Remote Ocean Forcing on Interannual-to-Decadal Climate Variability through Inter-Basin Interactions

Zachary F. Johnson
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REMOTE OCEAN FORCING ON INTERANNUAL-TO-DECADAL CLIMATE VARIABILITY THROUGH INTER-BASIN INTERACTIONS

by

Zachary F. Johnson

A dissertation submitted in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY in Climate Science

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UTAH STATE UNIVERSITY
Logan, Utah
2021
ABSTRACT

Remote ocean forcing on interannual-to-decadal climate variability through inter-basin interactions

by

Zachary F. Johnson, Doctor of Philosophy
Utah State University, 2021

Major Professor: Yoshimitsu Chikamoto, Ph.D.
Department: Plants, Soils and Climate

This dissertation explores interannual-to-decadal climate variability through the perspective of inter-basin interactions between the Pacific, Atlantic, and the Indian Oceans. The El Niño Southern Oscillation (ENSO) in the tropical Pacific can remotely influence sea surface temperatures (SSTs) in other ocean basins through atmospheric teleconnections. However, the Indian and Atlantic Oceans can feedback on tropical Pacific climate variability, highlighting the research area of inter-basin interactions. The mechanisms to describe these inter-basin interactions are an ongoing debate. To identify each ocean basin’s remote impact onto tropical Pacific climate variability, a suite of climate model experiments is conducted by incorporating observed ocean information into fully coupled climate models.

The dissertation is split into three manuscripts. After introducing the dissertation in Chapter 1, Chapter 2 examines the Atlantic and Pacific impact on Australian precipitation variability through inter-basin interactions. A significant Atlantic component of Australian precipitation variability was found through a modulation of the global Walker circulation. Chapter 3 investigates the remote Atlantic and tropical Pacific impacts on the Pacific Decadal Oscillation (PDO). Two pathways from the Atlantic to the Pacific are proposed to explain PDO variability, one at the equator and one in the north tropics. Chapter
4 explores how precipitation trends in the Indo-Pacific sector respond to the combination of external radiative, tropical Pacific, and Indian Ocean forcings from the 1980s through the 2010s. The results in this dissertation show that the three tropical oceans are more tightly connected than previously thought.
PUBLIC ABSTRACT

Remote ocean forcing on interannual-to-decadal climate variability through inter-basin interactions

Zachary F. Johnson

This dissertation explores the connection between ocean basins through the atmosphere by employing observational data analyses and a climate modeling approach. Sea surface temperature changes in the tropical Pacific, known as the El Niño Southern Oscillation, can influence worldwide weather and sea surface temperatures in other ocean basins. For instance, tropical Pacific sea surface temperatures can impact the Atlantic and Indian Oceans through airflow changes along the equator. However, Atlantic and Indian Ocean sea surface temperature changes can also influence the tropical Pacific through similar processes. Therefore, it is challenging to identify the mechanisms of these remote connections between ocean basins due to two-way interconnections. To better understand how ocean basins are connected by the atmosphere, this dissertation uses climate models to simulate the climate response from ocean forcing. Results in this dissertation show that the three tropical oceans are more tightly connected than previously thought. After introducing the dissertation in Chapter 1, Chapter 2 shows how the Atlantic Ocean can cause long-lasting changes in Australia rainfall through teleconnections. Chapter 3 reveals that Atlantic sea surface temperatures can influence north Pacific sea surface temperatures through remote teleconnections. Lastly, Chapter 4 explores trends in rainfall in the Pacific and the Indian Ocean from the 1980s to the 2010s. The results in this dissertation provide a path forward to increase the forecasting capability of the El Niño Southern Oscillation and help the climate science community enhance our understanding of climate variability.
ACKNOWLEDGMENTS

First, I would like to express my sincere gratitude to my advisor Dr. Yoshi Chikamoto. He was incredibly patient with me while I learned climate modeling and ocean dynamics coming from a traditional meteorology background. I will forever be thankful for him taking a chance on me. He spent an ungodly amount of time teaching, mentoring, and supporting me through my Ph.D. I will forever treasure the time I spent with him as a student.

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Zachary F. Johnson
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<td>AGCM</td>
<td>Atmospheric general circulation model</td>
</tr>
<tr>
<td>AOGCM</td>
<td>Atmosphere ocean general circulation model</td>
</tr>
<tr>
<td>AMIP</td>
<td>Atmospheric model intercomparison project</td>
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<td>AMO</td>
<td>Atlantic multidecadal oscillation</td>
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<td>Atlantic meridional overturning circulation</td>
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<td>Atlantic</td>
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<td>COBE</td>
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<td>CP-ENSO</td>
<td>Central Pacific El Niño southern oscillation</td>
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<td>GPCC</td>
<td>Global Precipitation Climatology Centre</td>
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<td>GPCP</td>
<td>Global precipitation climatology project</td>
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<tr>
<td>ECMWF</td>
<td>European Centre for Medium-Range Weather Forecasts</td>
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<td>Equatorial Pacific</td>
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<td>ERSST</td>
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ITCZ     Intertropical convergence zone
MIROC    Model for interdisciplinary research on climate
NCAR     National Center for Atmospheric Research
NCEP     National Centers for Environmental Prediction
NOAA     National Oceanic & Atmospheric Administration
NPGO     North Pacific gyre oscillation
OI       Optimum interpolation
OLR      Outgoing longwave radiation
PDO      Pacific decadal oscillation
PNA      Pacific north America
RCP      Representative concentration pathway
SAM      South Asian monsoon
SLP      Sea level pressure
SLPA     Sea level pressure anomaly
SOI      Southern oscillation index
SPCZ     South Pacific convergence zone
SSH      Sea surface height
SST      Sea surface temperature
SSTA     Sea surface temperature anomaly
SVD      Singular value decomposition
TBV      Trans-basin variability
WAF      Wave-activity flux
WP       Western Pacific
WWB      Westerly wind burst
CHAPTER 1
INTRODUCTION

1.1 Ocean-atmosphere interactions

1.1.1 Background

Examining ocean-atmosphere interactions on remote climate variability at interannual-to-decadal timescales and enhancing its predictability has broad socioeconomic implications on a wide range of sectors, such as water resources, agriculture, forestry, finance, energy, tourism, ecology, and environmental change. For instance, on interannual timescales, El Niño Southern Oscillation (ENSO) can cause remote impacts on drought and pluvial conditions in the United States, Australia, Asia, and South America (Barlow et al., 2001; Hu and Huang, 2009; McCabe et al., 2004; Taguchi et al., 2012; Tanimoto et al., 2003; Vance et al., 2015; Wang et al., 2014, 2009; Zanchettin et al., 2008). ENSO is split into three categories to describe sea surface temperature (SST) variability in the tropical Pacific: La Niña or cooling, neutral or lack of temperature anomalies, and El Niño or warming (Philander, 1983). ENSO’s remote influence on the climate system has intensified since 1980 due to an increase in greenhouse gases suggesting the potential for increased climate extremes in the coming century (Cai et al., 2019; Simon Wang et al., 2015). Similar to ENSO, the Pacific Decadal Oscillation (PDO), primarily forced by ENSO variability through ocean-atmosphere interactions, also has remote climate impacts and can cause changes in marine ecosystems and biogeochemical cycles over the North Pacific (Chavez et al., 2003; Francis and Hare, 1994; Mantua and Hare, 2002). From the perspective of these broad impacts, improving our understanding of Pacific interannual-to-decadal climate variability is essential not only for enhancing climate predictability but also its application to marine ecosystem predictions.
The scientific community accepts that ENSO can remotely impact the extratropics and other ocean basins; however, there is an ongoing debate regarding the Atlantic and Indian Ocean’s impact on the Pacific, highlighting the research area of inter-basin interactions (see review paper; Cai et al., 2019). By incorporating accurate predictions of inter-basin interactions, there may be significant improvements in climate predictability (Chikamoto et al., 2015, 2020; Izumo et al., 2010). However, many uncertainties exist in the extent of the Indian and Atlantic ocean’s impact on Pacific climate variability. In the Atlantic and Indian Ocean basins, there is a wide spectrum of climate variability both in time and spatial scale, so it is challenging to separate natural climate variability and anthropogenic forcing and whether they influence each other in the perspective of inter-basin interactions. Also, due to decadal-to-multidecadal variability in the Atlantic and Pacific basins, the short instrumental record limits research capability. To further complicate matters, there are errors and biases in simulating the Atlantic Ocean, causing uncertainty in the Atlantic’s influence on the Pacific (McGregor et al., 2018). The research presented in this dissertation explores some of the mechanisms of inter-basin interactions.

1.1.2 Bjerknes theory

We first review the climatological state of the tropical Pacific in both the atmosphere and the ocean. The tropical Pacific’s atmospheric climatological state consists of a zonal sea level pressure (SLP) gradient: high in the east and low in the west, which results in surface easterly winds in the central tropical Pacific. At the equator, these easterly winds cause westward flow at the ocean surface, allowing for ocean water to pile up in the western Pacific. However, a few degrees off the equator, these easterlies cause poleward oceanic Ekman transport due to the Coriolis force, allowing deep ocean cold water to upwell into the upper ocean layer at the equator. The easterly wind stress balances the zonal sea surface height (SSH) contrast, which consists of a deeper thermocline in the west but shallower in the east. The deeper thermocline in the western tropical Pacific leads to warmer SSTs and an unstable atmosphere resulting in deep convection. The shallower thermocline in the east leads to cooler SSTs and a stable atmosphere that inhibits convection. Bjerknes (1969)
described this tropical Pacific large-scale atmospheric circulation (i.e., surface easterlies, ascending motion in the west, upper-tropospheric westerlies, and descending motion in the east) as the Walker circulation.

*Bjerknes* (1969) found a positive atmosphere-ocean feedback to explain SST growth in the tropical Pacific through Sir Gilbert Walker’s Southern Oscillation (*Walker*, 1928; *Walker and Bliss*, 1932). The Southern Oscillation involves the seesawing of SLP in the tropical Pacific. It can be depicted as the Southern Oscillation Index (SOI), the difference in SLP between Darwin Australia and Tahiti, French Polynesia in the South Pacific. A positive SOI corresponds to a stronger zonal SLP gradient than normal, which leads to strengthened easterlies. These increased easterlies strengthen the zonal SSH gradient at the equator and cause a sharper tilt to the thermocline. The sharper tilt to the thermocline leads to a larger zonal SST gradient: colder than normal in the east and warmer than normal in the west. This SST gradient amplifies the initial SLP gradient in the tropical Pacific, increases the surface easterlies over the tropical Pacific, and leads to the ENSO cold phase. This positive feedback explains the developing stage of ENSO events.

### 1.1.3 ENSO theory

In order to end the positive feedback and allow the coupled ocean-atmosphere system to oscillate, a negative feedback occurs through ocean waves that can propagate thermocline depth anomalies across the equatorial Pacific (*Battisti*, 1988; *Godfrey*, 1975). Based on spatiotemporal ENSO scales, the two most relevant waves are Rossby waves and Kelvin waves (*Gill and Clarke*, 1974). Kelvin waves are eastward propagating waves trapped to the equator that take ~2.5 months to cross the equatorial Pacific, whereas Rossby waves are much slower westward propagating off-equatorial waves, which take ~8 months to cross the tropical Pacific. These waves can be triggered by westerly wind anomalies in the central tropical Pacific.

For example, assume climatological equatorial easterly winds, when westerly wind anomalies in the central tropical Pacific excite an eastward propagating Kelvin wave and a westward propagating Rossby wave. The Kelvin wave acts to deepen the thermocline in
the eastern Pacific, thereby favoring warm SSTs. Once the Kelvin wave reaches the eastern boundary, part of it reflects back as a Rossby wave, and the other part spreads poleward along the coasts of North and South America as a coastal Kelvin wave \( (Clarke, 1992) \). After the initial Kelvin wave, the deeper thermocline in the eastern Pacific leads to a weaker zonal SST gradient, which maintains westerly wind anomalies in the central tropical Pacific and increases the chance of triggering additional Kelvin waves. Hence, when westerly winds persist for many months, Kelvin waves can continue to warm the eastern tropical Pacific \( (Lengaigne et al., 2002, 2003) \). Meanwhile, the westerly wind anomaly excites a Rossby wave that acts to raise the thermocline in the western Pacific and eventually reflects on the western boundary as a Kelvin wave \( (Battisti, 1988; Boulanger et al., 2004; Kessler, 1991) \).

The reflected Kelvin wave on the western boundary helps raise the thermocline in the eastern Pacific about 6 months later after the initial onset of eastern Pacific warming. This process is termed the delayed oscillator and is responsible for the turnabout from warm to cold phases, or vice-versa \( (Battisti and Hirst, 1989; McCreary Jr, 1983; McCreary Jr and Anderson, 1984; Suarez and Schopf, 1988) \). The idea behind the delayed oscillator is that the initial warming of the tropical Pacific triggered by a Kelvin wave sows the seeds of its demise after the initial Rossby wave has traversed the Pacific and has come back again.

In addition to the delayed oscillator, other oscillator modes describe the ENSO phenomena, including the recharge oscillator \( (Jin, 1997) \), the western Pacific oscillator \( (Weisberg and Wang, 1997) \), and the advective-reflective oscillator \( (Picaut et al., 1997) \). The discharge oscillator, with a low frequency of \( \sim 3-5 \) years, considers the importance of heat content of the tropical Pacific to describe the behavior of ENSO. During an El Niño, poleward discharge of heat leads to a shallow thermocline anomaly in the eastern Pacific. The new thermocline depth facilitates colder SSTs in the eastern Pacific, which intensifies equatorial westward flow, and brings about a La Niña. During La Niña, equatorward recharge of heat content occurs, thereby causing a deeper thermocline anomaly in the eastern Pacific, eventually leading back to an El Niño. It is the recharge and discharge that allow the coupled ocean-atmosphere system to oscillate. The western Pacific oscillator emphasizes a negative
feedback without ocean wave boundary reflection, whereas the advective-reflective process can advect warm water eastward or westward (see review paper; Wang and Picaut, 2004).

1.1.4 Remote impact of ENSO

Tropical SST variability can affect convection activities, which result in remote atmospheric teleconnections (Klein et al., 1999). Convection triggered by an El Niño leads to upper tropospheric divergence over the equator (Gill, 1980). The anomalous poleward outflow causes a pair of off-equatorial upper tropospheric anticyclones in the form of atmospheric Rossby waves. Subsequent subtropical upper-tropospheric divergence triggers a Rossby wave train towards the extra-tropics impacting the midlatitude storm track through the Pacific North America (PNA) pattern and changes to the Aleutian low (Renwick and Wallace, 1996). Aleutian low variability triggers a North Pacific SST response through changes in surface wind speed. A similar mechanism can occur in the southern hemisphere, where ENSO causes changes in midlatitude storm tracks and extratropical SSTs (Newman et al., 2016). Differences in the pattern of SST anomalies and the location of maximum anomalies associated with ENSO, termed ”ENSO diversity,” will have an impact on the location of convection; hence, changes in the location of the extratropical response (see review paper; Capotondi et al., 2015). For instance, an eastern Pacific (EP) ENSO event typically triggers the PNA pattern while a central Pacific (CP) ENSO event causes changes in middle-to-upper tropospheric geopotential height in the Arctic and North Pacific, highlighting the Arctic Oscillation (Takahashi et al., 2011; Thompson and Wallace, 1998).

In addition to tropical-to-extratropical responses, ENSO can cause changes in the global Walker Circulation, thereby remotely impacting the tropical Indian and Atlantic oceans (Cai et al., 2019, and references therein). The tropical Pacific influences the Indian Ocean by the atmosphere and ocean via the Walker circulation and the Indonesian throughflow (ITF), respectively (Sprintall et al., 2014). ENSO-induced changes in the Walker circulation may lead to an SST dipole pattern in the Indian Ocean and basin-wide SST changes (Annamalai et al., 2003). In contrast to ITF transport, the tropical Pacific remote impacts on the Atlantic are primarily through the atmosphere in both tropical and extratropical
pathways. The tropical pathway involves changes in the global Walker Circulation, while the extratropical pathway occurs through the PNA pattern causing SLP changes over the western subtropical Atlantic (Wallace and Gutzler, 1981).

However, the Atlantic and Indian Oceans can feedback onto the Pacific from ENSO-induced SSTs highlighting inter-basin interactions. The Atlantic and Indian Oceans also have their own internal variability separate from ENSO forcing. This presents a challenge in separating internal and external forcing. An expanded review of ENSO remote forcing and inter-basin interactions is located in the introductions of Chapters 2 and 3. In order to examine these remote forcing mechanisms, appropriate climate modeling experiments are needed, given the two-way interaction of these ocean basins.

1.2 A brief history of climate model experiments

A climate model tries to represent processes that produce Earth’s climate through physical, chemical, and biological principles by using a series of equations calculated on supercomputers. However, in the late 19th century and early 20th century, the pioneers of climate modeling did not have the luxury of computers. An American meteorologist named Cleveland Abbe (1838-1916) recognized that "meteorology is essentially the application of hydrodynamics and thermodynamics of the atmosphere" and proposed a mathematical approach to weather prediction (Willis and Hooke, 2006). Moving one step further, Norwegian scientist Vilhelm Bjerknes (1862-1951) recognized that a set of nonlinear differential equations could be used to predict the weather (Nebeker, 1995). Lewis Fry Richardson (1881-1953) put Bjerknes’ method into practice and attempted to apply equations to forecast pressure change, but his result showed unrealistic pressure increases due to errors in his equations and initial conditions (Richardson, 2007). Electronic computers were developed during and after World War II, so by the 1960s, the Deutscher Wetterdienst (German Weather Service) produced an accurate simulation of the atmosphere in an operational setting. Phillips (1956) was the first to use a full set of "primitive equations" and ran the first long-ranged general circulation model to simulate the atmosphere. Following his work, in the 1960s, the Geophysical Fluid Dynamics Laboratory (GFDL) designed an experiment
that combines both oceanic and atmospheric processes.

