:

- What are the detailed mechanisms and processes of inter-ocean interactions?
- How do inter-ocean interactions affect climate variability and climate changes?
- Do inter-ocean interactions change or modulate our previous views of climate and climate changes?
- Do CMIP5/6 climate models have ability to simulate mechanisms and processes of inter-ocean interactions? If not, why? Can we fix the problems?
- How can climate predictability and prediction be improved by considering inter-ocean interactions?
- Can inter-ocean interactions improve our understanding of climate's impacts on extreme weather events?

It is hoped that we will have answers or partial answers for these questions in the near future.

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