To improve climate models, Manabe and Bryan (1969) recognized that the ocean and atmosphere needed to be coupled, leading to the first coupled ocean-atmosphere climate model (Bryan et al., 1975). In the 1980s and 1990s, further developments by incorporating other climate components like sea-ice, land surface, and atmospheric chemistry resulted in improvements in climate models (Washington et al., 1980). These vital components in the climate system established increased complexity requiring further computational resources. In a coupled climate model, different Earth system components like the ocean and atmosphere communicate and interact with each other producing feedbacks within the climate system. Each component runs separately as its own model, which requires a “coupler” to exchange information between the components. This exchange between models leads to a fully coupled atmospheric-ocean general circulation model (AOGCMs). In this dissertation, we examine inter-basin interactions, which requires a fully coupled AOGCM.

Several types of experiments were used to evaluate ocean-atmosphere interactions through coupled models. In the 1990s and 2000s, AGCM experiments prescribed observed SST (so-called AMIP-type experiment; Lau and Nath, 1994) to evaluate the direct atmospheric responses to the tropical Pacific SST forcing, but they do not capture the time evolution of the ocean response from the atmosphere due to a lack of an ocean dynamical model (McGregor et al., 2014). A role of thermodynamical air-sea interaction was further assessed with AMIP-type experiments coupled with mixed-layer ocean models (e.g., Alexander et al., 2002; Di Lorenzo et al., 2015; Lau and Nath, 1996). For example, Lau and Nath (2000) used an AGCM coupled with a simple 50-m mixed layer model to assess teleconnections from the tropical Pacific to the Indian Ocean. To assess the role of fully dynamical atmosphere-ocean interactions, Wang et al. (2012) proposed the partial decoupling experiment using the NCEP CFSv2 model, where SSTs are dampened toward climatological states in a targeted region, yet it was still challenging to isolate the remote ocean impacts from regional air-sea interaction.

To identify the climate variability originating from the tropical Pacific, a pacemaker
experiment was proposed, in which the observed SST variability in the tropical Pacific was prescribed in the ocean component of a climate model through heat flux forcing at the ocean surface (Kosaka and Xie, 2013). In these experiments, the ocean and atmosphere are able to evolve freely outside the targeted region but can also interact inside the targeted region. Using this pacemaker experiment, Zhang et al. (2018) demonstrated that the atmospheric bridge concept. In another pacemaker experiment, Ding et al. (2012) showed that the Atlantic Niño can impact the amplitude of ENSO by prescribing the observed SST field in the tropical Atlantic. The pacemaker experiment is one of the most advanced experiments because it allows one to investigate remote ocean-atmosphere response outside the area that is constrained.

1.3 Challenge simulating Atlantic variability

While tropical-to-extratropical processes in the Pacific have been well simulated by AMIP-type experiments, an experiment to capture the Atlantic impact on the Pacific is more challenging because subsurface temperature and salinity play a role in Atlantic Ocean dynamics, which may introduce systematic errors (Zhang and Zhao, 2015). Previous studies constrained the Atlantic Ocean using SST only, such as an idealized experiment by prescribing the AMO-induced SST pattern (Levine et al., 2018; Ruprich-Robert et al., 2017) or a pacemaker experiment targeted in the Atlantic (Ding et al., 2012). However, Boer et al. (2016) pointed out that pacemaker experiments forced by Atlantic SSTs may introduce energy and density imbalances due to a lack of salinity information, which causes artificial changes in atmosphere-ocean interaction and can alter the coupled model equilibrium. Therefore, a SST-only assimilation is not sufficient to constrain the ocean density structure of the Atlantic due to higher-frequency fluctuations. As a result, SST-only assimilation or pacemaker approaches may fail to properly simulate observed SST variability, thus impacting ocean-atmosphere interactions (Kajtar et al., 2018; Luo et al., 2018; McGregor et al., 2018; Sasaki et al., 2014).

To avoid errors in simulating ocean-atmosphere interactions in the Atlantic, this dissertation conducts experiments by assimilating temperature and salinity in the tropical Pacific,
Atlantic, and Indian Oceans in fully coupled climate models. Next, Chapter 2 investigates Australian interannual-to-decadal precipitation variability remotely forced by the tropical Pacific and Atlantic. Chapter 3 examines the remote response on PDO variability originating from the tropical Pacific and Atlantic Oceans. Chapter 4 explores precipitation trends in the Pacific and Indian Oceans from 1982-2014. Finally, Chapter 5 concludes this dissertation and give a perspective on future work. At the completion of this dissertation, the results of Chapter 2 and Chapter 3 were published in Johnson et al. (2018) and Johnson et al. (2020), respectively.
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CHAPTER 2
OCEAN IMPACTS ON AUSTRALIAN INTERANNUAL-TO-DECADAL PRECIPITATION VARIABILITY

2.1 Abstract

In Australia, successful seasonal predictions of wet and dry conditions are achieved by utilizing the remote impact of sea surface temperature (SST) variability in the tropical oceans, particularly the Pacific Ocean, on the seasonal timescale. Beyond seasonal timescales, however, it is still unclear which processes and oceans contribute to interannual-to-decadal wet/dry conditions in Australia. This research examines the interannual-to-decadal relationship between global SST anomalies (SSTAs) and Australian wet/dry variability by analyzing observational data and global climate model experiments conducted with the NCAR Community Earth System Model (CESM) and the Model for Interdisciplinary Research on Climate (MIROC). A 10-member ensemble simulation suite for 1960–2015 (CESM) and 1950–2010 (MIROC) is conducted by assimilating the observed 3-dimensional ocean temperature and salinity anomalies into fully coupled global climate models. In both observational analyses and ocean assimilation experiments, the most dominant annual mean precipitation variability shows a clear relation with SSTAs in the tropical Pacific and the Atlantic. Our partial ocean assimilation experiment, in which the ocean component of the CESM and MIROC are assimilated by the observed ocean temperature and salinity anomalies in the equatorial Pacific only, shows that the tropical Pacific SST variability is the main driver for the Australian precipitation variability at the interannual-to-decadal timescales. However, our additional partial ocean assimilation experiment, in which the climate models incorporate the observed anomalies solely in the Atlantic ocean, demonstrates that the Atlantic Ocean can also affect Australian precipitation variability at the interannual-to-decadal timescale through changes in tropical Pacific SSTAs and the mod-
ulation of the global Walker circulation. Our results suggest that about half of Australian interannual-to-decadal precipitation variability originates from the Atlantic Ocean.

### 2.2 Introduction

Australia experiences prolonged or intermittent severe droughts that devastate local economies, put significant strains on the agriculture industry, and contribute to wildfires (Dijk et al., 2013; Heberger, 2012). For example, the Millennium Drought (1997–2010) impacted the populated areas of Australia, including cities like Melbourne, Brisbane, and Sydney, causing severe water restrictions and a ban on outdoor water use (Kiem and Franks, 2004). In addition, a significant drop in the gross domestic product was observed and blamed on the drought in Australia (Horridge et al., 2005). Conversely, Australia is susceptible to flooding, which affects major metropolitan areas, such as the 1976 Brisbane and 2002 Townsville floods (Abrahams et al., 1976; Yeo, 2002). Due to Australia being surrounded by oceans on all sides, many studies have looked into the ocean’s impact on climate extremes in Australia, as introduced below.

The El Niño Southern Oscillation (ENSO), interactions between large-scale SST anomalies (SSTAs) and atmospheric circulation, has a significant impact on drought conditions around the world as well as Australia (Chiew et al., 1998). ENSO is split into three categories: La Niña or cooling of the equatorial Pacific, neutral or a lack of temperature anomalies in the equatorial Pacific, and El Niño or warming of the equatorial Pacific. Previous studies showed the linkage of El Niño to some of the major droughts in Australia (Allan, 1988; Dijk et al., 2013; McDonald et al., 2004; Nicholls, 1991; Philander, 1983). Through the tightly coupled atmosphere-ocean interaction, ENSO accompanies a zonal seesaw of atmospheric pressure anomalies in the Australian-Indonesian and eastern equatorial Pacific regions, which is known as the Walker Circulation (Bjerknes, 1969). As a result, many studies have shown ENSO to affect a significant portion of Australian precipitation variability. We note that some of the Australian precipitation events could be explained by the other drivers outside the tropical Pacific (Ummenhofer et al., 2009). Because of the predictability of ENSO (on the order of one or two years), climatologists and meteorologists
apply ENSO prediction to their seasonal climate forecasts (Luo et al., 2008; Meehl et al., 2014). Additionally, the Indian Ocean Dipole (IOD) can also affect precipitation variability over Australia and potentially have decadal predictability (Han et al., 2014a; Saji et al., 1999; Webster et al., 1999). It would be of great benefit for society if we could enhance the predictability of Australian precipitation anomalies and lead times beyond seasonal timescales.

While the Walker Circulation is mainly located within the Pacific region in the traditional sense of the term, similar atmospheric asymmetric zonal circulations are also observed in other ocean basins, such as the equatorial Indian Ocean and Atlantic region (Saji et al., 1999; Wang, 2002). It has been known that, on interannual timescales, the three tropical oceans display intimate interactions that affect the IOD, ENSO, and Atlantic Niño (Luo et al., 2010, 2017). On decadal timescales, recent studies have also suggested that the tropical Pacific interacts with the Indian and Atlantic Oceans through the modulation of global Walker circulation despite being separated by continents (Chikamoto et al., 2012; Kucharski et al., 2015; Li et al., 2015; Luo et al., 2012; McGregor et al., 2014; Rodríguez-Fonseca et al., 2009). Additionally, the Indian ocean can influence the Pacific Ocean through the atmospheric bridge (Izumo et al., 2010; Luo et al., 2012). This leads to the inter-basin scale decadal climate variability. Some state-of-the-art climate prediction systems demonstrate that the tropical inter-basin interaction between the Pacific, Atlantic, and Indian Oceans shows multi-year predictive skills (Chikamoto et al., 2015a; Choudhury et al., 2017; Luo et al., 2017). Moreover, there is evidence that oceans can provide decadal predictability in Australia, given the ocean’s large effect on the continent (Power et al., 1999). Nevertheless, it is still unclear whether and how the tropical oceans affect the precipitation variability at interannual-to-decadal timescales in Australia.

In this study, we examine Australian interannual-to-decadal precipitation variability originating from the ocean using two global climate models: the Community Earth System Model (CESM) and the Model for Interdisciplinary Research on Climate (MIROC). In these models, observed 3-dimensional ocean temperature and salinity anomalies are assimilated,
and external radiative forcings are prescribed. By identifying the consistent variability between observation and model simulation, we will discuss how the tropical Pacific and Atlantic Oceans contribute to Australian precipitation variability. We note that possible contributions of the Indian Ocean are not examined in the work considering the relatively small impact of the Indian Ocean on Pacific decadal change in recent decades (Chikamoto et al., 2015a); however, the potential role of the Indian Ocean is discussed in Section 2.4.

2.3 Data and model experiments

2.3.1 Observational datasets

Our model experiments are validated by observational datasets including precipitation from the Global Precipitation Climatology Centre (GPCC), monthly mean sea level pressure (SLP) derived from NCEP reanalysis, and sea surface temperature (SST) from the Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4). Precipitation from the GPCC is based on over 67,000 stations worldwide and has a grid spacing of 0.5° ranging from 1901–2016 (Becker et al., 2011). SLP extends from 1948–2015 with a 2.5° resolution (Kalnay et al., 1996). SST has a 2° resolution extending from 1854–2016 (Huang et al., 2015). These reconstructed datasets use statistical analyses for spatial completeness, especially since observations are more sparse in earlier years.

2.3.2 Models

This study uses two fully coupled low-resolution models: the CESM 1.0.3 (Shields et al., 2012) and the MIROC 3.2m (Nozawa et al., 2007). The CESM relies on the physics from the Community Climate System Model (Gent et al., 2011) version 4, which includes atmospheric dynamics from the Community Atmosphere Model (Neale et al., 2013) version 4. Resolution for atmospheric and land are T31 spectral (approximately 3.75°), including hybrid sigma pressure coordinates of 26 atmospheric levels and 15 soil levels (surface–35 m). The ocean model includes 60 vertical levels and is coupled with the sea-ice model, which has a resolution of approximately 3°. Within the CESM, the land, atmosphere, and
sea-ice communicate every 30 min, while the ocean couples with the atmosphere once a day. The land portion is from the Community Land Model version 4 and has a carbon-nitrogen biogeochemical cycle, a groundwater model, and a wildfire scheme (Lawrence et al., 2012). Details of the model’s basic performance in the configuration can be found in results of the CESM described in previous studies (Chikamoto et al., 2015b; Shields et al., 2012).

The MIROC uses a T42 spectral horizontal grid including 20 vertical levels of sigma pressure coordinates and a standard physics package cooperatively developed at the Center for Climate System Research (CCSR), the University of Tokyo’s Atmosphere and Ocean Research Institute, and the Frontier Research Center for Global Change (K-1 Model Developers, 2004). The ocean component has a higher resolution (1.4°-latitude and 0.56-1.4°-longitude) compared to the CESM with 44 levels. All components of atmosphere, ocean, land, and sea-ice modules are coupled without any flux corrections for exchanging heat, water, and momentum fluxes between the atmosphere and the ocean (Komuro et al., 2012). Details of the performance and settings of the MIROC are described in a previous study (Nozawa et al., 2007).

2.3.3 Partial assimilation experiments

In this study, we analyze ocean data assimilation experiments as described in the previous studies (Chikamoto et al., 2015b, 2016; Ham et al., 2017; Purich et al., 2016). Both the CESM and MIROC runs consist of 10 ensemble members with time-varying observed external forcings (solar, aerosols, land-use change, and greenhouse gases) prior to 2005. After 2005, we prescribe the RCP4.5 emission scenario for the CESM but the A1B-type scenario for the MIROC. Using these configurations for the external forcings, the observed global 3-dimensional ocean temperature and salinity anomalies (CESM: 0–3000 m; MIROC: 0–700 m) are assimilated into the ocean components of global climate models (the global ocean assimilation runs). By assimilating the surface and sub-surface ocean field, the models can simulate ocean variability in the mixed layer depth, thermocline, and thermodynamics more appropriately compared to a SST-only assimilation. As a result, our assimilation run would be suitable to investigate the interannual-to-decadal climate variability. These observations
are derived from the ECMWF ocean reanalysis product version 4 for 1958–2014 in the CESM but the objective analysis termed ProjD for 1945–2010 in the MIROC (Balmaseda et al., 2013; Ishii and Kimoto, 2009). The monthly observations of these datasets are linearly interpolated into daily values and then assimilated into the ocean model component in each climate model using an Incremental Analysis Update scheme (Bloom et al., 1996; Huang et al., 2002). These ocean assimilation systems are developed individually in each model, but they are applied by a similar method. More detailed methodology and its application for decadal climate prediction are described by the previous studies for the MIROC (Mochizuki et al., 2010; Tatebe et al., 2012) and the CESM (Chikamoto et al., 2017).

To evaluate the ocean contribution to the Australian precipitation, we partially assimilated the observed 3-dimensional ocean temperature and salinity anomalies in the equatorial Pacific (CESM and MIROC: 10°S–10°N) and the entire Atlantic (CESM: 30°S–70°N; MIROC: 50°S–60°N), respectively. Each partial assimilation and global runs include 10 ensemble members with different time ranges for the CESM (1960–2014) and MIROC (1950–2010) after the model spin-up (2 years for the CESM and 5 years for the MIROC). The ensemble mean was used to analyze the data. Initial conditions for those assimilation runs are obtained from 10-member initial conditions within the twentieth century historical simulation as designed by the Coupled Model Intercomparison Project-5 (i.e., prescribing the observed external forcings without ocean data assimilation). Our assimilation procedure takes into account the model climatological biases and has no significant model drift in the global and partial ocean data assimilation runs. These assimilation systems have been developed in order to conduct the decadal climate predictions in the previous studies (Chikamoto et al., 2017; Mochizuki et al., 2010). The partial assimilation experiments capture both the low-frequency atmospheric response to SST variability in the partially assimilated area (i.e., Atlantic or equatorial Pacific) and air-sea interaction responses that are remotely forced by the partially assimilated ocean. Partial assimilation and similar experiments were conducted by previous studies (Chikamoto et al., 2015a, 2016; Kosaka and Xie, 2013; Kucharski et al., 2015; Li et al., 2015; McGregor et al., 2014; Zhang and
Delworth, 2007) and using two separate general circulation models allows us to evaluate robust features as well as model uncertainty.

2.4 Results

2.4.1 Model validation

To capture the most correlated variability of Australian precipitation between observation and model simulation, we conducted the singular value decomposition (SVD) analysis (Wallace et al., 1992) (similar to the maximum covariance analysis but using correlation matrix) from the annual mean precipitation anomalies over Australia (45°S–10°S, 105°E–160°E) between GPCC and the ensemble mean of the model global runs (top panels in Figs. 2.1 and 2.2). Because of the different lengths of model simulations, the SVD analysis is applied from 1960–2014 in the CESM and 1950–2010 in the MIROC. Nevertheless, the first SVD mode in both the CESM and MIROC against observations explains most of the variance of the Australian precipitation variability through the square covariance fraction (SCF), the percentage of square covariance between the two fields explained in the leading SVD mode (SCF=85.5% and 88.6%, respectively). The model’s simulated temporal variations in the first SVD mode are highly correlated with the observed variations (correlation coefficient $R = 0.66$ in the CESM and 0.65 in the MIROC; Fig. 2.1c and 2.2c). The observed precipitation pattern associated with the first SVD mode shows a monopole structure in the entirety of Australia with a local maximum on the eastern side (Fig. 2.1a and 2.2a). This monopole pattern is also captured by the first SVD mode of global runs in both the CESM and MIROC very well (Figs. 2.1b and 2.2b), although both models overestimate the precipitation anomalies over western Australia. In particular, the first SVD mode captures the multi-year wet periods around the mid-1970s and late-1990s and additionally, the dry periods around the early-1990s and early-2000s (Figs. 2.1c and 2.2c). During these periods, Australia experienced severe flooding events at Brisbane (Abrahams et al., 1976) in 1974, Townsville (Yeo, 2002) in 1998, and in Queensland (van den Honert and McAneney, 2011) in 2010–2011, and also, drought in Queensland (Fensham, 1998) in the early-1990s and the
millennium drought (Dijk et al., 2013) in the early-2000s. In fact, the observed and model simulated principal components of these first SVD modes (black and red lines in 2.1c and 2.2c) exhibit a larger power in the lower frequency components with a spectral peak at 12.5 years (a frequency of 0.08 cycle per year) in both the CESM and MIROC (black lines in Figure 2.3). Our SVD analysis suggests that these wet and dry conditions are mainly attributed to the SSTAs on Australian precipitation variability.

First SVD Mode (1960-2014)

Fig. 2.1: The first SVD modes of annual mean precipitation anomalies in Australia between the observation (GPCC) and the CESM simulations of (a-c) the global assimilation run, and partial assimilation runs in (d-f) the equatorial Pacific and (g-i) the Atlantic during 1960–2014. Left and middle columns are homogeneous correlation maps in observation and model simulations, respectively. The squared covariance fraction (SCF) explained by the first SVD mode is indicated in the upper-right corner of these panels. The right column shows the principal components of the first SVD modes between observation (black) and model simulations (red). Correlation coefficients of these principal components in the first SVD mode are noted on the top right of each time series plot. A correlation coefficient of 0.35 is statistically significant at 99% level with 50 degrees of freedom. The SVD analysis is conducted based on the correlation matrix.
Fig. 2.2: Same as Figure 2.1 but for MIROC during 1950–2010.
Fig. 2.3: Power spectrum for the principal components of the first SVD mode in observations (broken lines) and model simulations (thick solid lines) in the global (black), equatorial Pacific (blue), and Atlantic (red) assimilation runs in (a) CESM and (b) MIROC. The thin solid line and thin dotted line correspond to the power spectrum of a fitted 1st order Markov process and its 95 and 5% confidence limits for model simulations in the Atlantic (red) and global (black) runs, respectively.

Figures 2.4a and 2.5a show the observed patterns of SST and SLP anomalies associated with the first SVD mode between the GPCC observation and the global assimilation runs. We found La Niña-like SST and SLP anomaly patterns that are characterized by the SST cooling in the central tropical Pacific, its surrounding warming in the western Pacific, and a zonal pressure gradient with positive SLP anomalies in the eastern tropical Pacific and negative ones in the Indo-western Pacific regions. This zonal SLP gradient associated with the first SVD mode corresponds to the strengthened (weakened) Walker circulation in the Indo-Pacific sector during enhanced (suppressed) Australian precipitation years. The La Niña-like patterns accompany the warm SSTAs around the Australian coastline and the lower-than-normal SLP covering its entire continent, which is consistent with the increase in Australian annual precipitation. These features are captured very well in the global runs of both the CESM and MIROC (Figs. 2.4b and 2.5b), indicating that the SSTAs play an important role for the Australian interannual-to-decadal precipitation variability. These
results support that our model experiments are reasonable to evaluate the ocean impacts on the Australian precipitation variability.

Fig. 2.4: Correlation maps of SST (shade) and SLP anomalies (contour: solid and broken lines are positive and negative correlations, respectively) associated with the first SVD mode between observation and the CESM simulation in the global (top), the equatorial Pacific (middle), and Atlantic assimilation runs (bottom) during 1960-2014. Contour interval is 0.2 and zero contour is omitted. Observed SST and SLP datasets are obtained from ERSST and NCEP, respectively.
Fig. 2.5: Same as Figure 2.4 but for the MIROC during 1950-2010.

### 2.4.2 Ocean impacts on Australian precipitation variability

As described in the previous subsection, the first SVD modes of Australian precipitation between observation and the global runs of both the CESM and MIROC accompany significant SST correlations in the equatorial Pacific with a wide swath exceeding $-0.6$ (Figs. 2.4b and 2.5b), suggesting that the equatorial Pacific SST is the primary driver for Australian precipitation variability. To further explore this, we applied the SVD analysis for the Australian precipitation variability between observation and the ensemble means of the equatorial Pacific partial assimilation runs (Figs. 2.1d-f and 2.2d-f). Similar to the results in the global runs, the first SVD mode between observation and the equatorial Pacific runs explains most of the variance in both the CESM and MIROC (86.6% and 89.1%). These first SVD modes show significant correlation coefficients of principal components between the observation and the model simulations in both the CESM and MIROC ($R = 0.63$ and 0.67 in Figs. 2.1f and 2.2f) and exhibit monopole precipitation patterns over Australia (Figs.
2.1c-d and 2.2c-d) that are almost identical with the results of the global runs (Figs. 2.1a-b and 2.2a-b). These first SVD modes of Australian precipitation variability in the equatorial Pacific partial assimilation runs demonstrate the La Niña-like pattern, including the central tropical Pacific SST cooling and the positive zonal gradient of SLP anomalies between the eastern tropical Pacific and the Indo-western Pacific region (Figs. 2.4d and 2.5d). This result strengthens the suggestion that the La Niña-like SST forcing in the equatorial Pacific is the main driver of the enhanced annual precipitation in Australia and vice versa. We can see the lower-than-normal SLP anomalies over the entire Australian continent associated with the weakened Pacific Walker circulation (Figs. 2.4d and 2.5d), which is consistent with the enhanced precipitation over Australia (Figs. 2.1e-f, 2.2e-f). There are also significant correlations in the model simulated principal components of the first SVD modes between the global runs and the equatorial Pacific partial assimilation runs for both the CESM (red lines in Figs. 2.1c and 2.1f; $R = 0.71$) and the MIROC (red lines in Figs. 2.2c and 2.2f; $R = 0.79$), further strengthening this result. Although ENSO is a major driver for Australian precipitation variability on seasonal timescales (Chiew et al., 1998), a similar mechanism showing the equatorial Pacific Ocean impacts on the Australian precipitation is found even on the decadal timescale (Figure 2.3).

There is evidence that the Atlantic ocean may affect the tropical Pacific climate variability (Chikamoto et al., 2015a; Kucharski et al., 2015; Li et al., 2015; McGregor et al., 2014), and in turn, potentially affecting Australian precipitation (Choudhury et al., 2017). To identify the Atlantic origin for Australian precipitation variability, we applied the SVD analysis to the Australian precipitation variability between observation and the ensemble means of the Atlantic partial assimilation runs (Figs. 2.1g–i and 2.2g–i). The first SVD modes in the Atlantic partial assimilation runs explain 43.5% in the CESM and 65.1% in the MIROC for total covariance, which are separated well from higher modes (the SCFs of second SVD modes in the CESM and MIROC are 17% and 11.1%, respectively). Correlation coefficients of the principal components in the leading SVD modes are statistically significant ($R = 0.37$ in the CESM and 0.46 in the MIROC). The Atlantic partial assimi-
lated experiment also captures a very similar positive monopole-like pattern for observed precipitation variability comparable to the global and equatorial Pacific runs (Figs. 2.1 and 2.2). The observed principal components of the first SVD modes in the global, equatorial Pacific, and Atlantic runs are almost identical with each other (black lines in Figs. 2.1c,f,i and 2.2c,f,i; $R > 0.95$). In addition to the Australian precipitation variability, the observed SST and SLP anomalies correlated with these first SVD modes and also show indistinguishable patterns from each other (left panels in Figs. 2.4 and 2.5). Even though the model simulated precipitation variability is driven by the different ocean basins, the SVD analysis captures the same observed precipitation variability in Australia.

In the Atlantic partial assimilation runs, the first SVD mode accompanies the equatorial Pacific SST cooling (Figs. 2.4f and 2.5f) and the enhanced precipitation over Australia (Figs. 2.1h and 2.2h) similar to the results of the global and equatorial Pacific runs. Previous studies indicate that the Atlantic Ocean variability can drive the tropical Pacific climate variability through the modulation of the global Walker circulation (Chikamoto et al., 2015a; Kucharski et al., 2015; Li et al., 2015; McGregor et al., 2014). Consistent with these results, our Atlantic partial assimilation run demonstrates that the SST cooling in the equatorial Pacific partially originates from the Atlantic Ocean variability, which further affects enhanced precipitation in Australia. This negative correlation between the eastern equatorial Pacific SST and Australian precipitation in the Atlantic partial assimilation run is stronger in the MIROC than the CESM (Figs. 2.4f and 2.5f). In fact, the first SVD mode in the MIROC Atlantic partial assimilation runs clearly shows equatorial Pacific SST cooling and lower-than-normal SLP over the Australian continent (Fig. 2.5f) though those responses are much weaker in the CESM Atlantic partial assimilation run (Fig. 2.4f). In any case, our results suggest that the Atlantic is able to trigger the central Pacific SST variability and the subsequent changes in Australian annual precipitation.

2.4.3 Atlantic origin in Australian precipitation

To identify the Atlantic origin for the central tropical Pacific SST variability, we produced a scatter plot of the annual mean SSTAs in the Niño 4 region (5°S–5°N, 160°E–
150°W) between the observation and the Atlantic partial assimilation runs of the CESM and MIROC (Figs. 2.6 and 2.7). Similar results are obtained when we use the global assimilation runs instead of the observations. The Niño 4 region was chosen over the Niño 3.4 region (5°S–5°N, 170°E–120°W) because it has a lower-frequency (Trenberth, 1997; Trenberth and Stepaniak, 2001). The points at the upper-right and lower-left quadrants imply that the SSTAs in the central tropical Pacific originate from the Atlantic Ocean variability, whereas those at the upper-left and lower-right quadrants indicate the lesser role of the Atlantic Ocean in causing the tropical Pacific SST variability. By choosing the consistent years in the central tropical Pacific SSTAs between the observation and the Atlantic partial assimilation runs, we evaluate the Atlantic impacts on the Australian precipitation variability via changes in the tropical Pacific climate. For the CESM, we can find 8 cold years of the central tropical Pacific SSTAs associated with the Atlantic forcing (below −0.5°C; 1964, 1984, 1988, 1989, 1998, 1999, 2010 and 2011) but only 3 warm years (above 0.5°C; 1977, 2004 and 2009). A higher number of years is extracted in the MIROC Atlantic partial assimilation run: 10 cold years (1964, 1971, 1973, 1974, 1984, 1985, 1988, 1989, 1999 and 2000) and 9 warm years (1957, 1958, 1977, 1987, 1993, 2002, 2003, 2004 and 2005). Clearly, the MIROC (Fig. 2.6) was able to produce more warm and cold years compared to the CESM (Fig. 2.7), which enhanced the correlation for the MIROC ($R = 0.29$) compared to the CESM ($R = 0.00$). This higher correlation in the MIROC compared to the CESM is consistent with the SVD analysis in the Atlantic partial assimilation run, in which the central tropical Pacific SST cooling is much larger in the MIROC than the CESM (Figs. 2.5f and 2.4f).
Fig. 2.6: Scatterplot of SSTAs averaged over the Niño 4 region in observation and the CESM Atlantic partial assimilation run from 1960–2014. Niño 4 values were standardized by the standard deviation of the dataset. Composite years were chosen for those exceeding or being less than 0.5 and −0.5 standard deviations for both the observation and the Atlantic partial assimilation run (black box). Dot size and color indicate the observed principal component in the SVD analysis between observation and the Atlantic partial assimilation run (black line in Fig. 2.1i). $R = 0.00$. 
Using the cold and warm years extracted from the scatter plots of the central tropical Pacific SSTAs (Figs. 2.6 and 2.7), we conducted the composite analysis of annual precipitation anomalies in Australia (Figs. 2.8 and 2.9). Consistent with the SVD analysis, the observed annual precipitation anomalies tend to show wet conditions in the entirety of Australia during cold years but dry during warm years (left panels in Figs. 2.8 and 2.9). During cold years, in particular, the enhanced precipitation anomalies in eastern Australia is well simulated in the equatorial Pacific and the Atlantic partial assimilation runs of both the CESM and MIROC (Fig. 2.8). We can also confirm the consistency with the SVD analysis, in which those cold years show the larger observational principal component in the first SVD mode (blue dots in Figs. 2.6 and 2.7), except for 1964, 1985, and 1988. Similar features but in an opposite phase are also obtained in the warm year composite in the MIROC (bottom panels in Figs. 2.9), though the warm year composite in the CESM shows a noisier pattern because of the smaller sampling number of warm years (only three years; upper panels in Figs. 2.9). Additionally, in the warm composite years, precipitation anomalies in
the CESM Atlantic run show almost no statistical significance over Australia (black dots in Figure 2.9c). Because of this lesser Atlantic impact on the warmer SSTAs in the central tropical Pacific, the Atlantic partial assimilation run in the CESM, compared to the MIROC, shows the weaker impacts on Australian precipitation variability.

Fig. 2.8: Annual precipitation anomalies (mm/day) in (a,d) observation, (b,e) the equatorial Pacific, and (c,f) the Atlantic partial assimilation runs during the cold year composites in the CESM (top) and the MIROC (bottom). The dotted region is above the 90% statistical significance level using Student’s t-test for precipitation anomalies. The cold years are obtained from scatter plots in Fig. 2.6 and 2.7, in which the annual SSTAs in the Niño 4 region are below −0.5 standard deviations in both observation and the Atlantic partial assimilation runs of the CESM (1964, 1984, 1988, 1989, 1998, 1999, 2010 and 2011) and the MIROC (1964, 1971,1973, 1974, 1984, 1985, 1988, 1989, 1999 and 2000).
Fig. 2.9: Same as Figure 2.8 but for warm year composites. Warm years correspond to 1977, 2004 and 2009 in the CESM and 1957, 1958, 1977, 1987, 1993, 2002, 2003, 2004 and 2005 in the MIROC.

Figures 2.10 and 2.11 show the SST and SLP anomaly pattern in the cold and warm year composites. For the cold composite years in both the CESM and MIROC, the observation shows the equatorial Pacific SST cooling and the zonal SLP anomaly gradient in the tropical Pacific with the dominant negative SLP anomalies over Australia and the eastern Indian Ocean (left panels in Fig. 2.11). These SST and SLP anomaly patterns are captured well in the equatorial Pacific partial assimilation runs in both the CESM and MIROC models (middle panels in Fig. 2.11), indicating that the equatorial Pacific SST cooling and subsequent Australian decreased SLP anomalies are the main factors for the enhanced precipitation in Australia. The Atlantic partial assimilation runs in both the CESM and MIROC also capture these features, albeit with almost half of the amplitudes of the equatorial Pacific SST cooling compared to the observation and the equatorial Pacific partial assimilation runs (right panels in Fig. 2.11). Similar features but opposite phases are obtained for the warm year composite in the MIROC (bottom panels in Fig. 2.10). These results strongly support our findings that the Atlantic Ocean variability contributes to the
Australian precipitation variability through changes in SLP anomalies over the Australian continent via the equatorial Pacific SST forcing.

On the other hand, the warm year composite in the CESM may have a different process regarding the Atlantic impacts on Australian precipitation variability. The CESM warm year composite shows the northwest-southeast SLP anomaly gradient over the Australian continent in all of the observation, the equatorial Pacific, and the Atlantic partial assimilation runs even though there are warmer SSTAs in the equatorial Pacific (upper panels in Fig. 2.10). Because of the geostrophic balance, this SLP gradient within the Australian continent accompanies the southerly wind anomalies, which then may cause suppressed precipitation in southern Australia through the meridional advection of dry air from higher latitudes. However, this process may depend on the local wind anomalies in Australia, and the CESM may hardly resolve this local process because of its lower horizontal resolution. As a result, the Australian precipitation anomalies in the CESM warm year composite exhibit diverse patterns in observation, the equatorial Pacific, and the Atlantic partial assimilation runs (upper panels in 2.9).

**Cold Composite of SLP and SST Anomalies**

Fig. 2.10: Same as Fig. 2.8 but for SST (color interval is 0.1°C) and SLP anomalies (contour interval is 0.2 hPa). The dotted region is above the 90% significance level using Student’s t-test for SLP anomalies.
2.5 Discussion

Consistent with previous studies (Chikamoto et al., 2012, 2015a, 2017; Kucharski et al., 2011, 2015; Li et al., 2015; McGregor et al., 2014), our Atlantic partial assimilation runs identifies the Atlantic’s important impacts on tropical Pacific climate variability through the tropical inter-basin gradients of SST and SLP anomalies between the Atlantic and the Pacific. In fact, our composite analyses in both observations and the Atlantic partial assimilation runs show the tropical inter-basin contrast of SSTAs between the two oceans: the equatorial Pacific SST warming (cooling) is associated with the SST cooling (warming) in the southern tropical Atlantic Ocean during the warm (cold) year composites in both the MIROC and CESM (left and right panels in Figs. 2.10, and 2.11). During the cold year composites, particularly, we can find the warm SST and negative SLP anomalies in the southern tropical Atlantic in observations and the Atlantic partial assimilation runs (left and right panels in Fig. 2.11), which consists of the tropical inter-basin SST and SLP gradients. Notably, the equatorial Pacific partial assimilation runs show unclear responses in the tropical Atlantic SSTAs (Figs. 2.4d, 2.5d, 2.10b, 2.10e, 2.11b and 2.11e), supporting our hypothesis about the Atlantic origin for the tropical Pacific climate variability. Our finding is also consistent with previous studies, in which the southern tropical Atlantic SST warming affects the inter-basin gradient of SLP anomalies, which triggers the central Pacific SST cooling through the modulation of global Walker circulation (Chikamoto et al.,
Previous studies referred to the inter-basin SST and SLP gradients as the trans-basin variability (TBV) (Chikamoto et al., 2015a; McGregor et al., 2014). Based on the perspective of TBV, our study implies the following process about the Atlantic impacts on the Australian precipitation variability. The SSTAs in the tropical Atlantic cause atmospheric vertical motions and its compensating vertical motions in the central tropical Pacific, which results in changes in the Pacific Walker circulation. Once the Pacific Walker circulation changes due to Atlantic forcings, it drives atmosphere-ocean interaction in the tropical Pacific through the local Bjerknes feedback. Due to the fully coupled climate models, we can identify the La Niña-like atmosphere-ocean response to the Atlantic forcing. However, it is still unclear what role the 3-dimensional ocean field has on the low-frequency ocean variability. Because Australia is located in the western part of the Pacific Walker circulation, the vertical motion associated with the changes in the western part of the Pacific Walker circulation directly affects Australian precipitation variability. As a result, the warmer (colder) SSTAs in the tropical Atlantic cause SST cooling (warming) in the central Pacific, which in turn induces the upward (downward) motions and enhanced (suppressed) precipitation over Australia through the strengthened (weakened) Pacific Walker circulation. It is still unclear from this research which mechanisms regulate the warming or cooling of the Atlantic.

In addition to the Pacific and Atlantic basins, the Indian Ocean may also have a role in Australian precipitation variability. SST correlations associated with the first SVD modes show a dipole-like pattern in the Indian Ocean: negative correlations in the west and positive in the east except for the CESM Atlantic run (Figs. 2.4a-e and 2.5a-f). These patterns resemble the IOD (Saji et al., 1999; Webster et al., 1999) and Ningaloo Niño (Feng et al., 2013), which is known to affect precipitation over Australia (Tozuka et al., 2014; Ummenhofer et al., 2009). In addition to these impacts on Australian precipitation variability, the Indian Ocean could influence the Pacific Ocean through an atmospheric bridge on interannual-to-decadal timescales (Han et al., 2014b; Izumo et al., 2010; Luo et al., 2012; Mochizuki et al., 2016a,b). Further analysis regarding the inter-basin interaction
among three oceans would be beneficial to understand and improve Australian precipitation predictability on interannual-to-decadal timescales.

From this research, there are implications for increased predictability of Australian wet/dry conditions. Whereas the current seasonal prediction of Australian precipitation relies mainly on the ENSO predictive skills, multi-year predictability for wet/dry conditions in Australia may be achieved by utilizing the TBV predictability (Lee et al., 2010). Generally speaking, climate variability with larger spatial scales has longer predictability (Baldwin et al., 2003). Because the inter-basin interaction such as the TBV show larger spatial scales than ENSO by including the Atlantic Ocean, the Atlantic impacts on the Australian precipitation variability may have longer timescales compared to ENSO. In fact, some state-of-the-art climate prediction systems show multi-year predictive skills of TBV, which reflects much longer predictability than ENSO (Chikamoto et al., 2015a; Choudhury et al., 2017). Moreover, the Atlantic partial assimilation runs show decadal variability with peaks and ridges lasting several years, particularly after the mid-1980s (Figs. 2.1i and 2.2i), and additionally, there is a decadal peak of spectral power for Australian precipitation variability, particularly in the MIROC (Figure 2.3). The longer predictability of Australian precipitation could translate into a better assessment of climate risks for agriculture, water resources, and fire probability (Chikamoto et al., 2015b, 2017).

2.6 Conclusions

This research evaluated the potential interannual-to-decadal predictability of Australian precipitation using fully coupled global climate models: the MIROC and the CESM. By assimilating the observed ocean temperature and salinity in the global, equatorial Pacific, and Atlantic Oceans with 10-member ensemble suites, we evaluate ocean impacts on Australian precipitation variability. Past researchers mainly focus on the Indian and Pacific basins for precipitation anomalies in Australia, particularly on seasonal-to-interannual timescales (Ashok et al., 2003; Dai and Wigley, 2000; England et al., 2006; Meehl, 1987; Ropelewski and Halpert, 1987; Ummenhofer et al., 2011). This research, however, took into consideration the Atlantic basin and inter-basin interaction for Australian precipitation
variability on interannual-to-decadal timescales. Our study also discusses the importance of inter-basin interaction for climate predictability as described in the previous studies (Chikamoto et al., 2012, 2015a; Kucharski et al., 2011, 2015; Li et al., 2015; McGregor et al., 2014; Okumura et al., 2009; Timmermann et al., 1998).

The results of this study show that Australia precipitation variability is mostly attributed to global ocean variability. In particular, the equatorial Pacific SST variability is the main driver for the observed precipitation variability in Australia. Interestingly, however, we also identified the Atlantic Ocean impacts on Australian precipitation variability by affecting changes in the tropical Pacific climate through the TBV mechanism. Our SVD analysis in the Atlantic partial assimilation runs suggests that predictability of Australian decadal precipitation anomalies would be enhanced by utilizing the multi-year predictive skill of TBV (Chikamoto et al., 2015a). Combining our result with the seasonal forecast of ENSO, Australian climatologists may provide more accurate forecasts of Australian wet/dry conditions on seasonal to decadal timescales. Future work should focus on additional experiments using different models and examining the TBV role in past and future climate. There is also a need to find which mechanisms affect the warming and cooling of the Atlantic Ocean because Atlantic changes are important for the Australian decadal precipitation variability through changes in the tropical Pacific climate.
REFERENCES


CHAPTER 3
PACIFIC DEcadAL OSCillATION REMOTELY FORCED BY THE EQUATORIAL PACIFIC AND ATLANTIC

3.1 Abstract

The Pacific Decadal Oscillation (PDO), the leading mode of Pacific decadal sea surface temperature variability, arises from two main mechanisms: regional air-sea interaction within the North Pacific and remote forcing outside the North Pacific, such as the tropical Pacific and the Atlantic. Because of the combination of these two mechanisms, a question remains as to how much PDO variability originates from these basins. To better understand PDO variability, the equatorial Pacific and the Atlantic impacts on the PDO are examined using 3-dimensional partial ocean data assimilation experiments conducted with two global climate models: the CESM1.0 and MIROC3.2m. In these partial assimilation experiments, the climate models are constrained by observed temperature and salinity anomalies solely in the Atlantic basin and solely in the equatorial Pacific basin but are allowed to evolve freely in other regions. These experiments demonstrate that, in addition to the tropical Pacific’s role in driving PDO variability, the Atlantic can also affect PDO variability by modulating the tropical Pacific climate through two proposed processes. One is the equatorial pathway, in which tropical Atlantic sea surface temperature (SST) variability causes an El Niño-like SST response in the equatorial Pacific through the reorganization of the global Walker circulation. The other is the north tropical pathway, where low-frequency SST variability associated with the Atlantic Multidecadal Oscillation induces a Matsuno-Gill type atmospheric response in the tropical Atlantic-Pacific sectors north of the equator. These results provide a quantitative assessment suggesting that 12–29% of PDO variance originates from the Atlantic Ocean and 40–44% from the tropical Pacific, while the remaining 27–48% of variance still require other processes such as regional ocean-atmosphere interactions in the
North Pacific and possibly some influence from the Indian Ocean.

3.2 Introduction

The Pacific Decadal Oscillation (PDO) is the dominant low frequency climate variability in the North Pacific and is defined as the leading mode of North Pacific detrended sea surface temperature anomalies (SSTAs) (20°–70°N) calculated by a statistical decomposition called empirical orthogonal functions (EOFs) (Mantua et al., 1997; Minobe, 1997; Zhang and Levitus, 1997). The positive (negative) phase of the PDO is characterized by cool (warm) sea surface temperatures (SSTs) in the Kuroshio-Oyashio Extension (KOE) region and its surrounding warm (cool) SSTs along the west coast of North America (Mantua and Hare, 2002; Newman et al., 2016; Qiu, 2002). Associated with these physical ocean changes, the coherent decadal variability is also observed in marine ecosystems and biogeochemical cycles in the North Pacific, such as decadal shifts in fish production, plankton biomass, and nutrient concentrations (Chavez et al., 2003; Francis and Hare, 1994; Mantua and Hare, 2002). In addition to these regional changes, the PDO has shown remote associations through atmospheric teleconnections with multi-decadal drought and pluvial conditions over the United States, Canada, Siberia, Australia, and northern South America (Barlow et al., 2001; Hu and Huang, 2009; McCabe et al., 2004; Taguchi et al., 2012; Tamimoto et al., 2003; Vance et al., 2015; Wang et al., 2014, 2009; Zanchettin et al., 2008). From the perspective of these broad impacts, improving our understanding of PDO mechanisms is essential not only for enhancing PDO predictability but also its application to marine ecosystem predictions (Chikamoto et al., 2015a, 2013; Kim et al., 2014; Mochizuki et al., 2010).

Two processes mainly categorize the PDO mechanisms: regional air-sea interactions in the North Pacific and remote forcings from other ocean basins (Kwon and Deser, 2007; Liu and Alexander, 2007; Newman et al., 2016), such as the tropical Pacific and Atlantic. While it is challenging to separate regional air-sea interactions and remote impacts on the PDO due to feedback mechanisms (Zhang and Delworth, 2015), there is a scientific consensus that ENSO variability is the primary forcing for the PDO on interannual timescales through the
"atmospheric bridge" and subsequent changes in Aleutian low variability (Alexander et al., 2004; Strong and Magnusdottir, 2009; Trenberth et al., 1998). Despite the previous research on the atmospheric bridge process through many observational analyses and modeling experiments (Alexander, 1990; Barlow et al., 2001; Deser et al., 2004; Lau and Nath, 1994, 1996; Nakamura et al., 1997), there is large model uncertainty regarding timing and strength of the ENSO-PDO relationship. In CMIP3 and CMIP5 models, maximum correlation coefficients of the ENSO-PDO relationship in pre-industrial control simulations range from 0.2–0.9 (Nidheesh et al., 2017), which has higher uncertainty compared to the observational estimates (∼0.6) (Alexander et al., 2008; Hu and Huang, 2009). In addition to strength, the time lag between the ENSO and PDO indices in CMIP3 and CMIP5 models ranges from almost instantaneous to 8 months as most models overestimate the observed ENSO-PDO relationship up to 1–2 months lag. As a result, there is still large uncertainty in the current estimate of PDO variability originating from the tropical Pacific or simply internal variability, which corresponds to roughly 25–50% of PDO variance (Alexander et al., 2002, 2010).

Uncertainty of the ENSO-PDO relationship may also come from the Atlantic remote forcing through inter-basin interactions. Recent studies pointed out that Atlantic decadal variability can affect tropical Pacific climate variability through a modulation of the global Walker circulation (Cai et al., 2019; Chikamoto et al., 2012; Kucharski et al., 2011, 2015; Li et al., 2015; McGregor et al., 2014; Rodríguez-Fonseca et al., 2009a). On interannual timescales, ENSO is also modulated by Atlantic SST variability such as the north tropical Atlantic variability and the Atlantic Niño (Cai et al., 2019; Ding et al., 2012; Ham et al., 2013a,b; Luo et al., 2010, 2017; Rodríguez-Fonseca et al., 2009a). Interestingly, most CMIP3 and CMIP5 models capture the inter-basin impact from the Atlantic to the Pacific, although the strength of the atmospheric response to Atlantic SST variability and the subsequent ENSO amplitude show vast diversity among the models (Ham and Kug, 2015).

The mechanisms of the Atlantic impact to the PDO are still under debate, but three major processes have been proposed: tropical inter-basin interactions, midlatitude jet stream
activity, and an Arctic Ocean process through the Bering Strait. As described above, Atlantic Ocean variability affects the frequency and amplitude of ENSO events through interbasin interactions (Ham and Kug, 2015), which subsequently influences the PDO through the atmospheric bridge (Alexander, 1990). Consistent with this mechanism, Levine et al. (2018) found that a positive Atlantic Multidecadal Oscillation (AMO) can favor a northward shift in the intertropical convergence zone (ITCZ) through changes in equatorial trade winds and precipitation anomalies, which then affect tropical Pacific SSTs. Furthermore, Sun et al. (2017) and Gong et al. (2020) found that the AMO causes anomalous low pressure and associated ascent in the North Atlantic and subsequent compensating anomalous descent and surface high pressure in the North Pacific with the strongest air-sea interaction in the tropics, which causes atmospheric and oceanic thermodynamic responses there. By contrast to these tropical inter-basin interactions, other studies proposed that the AMO can affect the PDO through midlatitude jet stream activities (Okumura et al., 2009; Wu et al., 2011; Zhang and Delworth, 2007). In this midlatitude process, three pathways are still under debate: a meridional shift in the midlatitude westerlies between the North Atlantic and Pacific (Zhang and Delworth, 2007), changes in the Arctic Oscillation originating from North Atlantic SSTs (Zhang and Zhao, 2015), and circumglobal Rossby wave propagation from the tropical Atlantic to the North Pacific via the South Asian jet (Okumura et al., 2009; Wu et al., 2011). The third process is through the Arctic and Bering Strait by introducing freshwater forcing in the Atlantic Meridional Overturning Circulation (AMOC) (Okumura et al., 2009). Because of these inter-basin and atmosphere-ocean interactions, appropriate modeling experiments using a fully coupled atmosphere-ocean general circulation model (AOGCM) are required to evaluate the most prominent pathway from the Atlantic to the North Pacific.

Whereas just a few studies simulated an Atlantic origin for PDO variability, many models captured the atmospheric bridge process, including CMIP, AMIP, mixed-layer, partial decoupling, pacemaker, and partial assimilation experiments. In the 1990s, AGCM experiments prescribing observed SST variability (so-called AMIP-type experiment; Lau and
Nath, 1994) evaluated the direct atmospheric responses to the tropical Pacific SST forcing. In the following papers, a role of thermodynamical air-sea interaction in the North Pacific was further assessed with subsequent AMIP-type experiments coupled with mixed-layer ocean models (e.g., Alexander et al., 2002; Di Lorenzo et al., 2015; Lau and Nath, 1996).

To assess a role of fully dynamical atmosphere-ocean interactions, Wang et al. (2012) proposed the partial decoupling experiment using the NCEP CFSv2 model, where SSTs are dampened toward climatological states in a targeted region, yet it was still challenging to isolate the remote ENSO impact on the PDO from regional air-sea interaction. To identify the climate variability originating from the tropical Pacific, a pacemaker experiment was proposed, in which the observed SST variability in the eastern tropical Pacific was prescribed in the ocean component of a climate model through heat flux forcing at the ocean surface (Kosaka and Xie, 2013). In these experiments, the ocean and atmosphere are able to evolve freely outside the region but can also interact inside the targeted region. Using this pacemaker experiment, Zhang et al. (2018) demonstrated that the atmospheric bridge concept is consistent with previous studies.

While the atmospheric bridge process from the tropical to the North Pacific has been well simulated by prescribing SST-only forcing into atmospheric general circulation models (AGCMs), a model experiment to capture the Atlantic impacts on the PDO is more challenging due to complicated ocean dynamics relating to the AMOC variability, which may introduce systematic errors in the Atlantic (Zhang and Zhao, 2015). Previous studies constrained the Atlantic Ocean using SST only, such as an idealized experiment by prescribing the AMO-induced SST pattern (Levine et al., 2018; Ruprich-Robert et al., 2017) or a pacemaker experiment targeted in the Atlantic (Ding et al., 2012). However, Boer et al. (2016) pointed out that pacemaker experiments forced by Atlantic SSTs may introduce energy and density imbalances due to a lack of salinity information, which causes artificial changes in atmosphere-ocean interaction and can alter the coupled model equilibrium. A recent study by Chikamoto et al. (2019) demonstrated that 3-dimensional temperature and salinity information is required to constrain the Atlantic to simulate AMOC variability. To
minimize the aforementioned artificial changes in the AOGCM, the partial assimilation approach is proposed here, in which 3-dimensional temperature and salinity anomalies in the targeted region are assimilated into the climate model. Using this partial assimilation approach, Johnson et al. (2018) identified an Atlantic modulation on Australian precipitation variability in the interannual-to-decadal timescale.

To more quantitatively assess how much of PDO variability originates from the equatorial Pacific and Atlantic basins, we conduct partial ocean data assimilation experiments using two state-of-the-art global climate models: the Community Earth System Model (CESM1.0) and the Model for Interdisciplinary Research on Climate (MIROC3.2m). By assimilating observed conditions in the global, equatorial Pacific or Atlantic Oceans, respectively, we can evaluate how much each ocean basin contributes to the PDO variability. More detailed model experiments and their validations are described in Sections 3.3 and 3.4. Next, we describe impacts from the tropical Pacific (Section 3.5) and the Atlantic on the PDO (Section 3.6), followed by a discussion and conclusion (Section 3.7).

3.3 Data and model experiments

3.3.1 Observations

This study examines observational SST datasets from the UK Met Office’s HadSST version 3 for 1950–2014 (Kennedy et al., 2011a,b), the National Centers for Environmental Prediction’s (NCEP) Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4) for 1960–2014 (Huang et al., 2015), and an objective ocean analysis compiled by the Japanese Meteorological Agency (JMA) (hereafter, referred to as ProjD) for 1950–2010 (Ishii and Kimoto, 2009). In addition, we use SLP datasets derived from HadSLP version 3 for 1950–2014, JMA’s JRA55 for 1960–2014 (Ebita et al., 2011), and NCEP derived SLP for 1960–2010. The analysis periods in ERSST/JRA55 and ProjD/NCEP are identical to those in the assimilation runs in the CESM and MIROC, respectively, while the HadSST/HadSLP spans both model time periods. By choosing these time periods, we can validate our model simulated SSTs and its associated air-sea interaction, as described in Section 3.4.
Table 3.1: Summary of model experiments

<table>
<thead>
<tr>
<th>Name</th>
<th>Model</th>
<th>Assimilated region</th>
<th>Period</th>
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<tbody>
<tr>
<td>CESM GLOB</td>
<td>CESM</td>
<td>Global oceans</td>
<td>1960–2014</td>
</tr>
<tr>
<td>CESM ATL</td>
<td>CESM</td>
<td>Atlantic (30°S–70°N)</td>
<td>1960–2014</td>
</tr>
<tr>
<td>MIROC GLOB</td>
<td>MIROC</td>
<td>Global oceans</td>
<td>1950–2010</td>
</tr>
<tr>
<td>MIROC eqPAC</td>
<td>MIROC</td>
<td>Equatorial Pacific (10°S–10°N)</td>
<td>1950–2010</td>
</tr>
<tr>
<td>MIROC ATL</td>
<td>MIROC</td>
<td>Atlantic (50°S–60°N)</td>
<td>1950–2010</td>
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observational datasets, the linear trends at each grid point are removed from the anomalies in order to remove the climate response to the radiative forcing (i.e., the global warming component). The linear detrend method is not perfect for removing the climate response to radiative forcing (Dong and McPhaden, 2017), but its impact is minor, as described in Section 3.4 (Model Validation).

3.3.2 Models

This study uses two fully-coupled AOGCMs: the CESM1.0 (Shields et al., 2012) and the MIROC3.2m (Nozawa et al., 2007). The CESM consists of 4 physical components for atmosphere, land, ocean, and sea ice (Holland et al., 2012; Hunke and Lipscomb, 2008; Lawrence et al., 2012; Neale et al., 2013; Rodríguez-Fonseca et al., 2009b; Smith et al., 2010). The atmosphere and land models are T31 horizontal resolution (approximately 3.75°) with 26 atmospheric vertical levels. The ocean and sea-ice models use a horizontal resolution of approximately 3° with a displaced North Pole and 60 ocean layers. Within the CESM, atmosphere, land, and sea-ice complete a timestep every 30 minutes, while the ocean couples with the atmosphere once per day. Details of the model’s basic performance in the configuration can be found in Shields et al. (2012) and Chikamoto et al. (2015b).

In the MIROC, atmospheric and land models use a T42 spectral horizontal grid (approximately 2.8°), while the ocean and sea-ice components of the MIROC use 43 vertical levels with a zonal resolution of approximately 1.41°. Its meridional grid is about 0.56° in the equatorial region but 1.4° at high latitudes. The physical processes in the ocean model of the MIROC include parameterizations for vertical convection and mixing, lateral mixing
of momentum and tracer, and vertical penetration of shortwave radiation. All components of the atmosphere, ocean, land, and sea-ice modules are coupled without any flux corrections for exchanging heat, water, and momentum fluxes between the atmosphere and ocean. Details of the performance and configuration are found in Nozawa et al. (2007).

3.3.3 Partial assimilation experiments

Using these models, we first conducted two model experiments: a historical run and a global ocean assimilation run (GLOB run). In the historical run, the model was prescribed by time-varying observed natural and anthropogenic radiative external forcings (solar, aerosols, land-use change, and greenhouse gases) from 1850 to 2005. After 2005, we prescribed the RCP4.5 emission scenario for the CESM and an A1B-type scenario for the MIROC. These experiments consist of 10 ensemble members from initial conditions compiled from 10 random years of pre-industrial control simulations. Historical runs for each model correspond to the externally forced component due to radiative forcing. In the GLOB run, we prescribed the same radiative forcings as the historical run but assimilated 3-dimensional observed ocean temperature and salinity anomalies into the global climate models. The assimilated observed ocean data was derived from the ECMWF ocean reanalysis product (version 4) for 1958–2014 (Balmaseda et al., 2013) in the CESM and the ProjD dataset for 1945–2010 (Ishii and Kimoto, 2009) in the MIROC. The reason for the difference in assimilated reanalysis products is because the CESM and MIROC were developed by different institutions, which are summarized in Table 3.1. In these observation-based datasets, the monthly mean was linearly interpolated into daily values. Next, analysis increments are estimated from a temporally, spatially, and vertically invariant model-to-observation ratio in analysis errors and added as forcing into the model’s temperature and salinity equations during the analysis interval of one day by an Incremental Analysis Update scheme (Bloom et al., 1996; Huang et al., 2002).

To identify the remote forcing from the equatorial Pacific and Atlantic to the PDO, we further conducted partial ocean assimilation experiments using these models: the eqPAC and ATL runs, respectively. In these partial assimilation experiments, we use the same
model configurations as the GLOB run, except the 3-dimensional observed ocean anomalies are assimilated only in the equatorial Pacific (CESM and MIROC: 10°S–10°N) and the Atlantic (CESM: 30°S–70°N; MIROC: 50°S–60°N) basins. Each partial assimilation run includes 10 ensemble members with different time ranges for the CESM (1960–2014) and MIROC (1950–2010) after a model spin-up (2 years for the CESM and 5 years for the MIROC). Applications of these partial ocean assimilation experiments and how they affect regional climate are described in previous studies (Chikamoto et al., 2015b, 2016; Ham et al., 2017; Johnson et al., 2018; Purich et al., 2016).

In this study, anomalies are defined as deviations from the climatological mean for 1960–2014 in the CESM and for 1950–2010 in the MIROC. To remove the externally forced component due to radiative forcing, we take the difference of ensemble-mean anomalies between the partial assimilation and the historical runs in each model, similar to Zhang et al. (2018). A 13-month running mean filter is applied in our analyses to reduce atmospheric stochastic variability for model and observational data.

3.4 Model validation

To define the PDO index as described in Mantua et al. (1997), we apply an empirical orthogonal function (EOF) analysis to monthly SSTAs in the North Pacific (20°–70°N) in observations and the GLOB runs (Fig. 3.1). Note that the linear trends in each grid point are removed in the observed SSTAs, whereas in the GLOB runs, SSTAs are obtained by subtracting the ensemble mean of the historical run in each model. The leading EOF modes (EOF1) for the North Pacific SSTAs in GLOB runs capture the observed PDO-like SSTA pattern very well (Fig. 3.1e and i): the cooler SSTs in the KOE region and its surrounding warmer SSTs along the west coast of North America (Mantua and Hare, 2002; Newman et al., 2016). The percentages of explained variance in EOF1 (Fig. 3.1) are well separated from the second and third modes (HadSST: 16.8%, 8.5%, 7.9%; ERSST: 22.3%, 14.9%, 8.0%; CESM GLOB: 29.8%, 14.7%, 7.8%; ProjD: 22.8%, 13.7%, 8.4%; MIROC GLOB: 28.6%, 13.9%, 6.2%, in first, second, and third modes, respectively). All of the second modes show an SSTA pattern similar to the North Pacific Gyre Oscillation (NPGO), as
described by Di Lorenzo et al. (2008). The GLOB runs of both the CESM and MIROC have comparable variances explained that are larger than the observations, and the monthly principal components are strongly correlated with each other \((R = 0.79)\) for the time period that overlaps between the two models (1960–2010). Whereas the principal components of EOF1 in observations are almost identical with each other for the 1950–2014 period \((R > 0.97)\), the GLOB runs in CESM and MIROC show slightly lower correlation coefficients with observations \((R = 0.88 \text{ and } 0.78 \text{ for HadSST, } R = 0.87 \text{ and } 0.81 \text{ for ERSSTv4, and } R = 0.87 \text{ and } 0.74 \text{ for ProjD, respectively})\). Because no atmospheric observations are assimilated into these models, higher-frequency PDO variability is filtered out based on the ensemble mean of 10 members in the GLOB runs. Alternatively, this lower correlation may arise from the linear detrending method to estimate the climate response to radiative forcing in observations. When we apply a 13-month running mean filter to the monthly principal components of EOF1, correlation coefficients of the GLOB runs in the CESM and MIROC with observations increase pronouncedly \((R = 0.93 \text{ and } 0.87 \text{ in HadSST, } R = 0.95 \text{ and } 0.85 \text{ in ERSSTv4, and } R = 0.94 \text{ and } 0.81 \text{ in ProjD, respectively})\). Additionally, the observations and GLOB runs show a similar total SSTA variance in the North Pacific after the running mean is applied, lending support that the GLOB runs capture similar variability in the North Pacific (Fig. 3.2). These results support our approach in capturing the interannual-to-decadal components of the North Pacific, including the PDO. Hereafter, a 13-month running mean filter is applied to all monthly anomalies, and the PDO index is defined as the 13-month running mean for the principal component of EOF1 (black lines in the right panels of Fig. 3.1).

To validate the Aleutian low variability in our models, we correlated sea level pressure anomalies (SLPAs) and SSTAs with the observed and GLOB PDO indices in Figure 3.3. As pointed out by previous studies (e.g., Newman et al., 2016), we find that the positive phase of the PDO (SST cooling in the KOE region) accompanies the strengthened Aleutian low variability (negative SLPAs in the North Pacific) in all of observations and GLOB runs \((Qiu et al., 2007)\). Consistent with the SLP patterns, the GLOB runs in the CESM and MIROC
Fig. 3.1: Regressed SSTA patterns (left panels) and principal components (right panels) associated with the leading EOF mode of monthly SSTAs in the North Pacific based on the (a, b) UK Met Office’s HadSST version 3 (1950–2014), (c, d) National Centers for Environmental Prediction’s ERSST version 4 (1960–2014), (g, h) Japanese Meteorological Agency ProjD (1950–2010) and the GLOB runs of the (e, f) CESM (1960–2014) and (i, j) MIROC (1950–2010). Units in regression plots are °C. The variance explained by the first EOF mode is indicated in the upper-right corner of each plot. Positive (negative) values are red (blue). A black line in the right column is a 13-month running average for the principal component, which is defined as the PDO index.
Fig. 3.2: 13-month running mean total SSTA variance (colors) and the ratio of monthly and 10-year low-pass filtered total SSTA variance (values exceeding 40% are stippled) in (a, b) observations, (c, d) GLOB, (e, f) eqPAC, and (g, h) ATL runs in the CESM (left) and MIROC (right). Observations include (a) ERSSTv4 and (b) ProjD. Light grey dots indicate 40% of the variance is in the decadal component, whereas dark grey dots indicate 50% of the variance is in the decadal component.
correlate with the observed SLP variability in the Aleutian low region (30° S–65° N, 160° E–140° W; $R = 0.41$ and 0.46, respectively), otherwise known as the North Pacific Index (*Trenberth and Hurrell*, 1994). In addition to Aleutian low variability, the PDO-like pattern accompanies positive SLPAs over the Indian and Australia region and negative SLPAs over the central and eastern tropical Pacific in the observations and GLOB runs. This zonal SLP gradient in the Indo-Pacific sector is closely related to El Niño-like SST warming in the tropical Pacific, which is the primary driver of the PDO on interannual timescales through the atmospheric bridge (*Alexander et al.*, 2002, 2010). Even though atmospheric data are not assimilated into the models, the GLOB runs did simulate the Aleutian low variability associated with the PDO index quite well, albeit overestimating positive SLPAs in the northwestern subtropical Pacific.

In the Atlantic basin, however, the SSTA and SLPAs associated with the PDO show more diverse patterns among observations and GLOB runs. The observations and CESM GLOB run (Fig. 3.3a–c and e) shows that the PDO index positively correlates with SLPAs in the tropical Atlantic and Indian Oceans, whereas the tropical Atlantic SLPA correlations are unclear in the MIROC GLOB run (Fig. 3.3d). Moreover, the MIROC GLOB run shows SST warming in the tropical Atlantic associated with a positive PDO-like pattern, which is differentiated from the weak SST cooling in observations and the CESM GLOB run. By examining the ProjD (Fig. 3.3c), we also note stronger SSTA correlations in the south tropical Atlantic compared to the other observations and GLOB simulations. In the North Atlantic, there are moderate negative SSTA correlations in observations and the CESM GLOB, which accompanies negative SLPA correlations. These discrepancies in the Atlantic may impact the atmospheric bridge given inter-basin impacts to the Pacific.

### 3.5 Tropical Pacific origin

#### 3.5.1 Equatorial Pacific impact on PDO

To identify the tropical Pacific impacts on the PDO, we apply an EOF analysis to the ensemble mean of North Pacific SST variability in the equatorial Pacific (eqPAC) runs
Fig. 3.3: Correlation maps of SLPAs (contours: solid and broken lines are positive and negative correlations, respectively) and SSTAs (colors) associated with the PDO index in the (a) ERSST and JRA55, (c) ProjD and NCEP, (e) HadSST and HadSLP, and the GLOB runs in the (b) CESM and (d) MIROC. The contour interval is 0.2 with the zero contour omitted.
Fig. 3.4: Regressed SSTA patterns (left panels) associated with the leading EOF mode of monthly SSTAs in the North Pacific (20°–70°N) in (a, b) the eqPAC and (c, d) ATL runs during 1960–2014 for the CESM (left) and 1950-2010 for the MIROC (right panels). Units are in °C. Note that the color scale is the same as Fig. 3.1. The variance explained by EOF1 in each run is denoted on the top right corner of each sub-panel.

(Figs. 3.4a-b and 3.5a-b). The leading EOF modes in these eqPAC runs show a PDO-like SST pattern in the North Pacific (Fig. 3.4), which is consistent with the peaks of total SSTA variance in the North Pacific (Fig. 3.2). The 2nd modes (not shown) in the eqPAC runs depict an NPGO-like feature in the CESM and an unclear SST signal in the MIROC (16.9%, 7.3% and 8.0%, 7.5% of explained variability for the 2nd and 3rd modes of the MIROC and CESM, respectively). Their variances explained by EOF1 are slightly larger than those in the GLOB runs and observations. The EOF1 principal components in the eqPAC runs also significantly correlate with the GLOB runs ($R = 0.63$ in CESM and 0.66 in MIROC, respectively). Consistent with the idea of the atmospheric bridge proposed by previous studies ([Liu and Alexander, 2007; Trenberth and Hurrell, 1994; Trenberth et al., 1998], in particular, the eqPAC runs capture the positive phase of the PDO during the early 1980s and 1990s, and its negative phase during the mid 1970s and early 2000s.

Associated with the atmospheric bridge, Aleutian low variability is the primary factor
Fig. 3.5: Principal components associated with the leading EOF mode of monthly SSTAs the North Pacific SSTAs (20°–70°N) in (a, b) the eqPAC and (c, d) ATL runs during 1960–2014 for the CESM (left) and 1950-2010 for the MIROC (right panels). The variance explained by the first EOF mode in each run is denoted on the top right corner of each sub-panel.

in connecting tropical Pacific SST forcing and the PDO-like SST response in the North Pacific. To examine these connections, we produced lead-lag correlations of the PDO index (the first principal components) with Aleutian low variability (SLPA in 30°N–65°N, 160°E–140°W) and the Niño 3.4 index (SSTAs in 5°S–5°N, 170°W–120°W) in the GLOB and eqPAC runs (black and blue lines in Fig. 3.6). The eqPAC runs show that Aleutian low variability negatively correlates with the PDO index (Fig. 3.6a and b), with the strongest correlation leading the PDO by 5 months. Figure 3.7 shows correlation maps of SSTA and SLPAs leading the PDO index by 5 months. Similar to the simultaneous correlation maps in Fig. 3.3, we find the strengthened Aleutian low (i.e., negative SLPAs) and PDO-like SST patterns in the North Pacific for all of the observations and model simulations (Fig. 3.7a–f). Associated with these North Pacific SST and SLP patterns, all of the observations and model simulations demonstrate the El Niño-like SST warming and a weakened Pacific Walker circulation. These results suggest that tropical Pacific SST variability is the main driver for the PDO-like SST response in the North Pacific, confirming the results of previous studies.

The ENSO-associated atmospheric bridge is also clearly delineated by the eqPAC runs.
Fig. 3.6: Lead-lag correlations between (a, b) the PDO and Aleutian low indices and (c, d) the PDO and Niño 3.4 indices for the GLOB (black), eqPAC (blue), and ATL (red) runs in (left) the CESM and (right) MIROC. Negative lag months indicate that (a, b) the NPI leads the PDO index and (c, d) the Niño 3.4 index leads the PDO, whereas the positive lag months indicate the opposite. The Y axis indicates correlation coefficients, and the X-axis shows months. A horizontal dashed line indicates a statistical significance at the 95% confidence level using the two-sided Student’s t-test.
We can find that the eqPAC runs of the CESM and MIROC show the strongest response of equatorial Pacific SSTAs with the PDO (Fig. 3.7e-f). The correlation coefficients between the PDO and the Niño 3.4 index in the eqPAC runs peak 0.75 in both the CESM and MIROC, respectively, when the Niño 3.4 index leads the PDO by 5 months (blue lines in Fig. 3.6c and d). The equatorial Pacific SST warming enhances local precipitation, which causes the Matsumo-Gill type atmospheric response (Taschetto et al., 2010); this is shown in a quadrupole structure of upper-tropospheric streamfunctions over the tropical Indo-Pacific sectors (Fig. 3.8a-b). Consistent with the atmospheric bridge process (Trenberth et al., 1998), this tropical Pacific SST warming drives the PNA-like wave train pattern in the upper-troposphere and its associated wave-activity flux (arrows): positive streamfunction anomalies at 250 hPa (corresponding to the higher pressure) with the PDO over the northern tropical Pacific and alternating lower and higher pressure centers in the North Pacific and North American regions, respectively (Fig. 3.8a-b). In particular, the upper-tropospheric low pressure center in the North Pacific consists of an equivalent barotropic structure and affects Aleutian low variability (Yu and Zwiers, 2007). Our eqPAC runs clearly demonstrate that tropical Pacific climate variability is the main driver for the PDO-like SST response with strong agreement between the CESM and MIROC, a feature that also lends credibility to our partial assimilation experiments.

3.5.2 Equatorial Pacific Impact on Atlantic

In the eqPAC runs, we see positive SSTA and positive correlations of SLPAs over the tropical Atlantic with the PDO index (Fig. 3.7e and f), and it suggests the equatorial Pacific impact on the tropical Atlantic basin. According to the previous studies (Chang et al., 2006; Chiang and Lintner, 2005; Chiang and Sobel, 2002; Chikamoto and Tanimoto, 2005; Sobel et al., 2002), El Niño events lead to tropospheric warming in the entire tropics due to atmospheric Kelvin-wave propagation, which then induces SST warming in the tropical Atlantic and Indian Oceans through humidity-driven latent heat fluxes (called the TT-mechanism). The El Niño event also induces the SST warming in the northern tropical Atlantic through the PNA-like wave propagation from the Pacific to the North Atlantic
Fig. 3.7: Correlation maps of SLPAs (contours: solid and broken lines are positive and negative correlations, respectively) and SSTAs (colors) at 5 months lead associated with the PDO index in (a, b) observations, (c, d) GLOB, (e, f) eqPAC, and (g, h) ATL runs in the CESM (left) and MIROC (right). Observations include (a) ERSSTv4 and JRA55 and (b) ProjD and NCEP. The contour interval is 0.2 with the zero contour omitted. A 13-month running mean filter is applied to all anomalies.
Fig. 3.8: Regression maps of streamfunction (contours: solid and broken lines are positive and negative regressions, respectively) and precipitation anomalies (green: positive; tan: negative) at 5 months lead time associated with the PDO index in (a, b) the eqPAC and (c, d) ATL runs in the CESM (left) and MIROC (right). Only regressed precipitation anomalies that have a corresponding correlation coefficient stronger than 0.27 are plotted. Precipitation has units in $\text{mm day}^{-1}$. The zonal mean was removed from the streamfunction anomaly (contour interval: $0.4 \times 10^6 \text{ m}^2\text{s}^{-1}$ in the eqPAC runs and $0.2 \times 10^6 \text{ m}^2\text{s}^{-1}$ in the ATL runs). Positive (negative) streamfunction anomalies indicate a clockwise (anti-clockwise) circulation. Vectors represent wave-activity flux (units: $\text{m}^2\text{s}^{-2}$) calculated from streamfunctions at 250 hPa regressed on the PDO index (Takaya and Nakamura, 2001). Note that wave-activity vectors are displayed only in the northern hemisphere. Reference vectors are denoted in the upper right-hand corner of each plot. Wave activity fluxes exceeding 0.02 and 0.05 $\text{m}^2\text{s}^{-2}$ are plotted in the CESM and MIROC, respectively.
(Cai et al., 2019; Wallace and Gutzler, 1981). Consistent with these mechanisms, the eqPAC runs (Fig. 3.7e and f) clearly demonstrate that the El Niño-like SST warming in the equatorial Pacific induces higher SLPs in the tropical Atlantic and Indian Oceans and a subsequent ocean response as seen in the SST warming in those oceans.

In observations, by contrast, we find tropical Atlantic SST cooling associated with the PDO index despite higher SLPs at the same location (Fig. 3.7a and b). Because SST cooling contributes to locally higher SLPs, the higher observed SLPs in the tropical Atlantic associated with the PDO index are mainly driven by the local SST cooling rather than the remote impact from the equatorial Pacific. In other words, our results suggest that the SLP variability in the tropical Atlantic has two components: a remotely forced component from the equatorial Pacific and a locally driven component within the tropical Atlantic. Whereas the former leads the same phases of SSTAs in the tropical Pacific and Atlantic, the latter has the potential to induce the opposite phases of SSTAs between the tropical Pacific and Atlantic, as seen in the observations and CESM GLOB run (Fig. 3.7a, b, and c). The discrepancy of tropical Atlantic SSTAs between observation and eqPAC runs suggest that the Atlantic has its own variability independent from the tropical Pacific, which opens up the possibility of the Atlantic impact to the PDO.

### 3.6 Atlantic impacts to the PDO

To identify the Atlantic impacts on the PDO, we apply an EOF analysis of the ensemble mean of North Pacific SST variability in the Atlantic (ATL) partial assimilation runs (bottom panels in Figs. 3.4 and 3.5). We also examined the 2nd and 3rd modes of the ATL runs (not shown) and found the only PDO signal was in the 1st mode with well separated explained variance (CESM: 33.1%, 18.9%, 7.7%; MIROC: 28.6%, 9.1%, 7.2%). As one might expect, there are significant differences between the eqPAC and ATL runs. The EOF1 principal components in the ATL runs also significantly correlate with the GLOB runs ($R = 0.54$ in the CESM and 0.35 in the MIROC, respectively), and there are weak correlations of the first principal components between the ATL and eqPAC runs ($R = 0.26$ in the CESM and 0.20 in the MIROC, respectively). These significant ATL correlations
with the GLOB runs, but not the eqPAC runs, supports our hypothesis that Atlantic Ocean variability, which has a unique timescale from the Pacific counterpart, can trigger the PDO-like SST pattern. In the ATL runs, we can find negative SSTAs in the tropical Atlantic and Indian Oceans in both the CESM and MIROC (Fig. 3.7g-h), which is differentiated from the eqPAC runs (Fig. 3.7e-f). This sharp SSTA contrast suggests that the inter-basin climate interaction originating from the Atlantic Ocean also induces the PDO-like SST response. It is noteworthy that in the ATL run, the SST warming in the equatorial Pacific is stronger in the MIROC than the CESM, whereas the CESM shows AMO-like SST cooling in the North Atlantic more clearly than the MIROC. Additionally in the ATL runs, both models show total SSTA variance in the North Pacific, but the CESM shows a larger portion of total SSTA variance in the decadal component compared to the MIROC (dotted region in Fig. 3.2g-h). These different characteristics between the CESM and MIROC ATL runs imply two inter-related pathways from the Atlantic to the Pacific, as described in the following subsections.

3.6.1 Equatorial pathway

According to previous studies (e.g., Cai et al., 2019; Chikamoto et al., 2012, 2015c; Ding et al., 2012; Kucharski et al., 2015; Li et al., 2015; McGregor et al., 2014; Rodríguez-Fonseca et al., 2009a; Ruprich-Robert et al., 2017), Atlantic SST variability modifies tropical inter-basin climate interactions between the Pacific and Atlantic Oceans through the reorganization of the global Walker circulation. Consistent with this idea, we can find cooler SST and higher SLP in the tropical Atlantic and warmer SST and lower SLP in the tropical Pacific in both CESM and MIROC ATL runs (Fig. 3.7g-h). These SSTA/SLPA contrasts between the Pacific and Atlantic accompany less precipitation in the eastern tropical Pacific but more precipitation in the central tropical Pacific (Fig. 3.8c-d), suggesting the reorganization of the global Walker circulation. As shown by Hovmöller diagrams (Fig. 3.9) of SST, SLP, precipitation, and upper-tropospheric zonal wind anomalies at the equator (5°S-5°N) associated with the PDO indices in the ATL runs, SSTA and SLPA contrasts between the Pacific and Atlantic involve upper-tropospheric westerly anomalies over the eastern Pacific
and Atlantic and easterly anomalies over the western and central Pacific (Fig. 3.9b and d). These features highlight the equatorial pathway of inter-basin effects from the Atlantic to the tropical Pacific.

This reorganization of the global Walker circulation through the equatorial pathway is more prominent in the MIROC ATL run compared to the CESM ATL run. In the MIROC ATL run (Fig. 3.9a), we see SST cooling in the equatorial Atlantic that initiates about 36 months before the mature stage of the PDO. This Atlantic SST cooling accompanies locally higher SLP and then its eastward propagation from the Atlantic to the Indian Ocean. The SST cooling and higher SLPs cause decreases in precipitation over the Atlantic-Indian region (Fig. 3.9a and b) and its compensating atmospheric dynamical response in the equatorial Pacific, as seen in warmer SSTs, lower SLPs, and more precipitation. These atmospheric changes imply that the Atlantic SST forcing triggers strong atmosphere-ocean interactions in the equatorial Pacific, such as the Bjerknes feedback. The resultant precipitation changes in the equatorial Pacific lead to the Matsuno-Gill type atmospheric response as seen in the equatorial anti-symmetric structure of the upper tropospheric streamfunction anomalies over the tropical Pacific (Fig. 3.8d), which induces PNA-like Rossby-wave propagation (arrows) and a PDO-like atmosphere-ocean response in the North Pacific.

In the CESM ATL run, however, equatorial Pacific SST and SLP responses to the Atlantic forcing are much weaker and have shorter timescales compared to the MIROC ATL run (Fig. 3.7g and 3.9c). For example, we can find the CESM ATL run shows local peaks of Atlantic SSTA and SLPAs 6–18 months before the mature stage of PDO (Fig. 3.9c), which is a shorter time-lag compared to the MIROC ATL run that has a 36 month lag. These Atlantic anomalies accompany immediate but weaker responses of equatorial Pacific SSTA and SLPAs (Fig. 3.9c), suggesting weaker Bjerknes activation. In addition to the equatorial Pacific responses, we find totally opposite phases of upper-tropospheric zonal wind anomalies over the Indian Ocean among the MIROC and CESM ATL runs (Fig. 3.9b and d). Because of this weaker equatorial Pacific response, the upper-tropospheric streamfunction anomalies are unclear over the Indian Ocean in the CESM ATL run (Fig. 3.8c), compared
Fig. 3.9: Hovmöller diagrams for the lead-lag correlation of (a, c) SLP (contours), SST (colors), (b, d) 250 hPa zonal wind (contours), and precipitation anomalies (colors) averaged over the 5°S–5°N latitude band with the PDO index in the MIROC (top) and CESM (bottom) ATL runs. Contour interval is 0.2 with the zero contour omitted. Solid and broken contours are positive and negative correlations, respectively. A 13-month running mean filter is applied to all anomalies.
to the quadrupole structure in the MIROC ATL run (Fig. 3.8d). Nevertheless, the PDO index in the CESM ATL run is significantly correlated with the CESM GLOB run and displays a multidecadal timescale (Fig. 3.5c). These CESM results suggest that, in addition to the equatorial pathway, there is another pathway to explain the Atlantic influence on the North Pacific.

3.6.2 North tropical pathway

As described in the previous section, the PDO index in the CESM ATL run has multidecadal components with negative phases during the 1960s and 2000s but positive phases during the 1970s and 1980s (Fig. 3.5c), which reminds us of the temporal variations of the AMO (Knight et al., 2005; Trenberth and Shea, 2006). In fact, we can find significant instantaneous correlations between the PDO index in the CESM ATL run and the AMO index in observations ($R = -0.60$), but even stronger if there is a 7–9 month lag ($R = -0.67$). Previous literature also pointed out that the PDO is modulated by the remote impact of the AMO (Gong et al., 2020; Kucharski et al., 2015; Okumura et al., 2009; Sun et al., 2017; Wu et al., 2011; Zhang and Delworth, 2007). In addition to the multi-decadal timescales, the CESM ATL run shows AMO-like SST cooling in the North Atlantic associated with the PDO index (Fig. 3.7g) and a large percentage of variance in the decadal time scale (Fig. 3.2g). Therefore, these results lead us to examine the dynamical process that connects the AMO and PDO in the CESM ATL run more closely.

To examine the impact of the AMO on the PDO, we performed a lead-lag correlation between the AMO and PDO indices in the CESM and MIROC partial assimilation runs (Fig. 3.10). The AMO index is defined as SSTAs over the North Atlantic (0°–60°N, 0°–80°W) subtracted from the global mean (60°S–60°N) SSTAs as described by Trenberth and Shea (2006). In the GLOB simulations (black lines in Fig. 3.10a and b), there is a long period of negative correlations when the AMO leads the PDO and subsequent long-term positive correlations when the AMO lags the PDO, which is consistent with observed correlations (see Fig. 2 and 3 in Wu et al. (2011)). Additionally, we note the statistically significant correlations 11–12 years leading the PDO in the GLOB runs (Wu et al., 2011; Zhang and...
Delworth, 2007); however, these correlations disappear in the CESM and MIROC ATL runs where we find a local peak of correlations when the AMO leads the PDO by 9 months ($R = -0.77$ and $-0.36$, respectively). Using the 9 month lag timeframe, we made correlation maps of SST and SLP and regression plots of precipitation and 250 hPa streamfunction anomalies at 9 months lag with the AMO index in the ATL runs (Fig. 3.11). We find positive correlations of SSTAs extending from the equator to Greenland (shaded in Fig. 3.11a and b), highlighting the strong positive phase of the AMO. This SST warming accompanies negative SLPA correlations over the entire North Atlantic, particularly in the CESM ATL run (Fig. 3.11a). In the tropical North Atlantic, we can find increases in both precipitation and upper-level pressure (positive streamfunction anomalies) over and north of the warm SSTs (Fig. 3.11c and d), similar to the findings of Levine et al. (2018) and Wu et al. (2019), where they found that a positive AMO leads to a northward shift in the ITCZ and global Hadley circulation. These results suggest that the SST warming in the tropical North Atlantic associated with the positive phase of the AMO enhances local precipitation activity and leads to the atmospheric dynamical response, such as the baroclinic structure in the tropical North Atlantic and Rossby wave propagation toward extratropical Atlantic. The similar but opposite signs for baroclinic structure and extratropical Rossby wave propagation are also found in the northwestern tropical Pacific in the CESM ATL run (Fig. 3.11c). At 15°N, the SST cooling accompanies the decrease in precipitation, higher SLP, and upper-level lower pressure (negative streamfunction anomalies) over the northern tropical Pacific, which is consistent with Levine et al. (2018). These features, which are unclear in the eqPAC runs of both the CESM and MIROC, highlight the strong north tropical contrast between the Atlantic and Pacific associated with the AMO.

To further examine the inter-basin connection, we produced Hovmöller plots of SST, SLP, precipitation, and 250 hPa zonal wind anomalies over the north tropics (10°–20°N) correlated with the AMO index (Fig. 3.12). Consistent with the correlation maps of SSTAs and SLPAAs with the AMO index (Fig. 3.11), we see significant inter-basin contrasts of SSTAs and SLPAAs between the subtropical Atlantic and Pacific, particularly in the CESM
Fig. 3.10: Lead-lag correlations between the PDO and AMO indices in the GLOB (black), eqPAC (blue), and ATL (red) runs for the (a) CESM and (b) MIROC. Negative (positive) months lag indicates that the AMO leads (lags) the PDO. A horizontal dashed line indicates a statistical significance at the 95% confidence level using the two-sided Student’s t-test.

Fig. 3.11: Correlation maps of SLPAs (contours in top) and SSTAs (shaded in top), and regression maps of precipitation anomalies (green: positive; tan: negative) and 250 hPa streamfunction anomalies (contours in bottom panels), with the AMO index at 9 months lag for (a, c) CESM and (b, d) MIROC ATL runs. Precipitation has units in mm day$^{-1}$. Vectors (c, d) represent 250 hPa wave-activity flux (units: $m^2 s^{-2}$) calculated from streamfunctions (contour interval: $0.2 \times 10^6 m^2 s^{-1}$) regressed on the AMO index. Solid and broken contours are positive and negative correlations, respectively. A 13-month running mean filter is applied to all anomalies. Wave-activity fluxes exceeding 0.01 $m^2 s^{-2}$ are plotted. Note that wave-activity vectors are displayed only in the northern hemisphere.
ATL run (Fig. 3.12c). Precipitation and upper-level zonal wind anomalies also show the inter-basin contrast in the CESM ATL run with positive precipitation and negative 250 hPa zonal wind anomalies extending from the Atlantic into the eastern Pacific (Fig. 3.12d). These features suggest that SST warming in the tropical North Atlantic associated with the positive phase of the AMO causes lower SLP, an increase in precipitation, and upper-level easterly wind anomalies over the Caribbean Sea, which leads the atmospheric dynamical response in the subtropical North Pacific. In fact, the timing for correlation peaks of SST cooling, decreases in precipitation, and upper-level westerly wind anomalies in the subtropical Pacific are several months later than the correlation peaks in the Atlantic (Fig. 3.12c and d). This zonal dipole in SLPAs associated with the AMO has been reported by Sun et al. (2017) and Gong et al. (2020), particularly in the north tropical Atlantic and Pacific. In the MIROC ATL run (Figs. 3.11b, 3.11d, 3.12a, and 3.12b), we can find the same changes of SST, SLP, precipitation, and upper-level zonal wind anomalies as the CESM ATL run, albeit much weaker in amplitude. The reason for this weaker AMO-PDO relationship in the MIROC ATL run is unclear and may arise from model systematic error, mean-state biases, and model climate sensitivity.

3.7 Discussion and conclusion

Our partial ocean assimilation approach demonstrates that the tropical Pacific is the primary driver for PDO variability, consistent with the atmospheric bridge process proposed by previous studies (Alexander et al., 2010; Alexander and Scott, 2008; Miller et al., 1994; Newman et al., 2016). According to our eqPAC runs, about 40–44% of PDO variability originates from the tropical Pacific, similar to the findings of Alexander et al. (2002) and Liu and Alexander (2007). Although the ENSO-PDO relationship shows uncertainty in the GLOB runs with a correlation ranging in the CESM and MIROC from 0.60–0.65 and a time lag ranging from 1–4 months, this relationship becomes more consistent in the CESM and MIROC eqPAC runs with a correlation value of 0.75 and a time lag of 5–7 months. Such differences between the GLOB and eqPAC runs support our hypothesis that the Atlantic remote forcing affects the ENSO-PDO relationship through tropical inter-basin interactions.
Fig. 3.12: Hovmöller diagrams for the lead-lag correlation of (a, c) SLP (contours), SST (colors), (b, d) 250 hPa zonal wind (contours), and precipitation anomalies (colors) averaged over the 10°N–20°N latitude band with the AMO index in the MIROC (top) and CESM (bottom) ATL runs. Contour interval is 0.2 with the zero contour omitted. Solid and broken contours are positive and negative correlations, respectively. A 13-month running mean filter is applied to all anomalies.
The results of the eqPAC partial assimilations show the following process: equatorial Pacific SST warming enhances local convection activity and then causes a Matsuno-Gill type atmospheric response in the tropical Indo-Pacific region. This atmospheric forcing, in the form of a PNA-like pattern, contributes to the strengthening of Aleutian low variability. As a result, a PDO-like SST pattern emerges and persists on interannual-to-decadal timescales.

One of the novel results of this study is that the Atlantic Ocean also affects PDO-like SST variability through two processes: the equatorial and north tropical pathways (Fig. 3.13). In the equatorial pathway, Atlantic Ocean variability induces changes in the equatorial Pacific through the reorganization of the global Walker circulation, which further affects the PDO through the atmospheric bridge process, as shown by the eqPAC runs (Fig. 3.13a). This mechanism of the equatorial pathway is consistent with previous studies about the inter-basin interaction between the Atlantic and Pacific (Cai et al., 2019; Chikamoto et al., 2012, 2015c; Ding et al., 2012; Kucharski et al., 2015; Li et al., 2015; McGregor et al., 2014; Rodríguez-Fonseca et al., 2009a; Ruprich-Robert et al., 2017). This equatorial pathway is especially evident in the MIROC ATL run but comparably weak in the CESM ATL run. In contrast to the equatorial pathway, the north tropical pathway emphasizes an AMO-induced inter-basin connection between the north tropical Atlantic and Pacific Oceans (Fig. 3.13b). In the tropical North Atlantic (10°–20°N), warm AMO SSTs cause increases in precipitation and lead to higher upper-tropospheric pressure through a Matsuno-Gill type atmospheric response (Gill, 1980). Simultaneously, decreases in precipitation and lower pressure in the upper-troposphere emerge over the northwestern subtropical Pacific through local atmosphere-ocean interactions modulated by the Atlantic forcing. This teleconnection in SLPAs corresponds to a zonal dipole in velocity potential forced by North Atlantic SSTs found in Sun et al. (2017) and Gong et al. (2020). These north tropical atmospheric changes in the Atlantic and Pacific provide energy sources for atmospheric Rossby wave propagation toward the North Pacific resembling the Western Pacific (WP) pattern (Wallace and Gutzler, 1981) rather than the PNA pattern. Overall, our ATL runs suggest that 12–29% of PDO variability originates from Atlantic Ocean variability. Given the strong
correlations between the AMO and the PDO and the associated multi-decadal signal, it is suggested that persistent AMO forcing may provide a source for the low-frequency component of the PDO variability compared to higher frequency forcing from the equatorial Pacific (Fig. 3.2g). The north tropical pathway associated with the AMO takes about 1 year to fully impact the PDO, while the equatorial pathway takes about 3 years originating from symmetric equatorial Atlantic SSTAs.

However, a question remains what accounts for these 1- and 3-year time lags in north tropical and equatorial pathways from the tropical Atlantic to PDO. According to the atmospheric bridge process in the eqPAC runs, it takes about half a year for ENSO to impact the PDO. In the north tropical pathway, therefore, it is suggested that the process from the tropical Atlantic to the tropical Pacific also takes about half a year. In the equatorial pathway, by contrast, the tropical Atlantic-Pacific process takes significantly longer. In the MIROC ATL run, tropical Atlantic SST cooling and associated high pressure persist 36 months before the PDO with significantly delayed responses in upper-tropospheric westerlies and precipitation over the equatorial Atlantic (Fig. 3.9b). These delayed atmospheric responses to the tropical Atlantic SST forcing suggest that, in the MIROC, persistent equatorial Atlantic SST forcing is important for driving local atmospheric response in order to overcome the remote influence from the tropical Pacific. This stronger remote influence in MIROC may explain why the MIROC GLOB run shows tropical Atlantic SST warming associated with PDO, instead of the tropical Atlantic SST cooling demonstrated by observations and CESM GLOB run (Fig. 3.7a-d). This result suggests that uncertainty in our study relates to the model dependency with respect to the strength of the atmospheric response in the tropical Atlantic accounting for local SST forcing and remote impacts.

The equatorial pathway and the north tropical pathway in this study may not be independent of each other. In this study, we find that the PDO index correlated with SSTAs in the ATL runs (Fig. 3.7g-h) show an AMO pattern with an asymmetric SST response in the equatorial Atlantic in the CESM. In contrast, the MIROC shows a symmetric equatorial response in both SSTs and SLPs in the equatorial Pacific and Atlantic sectors, which
Fig. 3.13: Schematic figure of (a) the equatorial and (b) north tropical pathways for the Atlantic impact on the PDO. The equatorial pathway highlights the Walker circulation (black vertical and horizontal arrows) and a Rossby wave train. The north tropical pathway indicates the correlation maps of SSTA and SLPAs (bottom) and 250 hPa anomalies of streamfunctions, precipitation, and wind vectors (top) with the AMO index at 9 months lag.
accompany Kelvin wave-like propagation in Figure 3.9a. Additionally, in the equatorial pathway, we see an equivalent barotropic response in 250 hPa streamfunctions and SLPs in SPDO (South Pacific decadal oscillation), whereas in the north tropical pathway, we only clearly see an equivalent barotropic response in the North Pacific. In the equatorial pathway, the forcing mechanism for Rossby wave propagation shows symmetric precipitation in the central and western tropical Pacific (Fig. 3.8c-d); whereas in the north tropical pathway, the ITCZ is shifted further north due to the hemispheric energy imbalance associated with the positive AMO and penetrates into the tropical Pacific providing a source for Rossby wave propagation (Levine et al., 2018). Despite these differences, both processes show a Matsuno-Gill type circulation modulated tropical Atlantic SSTs: the north tropical pathway and the equatorial pathway correspond to the AMO and north tropical Atlantic SSTs, respectively (Keenlyside et al., 2013; Servain et al., 1982). In fact, the Atlantic Niño index (SSTAs in $3^\circ$S–$3^\circ$N, $20^\circ$W–$0^\circ$W) is correlated with the PDO index in the MIROC ATL run ($R = 0.48$). Further model experiments are needed to determine if the two processes are independent of each other.

Previous studies found that the AMO negatively correlates with the low-frequency component of the PDO at an 11–12 years lag in addition to a higher frequency component at a 1 year lag (Wu et al., 2011; Zhang and Delworth, 2007). Whereas our CESM and MIROC GLOB runs also show a statistically significant peak of negative correlation coefficients between the AMO and PDO indices at about $-11$ years (i.e., the AMO leads the PDO), this peak disappears in both CESM and MIROC ATL runs (Fig. 3.10). Regarding the higher frequency component at the 1 year lag, the CESM ATL run shows Rossby wave propagation from the north tropical Atlantic to Greenland (Fig. 3.10) but no propagation toward the North Pacific across the Eurasian continent (Okumura et al., 2009). The tropical atmospheric response to the AMO in the CESM ATL run resembles a northward shift in the ITCZ described by Levine et al. (2018). Therefore, our result supports the north tropical pathway through inter-basin interaction rather than a midlatitude pathway through jet stream activities (Okumura et al., 2009; Zhang and Zhao, 2015; Zhang and Delworth, 2007),
or an Arctic Ocean pathway through the Bering Strait (Okumura et al., 2009). However, there is uncertainty in the north tropical pathway in terms of model and seasonal dependencies as Zhang and Delworth (2007) found that the AMO can cause changes to winter time changes in the mid-latitude storm track, whereas this study does not take into consideration seasonality. To better understand the robustness of the AMO-PDO relationship, more research is needed, including inter-model comparisons using CMIP6 pacemaker experiments (Boer et al., 2016; Ruprich-Robert et al., 2017) as well as partial assimilation experiments using many other climate models.

This study focuses specifically on the North Pacific region, but the PDO is related to the Interdecadal Pacific Oscillation (IPO), which spans both northern and southern hemispheres. Since the tropical Pacific strongly impacts southern hemispheric Pacific decadal variability (Folland et al., 2002; Newman et al., 2003; Power and Colman, 2006), our mechanism regarding the Atlantic impact on the PDO may be applicable to explain SPDO variability. Correlation coefficients between the IPO Tripole Index (Henley et al., 2015) with a 13-month running average and our PDO index in ERSST, HadSST, and ProjD are 0.73, 0.75, and 0.73, respectively. Our ATL CESM and MIROC runs also show significant correlations between those indices (0.76 and 0.75), suggesting that our ATL simulations may force decadal variability in the entirety of the Pacific ocean, but future studies also need to confirm this relationship.

As others have noted (e.g., Newman et al., 2016), the PDO represents a combination of many atmosphere-ocean processes originating from the global ocean basins. Even though tropical Pacific climate variability is the primary driver for the PDO, it is also modulated by multi-basin interactions originating from the Atlantic, as shown by this study. The tropical Pacific and Atlantic basins explain roughly 52–73% of PDO variability; however, the remaining 27–48% may originate from regional ocean-atmosphere interactions in the North Pacific itself through locally driven stochastic forcing (Frankignoul and Hasselmann, 1977; Hasselmann, 1976), a reemergence mechanism (Alexander and Deser, 1995; Alexander et al., 1999; Hanawa and Sugimoto, 2004), and wind-driven adjustments of the ocean gyre
circulation (Miller et al., 1998; Taguchi et al., 2007). Additionally, the Indian Ocean has a potential to affect PDO through the inter-basin interaction between the Indian and tropical Pacific Oceans (Dong and McPhaden, 2018; Izumo et al., 2010, 2014; Kug and Kang, 2006; Saji and Yamagata, 2003; Yu et al., 2002). Although we suggest a total of 52–73% of PDO variability originates from the Atlantic and Pacific independently, this percentage may be less because the Atlantic and Pacific interact with each other, consequently causing an overlapping of explained variance among the two basins. On decadal timescales, specifically, the Atlantic and the Indian Ocean play active roles for tropical Pacific climate variability (Cai et al., 2019), which can affect the PDO through tropical-extratropical teleconnections. The potential for these timescales to be affected by climate change with respect to inter-basin impacts on the PDO needs to be further investigated (Zhang and Delworth, 2016). More studies regarding multi-basin interactions will contribute to advancing our understanding of PDO mechanisms and predictability, as well as the recent "hiatus" of global warming (Trenberth, 2015), which also involved inter-basin interactions.
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CHAPTER 4
PRECIPITATION TRENDS IN THE INDO-PACIFIC SECTOR FOR THE 1982-2014 PERIOD

Abstract

The globally averaged sea surface temperature (SST) shows a notable multidecadal warming trend during the second half of the 20th century, but its warming rate considerably weakened since the 1980s. This global warming hiatus event is mainly attributed to a combination of global SST warming due to external radiative forcing and a La Niña-like tropical Pacific SST cooling pattern due to internal climate variability. However, the precipitation trend pattern during this period is still obscured because of the unknown precipitation response to the combined SST forcings. Here we demonstrate how precipitation trends in the Indo-Pacific sector respond to a combination of external radiative, tropical Pacific, and Indian Ocean forcings from the 1980s through the 2010s. Linear trends of satellite-based observations depict a wetting trend in the western tropical Pacific and a drying trend in the eastern tropical Pacific associated with a La Niña-like SST cooling trend, supporting the warmer-get-wetter and colder-get-drier hypothesis. In the Indian Ocean, the observed precipitation trend pattern shows an interhemispheric contrast with a wetting trend in the north and a drying trend in the south. However, the cause of this interhemispheric contrast is unclear given the monopole sea surface temperature warming trend in the Indian Ocean, indicating a minor role of local ocean forcing. We propose climate model experiments to identify the precipitation responses to the radiative, tropical Pacific, and Indian Ocean forcings and additional analyses to reveal the physical mechanisms for the Indian Ocean precipitation trend pattern.
### 4.1 Introduction

Since the early 1980s, Earth’s global water cycle has undergone pronounced precipitation changes under a warming climate. The underpinning question is to what extent has external radiative forcing and internal climate variability contributed to these precipitation changes. From the 1980s through the 2010s, the combination of external radiative forcing and low-frequency internal climate variability induced many unusual regional precipitation changes over the Indo-Pacific sector (Balaji et al., 2018; Tu et al., 2009) and an amplified SST warming trend in the western Pacific (Chikamoto et al., 2012; Ding et al., 2013). These climate shifts include low-frequency internal variability characterized by the Interdecadal Pacific Oscillation (IPO) phase shift, from positive to negative, that began in the early 1980s and lasted through the mid 2010s, leading to the emergence of SST cooling in the eastern tropical Pacific. Combined with the externally forced SST warming trend, this tropical Pacific SST cooling is responsible for a noteworthy decrease in the average global warming rate from the late 1990s through the mid 2010s, termed the ”global warming hiatus” (England et al., 2014; Hartmann et al., 2013; Kosaka and Xie, 2013). Although these climate shifts may lead to water cycle changes, it is still unclear what precipitation trend patterns emerge in the Indo-Pacific sector during these recent decades. To answer that question, this study aims to identify the precipitation responses to radiative, tropical Pacific, and Indian Ocean forcings from the early 1980s to the mid 2010s.

The tropical precipitation response to external radiative forcing generally follows the ”warmer-get-wetter” and ”colder-get-drier” hypothesis (Xie et al., 2010). In this hypothesis, we assume that relative humidity stays constant in the tropics so that the atmosphere can contain more water vapor under warmer temperatures due to the Clausius-Clapeyron relationship. Regional SST warming can warm and moisten the lower atmosphere, which induces an unstable atmosphere and a source for convective instability. The SST warming also accompanies lower sea level pressure (SLP), which can trigger atmospheric upward motion, low-level moisture convergence, and increased precipitation (Lindzen and Nigam, 1987). The opposite is true for colder SSTs and decreased precipitation. As a result,
warming SST trends favor positive precipitation trends, whereas cooling SST trends prefer negative precipitation trends. In the tropical Pacific, the external radiative forcing causes a peak of SST warming along the equator, which provides a source for enhanced precipitation according to the warmer-get-wetter framework (Cai et al., 2019; Deser et al., 2010; Liu et al., 2005; Xie et al., 2010). The Indian Ocean has warmed much faster than the tropical Pacific since the early 1980s, possibly leading to a greater wetting trend compared to tropical Pacific precipitation trends (Dhame et al., 2020; Hu and Fedorov, 2019; Schott et al., 2009; Zhang et al., 2019).

In addition to external radiative forcing, low-frequency climate variability can influence precipitation trends in the Indo-Pacific sector during the 1980s through the 2010s. For instance, the recent IPO phase change accompanied downward motions over the eastern Pacific but upward motions over the western Pacific due to the strengthened Walker circulation, which exhibits an east-west contrast of precipitation trends and a La Nina-like SST trend pattern in the tropical Pacific (Dong et al., 2016). The upward branch of the strengthened Walker circulation may favor SST cooling in the Indian Ocean due to increased cloudiness and reduced net solar radiation, which counteracts external radiative forced SST warming (Dong and McPhaden, 2017). This inconsistency of Indian Ocean SST trends between external radiative forcing and the IPO remote impact makes it difficult to reveal precipitation trend patterns during the recent three decades.

The precipitation trends in the Indian Ocean may also rely on the phase shift of local climate modes, such as an Indian Ocean Basin (IOB) mode, an Indian Ocean Dipole (IOD), and the South Asian monsoon (Klein et al., 1999; Saji et al., 1999; Webster et al., 1999). The IOB features a basin-wide SST and precipitation pattern, while the IOD highlights a zonal contrast of SST and precipitation. These climate modes may also feedback onto the Pacific, potentially leading to changes in precipitation there (Cai et al., 2019). The Indian Ocean is unique compared to the Pacific because the South Asian monsoon highlights a seasonal reversal of winds bringing rich southerly moisture towards the Indian subcontinent during the summer and dry continental northerly winds during the winter (Flohn, 1957;
Ramage, 1971; Wu et al., 2012). During the summer, convection activities associated with the South Asian monsoon are seen extending from the Arabian Sea to the Bay of Bengal (Chen et al., 2005; Yoon and Chen, 2005). Since the South Asian monsoon relies on the thermal contrast between ocean and land to initiate moist southerly flow through meridional sea level pressure (SLP) gradients, rapid external radiative forced warming of the Indian Ocean may reduce the land-sea thermal contrast, thus monsoonal rainfall (Roxy et al., 2015). Consequently, a question remains regarding which components regulate the precipitation trend pattern in the Indo-Pacific sector.

The purpose of this study is to determine how the precipitation trend in the Indo-Pacific sector responds to a combination of radiative, tropical Pacific, and Indian Ocean forcings from the 1980s to the 2010s. The contributions of these components are estimated from observed data analyses and global climate model experiments. In the model experiments, we prescribe historical records of natural and anthropogenic radiative forcings and a future emission scenario into the Community Earth System Model (CESM) historical run. Precipitation responses to the combination of radiative and ocean forcings are assessed by incorporating the observed ocean information into CESM while prescribing the radiative forcings (the ocean data assimilation run). To isolate the precipitation responses to the tropical Pacific and Indian Ocean forcings, this study utilizes partial ocean assimilation experiments (see following section).

4.2 Data and model experiments

4.2.1 Observations

This study uses observational datasets from National Oceanic and Atmosphere Administration (NOAA) Extended Reconstruction Sea Surface Temperature (ERSST) (Huang et al., 2015), Optimum Interpolation (OI) (Reynolds et al., 2002), and the Japan Meteorological Agency’s Centennial Observation-Based Estimate (COBE) datasets (Ishii et al., 2005). ERSST and OI reanalysis datasets use ships and buoys, whereas the COBE dataset incorporates some satellite data. For precipitation, we use the Climate Prediction Center’s
(CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin, 1997) and the Global Precipitation Climatology Project (GPCP) (Adler et al., 2003; Huffman et al., 1997). Both of these datasets incorporate rain gauges and satellite data but use different data sources and assimilation techniques. For example, CMAP uses gauges on southeastern Pacific atolls, whereas GPCP does not assimilate atoll data (Gruber et al., 2000). These differences may introduce some biases in precipitation data over the ocean, but in this study, we also utilize NOAA’s Outgoing Longwave Radiation (NOAA-OLR) to verify these datasets (Liebmann and Smith, 1996). By using three separate datasets that obtain precipitation from different sources, we can better evaluate the robustness of our results.

4.2.2 Models

In addition to a suite of observational datasets, this study uses a fully coupled atmosphere-ocean general circulation model: The Community Earth System Model (CESM1.0) (Shields et al., 2012). Within the CESM, the ocean has a zonal resolution of $3^\circ$ and meridional resolution $1^\circ$ near the equator with 60 ocean layers. The ocean resolution decreases at higher latitudes. The atmospheric component has a T31 horizontal resolution (approximately $3.75^\circ$) with 26 vertical atmospheric levels.

4.2.3 Partial assimilation experiments

To examine precipitation trends in the Indo-Pacific, we first conduct a historical run (HIST) that prescribes time-varying observed natural and external radiative forcings (solar, aerosols, land-use change, and greenhouse gases) from 1850 to 2005 and an RCP4.5 emissions scenario after 2005. The HIST run consists of 10 ensemble members with initial conditions obtained from 10 random years in pre-industrial control simulations. The HIST simulation corresponds to the externally forced component due to radiative forcing. Next, we conduct a global ocean assimilation run (GLOB) by prescribing the same radiative forcing as the HIST run, but assimilate ECMWF ocean reanalysis product (version 4) global 3-dimensional temperature and salinity anomalies into the ocean component of CESM.

To identify trends in precipitation forced by the Indian Ocean and the equatorial Pacific
separately, we conduct two further experiments that assimilate 3-dimensional temperature and salinity anomalies in the Indian Ocean (IO) and equatorial Pacific (eqPAC), separately, in the ocean component of CESM. The assimilation region in the equatorial Pacific is $10^\circ$S-$10^\circ$N and north of $30^\circ$S in the Indian Ocean. The GLOB, eqPAC, and IO experiments run from 1958-2014 with a 2-year spin-up period and consist of 10 ensemble members. Applications of these partial assimilation experiments and how they affect regional climate are described in many previous studies (Chikamoto et al., 2015, 2016, 2020; Ham et al., 2017; Purich et al., 2016).

4.2.4 Trend analysis

We analyze precipitation ratio (p-ratio), defined by the annual precipitation anomaly divided by its climatology (%). A positive (negative) p-ratio trend is synonymous with an increasing (decreasing) annual precipitation trend. Here, we complete linear trends of SST and p-ratio in the Indo-Pacific sector from 1982–2014 in observations and our assimilations. We chose this period to cover the same length of satellite observations for precipitation and SST. Climatologies are defined by the 1982–2014 period, and trends are calculated by annual anomalies.

4.3 Preliminary results

4.3.1 Tropical Pacific precipitation trends

Tropical Pacific precipitation trends feature an equatorial zonal contrast with drying towards the east and a wetting towards the west (Fig. 4.1). The transition zone from positive to negative precipitation trends is about $160^\circ$E. The NOAA-OLR dataset shows the most robust zonal contrast, with consistent trends in CMAP and GPCP with some spatial noise, which may be related to differences in data sources for each dataset (Yin et al., 2004). Fig. 4.2a shows density plots of p-ratio trends in the eastern and western tropical Pacific denoted by the rectangles in Fig. 4.1, further highlighting the zonal contrasts of p-ratio trends, with strong agreement with the GLOB run. The agreement of p-ratio
trends between observation and our GLOB run demonstrates that our assimilation method captures accurate precipitation responses in the tropical Pacific.

Fig. 4.1: Trends in (a) NOAA-OLR, (c,e,g) p-ratio, and (b,d,f,h) SST using various (a-f) reanalysis datasets and (g-h) the GLOB run from 1982–2014. SST is in °C/month.

Trends in Pacific SSTs feature a cooling pattern in the central and eastern Pacific and a warming pattern in the western Pacific in observations and the GLOB simulation, which is suggestive of an IPO phase change (Fig. 4.1). The most robust cooling pattern in observational SST datasets (Figs. 4.1d, f, h) is along the eastern Pacific boundary at 20°-10°S along South America, with another regional cooling trend at 20°-25°N off the coast of Baja California. Additionally, an elongated cooling pattern extends from the central tropical Pacific to the southeast and northeast, connecting these sharp negative SST trends, respectively. These elongated warming bands are in a similar area as the South Pacific Convergence Zone (SPCZ) and a mirroring band in the northern hemisphere. The GLOB
Fig. 4.2: Probability distributions of p-ratio (left) and SST trends (right) over the western tropical Pacific (grey) and eastern tropical Pacific (blue) in observations (top) and the GLOB run (bottom). Observations are from COBE SST and CMAP datasets. The western and eastern tropical Pacific are outlined by boxes in Fig. 4.1.

Simulation captures a similar spatial pattern of trends but a more gradual cooling rate in the central and eastern equatorial Pacific. The COBE, ERSST, and OI datasets are consistent with one another regarding the cooling trend in the central and eastern tropical Pacific, which suggests the phase shift of the IPO is the primary reason for cooling during 1982-2014 in the eastern tropical Pacific. In the western tropical Pacific, the warming rate is similar but opposite to the cooling rates in the eastern tropical Pacific. Observations and GLOB simulations show a zonal contrast of equatorial SST trends lending support to the "global warming hiatus" associated with increased trade wind strength during the negative IPO.
When comparing the spatial patterns of SST and precipitation over the tropical Pacific (Fig. 4.1), a warming (cooling) SST trend is associated with an increasing (decreasing) precipitation trend: highlighting the warmer-get-wetter process. For instance, in the eastern tropical Pacific, the SST cooling trend leads to a stable atmosphere and decreases in precipitation, while the opposite is true for the western tropical Pacific. We further show this relationship through a scatterplot of p-ratio and SST trends in Figure 4.3. In tropical Pacific observations (grey dots on left), we see a distinct positive spatial correlation of trends ($R = 0.55$), suggesting that even during the IPO phase shift, the warmer-get-wetter process is consistent on a local basis (Xie et al., 2010). Fig. 4.3 also shows that the local rate of warming is important for precipitation: If SST warming is greater in one region compared to another, positive precipitation trends will also be greater. Finally, in Fig. 4.4, where we subtracted the eastern and western Pacific SST and p-ratio anomalies, the linear trends (black lines) clearly show that zonal contrast in SST trends is associated with the zonal contrast of p-ratio trends. The slopes between the GLOB and observations are similar in Fig. 4.4, supporting that the warmer-get-wetter scenario is consistent in the tropical Pacific despite a range of climate variability at different time scales.

![Fig. 4.3: Scatter plot of 1982–2014 p-ratio and SST trends at grid points in the Pacific (grey) and Indian Ocean (orange) in observations (left) and the GLOB run (right). Observations are from COBE SST and CMAP datasets.](image)
SST and precipitation trends in the eqPAC run (Fig. 4.5a-b) also show consistency with observations and the GLOB run, further highlighting the warmer-get-wetter hypothesis. Although the SST cooling in Fig. 4.5b is not as strong compared to observations, we clearly see the associated drying trend in the eastern tropical Pacific. These results suggest that during 1982–2014, we had increased precipitation in the western Pacific and decreased precipitation in the eastern Pacific in the eqPAC run, an indication of a strengthened trend in the Walker circulation that is supported by the IPO phase change as seen in SST trends. Our eqPAC runs demonstrate that tropical Pacific precipitation trends are caused primarily by regional tropical Pacific SST forcing.
Fig. 4.5: Trends in (left) p-ratio and (right) SST in the (a,b) eqPAC, (c,d) IO, and (e,f) HIST runs from 1982–2014. SST is in °C/month.

### 4.3.2 Indian Ocean precipitation trends

In the Indian Ocean, the OLR, CMAP, and GPCP datasets all support a similar precipitation trend pattern: a wetting trend over the northern Indian Ocean and a drying trend over the southern Indian Ocean. This meridional contrast in precipitation trends is further depicted in the density functions of observed northern and southern Indian Ocean p-ratio trends (Fig. 4.6). The GLOB run does not show this meridional contrast as clearly on Fig. 4.6c, but we can still see the contrast spatially on Fig. 4.1g. In the NOAA-OLR and the GLOB run, we also see a distinct drying pattern in the southern Indian Ocean. Meanwhile, in the northern Indian Ocean, there is a wetting pattern in the eastern portion of the Arabian Sea, the Bay of Bengal, and absent precipitation trends over India in CMAP and GPCP. However, NOAA-OLR additionally shows positive precipitation trends over India in CMAP and GPCP. The interpolation method over the landmass may be obscuring India precipitation trends as NOAA-OLR only uses satellite estimates compared to station observations in CMAP and GPCP.

Interestingly, the SST pattern does not represent a meridional contrast in trends during 1982–2014 as we might expect in the warmer-get-wetter framework (Fig. 4.1). Instead, we see a monopole SST warming trend. In all three datasets, we see some variations in the
Indian Ocean SST warming rate, with slightly weaker warming rates around India and slightly stronger warming rates in the southern Indian Ocean. The GLOB run shows a slightly stronger warming rate basin-wide compared to observations. By examining the density functions of the northern and southern Indian Ocean, SST trends are positive basin-wide in both observations and the GLOB run, further suggesting a monopole warming trend (Fig. 4.6b and d). One notable difference is that the southern Indian Ocean is warming slightly faster compared to the north in the COBE dataset, whereas SSTs in the GLOB run warm at nearly the same rate in both hemispheres.

Unlike the warmer-get-wetter pattern as seen in the tropical Pacific, this pattern is
observed in observations over the Indian Ocean (orange dots in Fig. 4.3; R = -0.15). The negative correlation suggests that the SST and precipitation trends are unlinked. Interestingly, Fig. 4.7a-b features an interhemispheric precipitation contrast during the 1980s to mid 1990s with a drier northern Indian Ocean. This pattern was reversed from the late 1990s through the 2010s where the southern Indian Ocean was drier. Meanwhile, observed SSTs have an opposite trend in the meridional contrast. In the GLOB run, we see an interhemispheric contrast in precipitation consistent with observations, however, SST trends featured no apparent interhemispheric difference from the earlier to the later period. To that end, there is certainly annual-to-decadal interhemispheric SST variability in the GLOB run, yet still no apparent long-term trend. If SST and precipitation trends were linked with each other in the warmer-get-wetter framework, we would expect similar linear trends in the meridional contrast of SSTs and precipitation like the zonal contrast in the Pacific (Fig. 4.4). But instead, there are opposite contrasts of SST and precipitation trends; therefore, the warmer-get-wetter hypothesis breaks down in the Indian Ocean during 1982–2014. These results suggest another process must be occurring other than regional ocean-atmosphere interactions to describe the interhemispheric contrast of precipitation trends.

To investigate the contribution of remote forcing, regional forcing, and external radiative forcing on precipitation trends over the Indian Ocean, we investigate the IO, eqPAC, and HIST experiments, respectively, in Fig. 4.5. The eqPAC run depicts drying in both hemispheres with wetting over the India landmass (Fig. 4.5a). This result suggests that some of the drying trend observed over the southern Indian Ocean basin is related to remote tropical Pacific forcing (Fig. 4.1). The wetting trend over the India landmass as depicted in both NOAA-OLR, GLOB, and eqPAC p-ratio trends may be related to a negative IPO fostering a more efficient South Asian monsoon (Joshi and Pandey, 2011). Yet, the IO run also features a drying trend over both hemispheres and over the India landmass with monopole SST warming (Fig. 4.5c-d). The combination of the IO and eqPAC runs suggest that vertical motions associated with Pacific forcing combined with regional ocean-atmosphere interactions promote a drying trend in the southern Indian Ocean. However,
the physical process linking ocean-atmosphere interactions in the IO run to this drying trend remains elusive because Indian Ocean SST trends appear to have the opposite effect of the warmer-get-wetter pattern in the southern Indian Ocean. The HIST run (Fig. 4.5d-f) depicts a wetting trend that corresponds to a stronger SST warming rate over the Arabian Sea. Yet, in observations (Fig. 4.1b,d, and f), northern Indian Ocean SSTs tend to have a slower warming rate compared to the rest of the basin. These differences in northern Indian Ocean SST warming rates related to local precipitation trends require further study.
4.4 Conclusions

This study aimed to determine how the precipitation trends in the Indo-Pacific sector respond to a combination of external radiative forcing and internal climate variability in the Pacific and Indian Oceans during the 1982-2014 period. Preliminary observational analyses demonstrate that the warmer-get-wetter hypothesis is consistent in the tropical Pacific basin during 1982–2014, but not in the Indian Ocean. A monopole pattern characterizes SST trends in the Indian Ocean during the same period, but an interhemispheric contrast characterizes precipitation trends. This interhemispheric contrast is consistent among all observed datasets and the GLOB run. However, a question remains as to the physical processes that represent the precipitation trends over the Indian Ocean during 1982-2014. Specifically, (i) does seasonality play a vital role in precipitation trends over the Indian Ocean, and (ii) what variables, instead of SST, dynamically connect with the interhemispheric contrasts in Indian Ocean precipitation trends?

4.5 Proposed work

To answer these questions, we will analyze seasonal trends, given that seasonality plays a significant role in regional climates in and around the Indian Ocean. For instance, winds over the Indian Ocean are strikingly different depending on the season by virtue of the South Asian monsoon, which has implications on regional moisture transport. Therefore, seasonal trends of SST, p-ratio, and winds (moisture advection and moisture convergence) will be further examined using observations and our partial assimilations. Additionally, the land-sea temperature contrast needs to be further examined in observations and our model experiments, given its importance for triggering meridional interhemispheric moisture transport over the Indian Ocean. These results may lead to uncovering the mechanism behind the interhemispheric contrast in precipitation trends.

Because direct SST forcing plays an inconsistent role in precipitation trends over the Indian Ocean, we will examine additional variables, such as vertical velocity, the vertical structure of winds, and SLP. Assessing the contribution of these variables may reveal the impact of low-frequency climate variability on precipitation trends over the Indian Ocean,
such as changes in the Walker circulation from the IPO phase shift. Vecchi and Soden (2007) suggest an anthropogenically forced weakening of the Walker circulation, yet the recent IPO phase change suggests a strengthening of the Walker circulation (England et al., 2014), which may lead to the unusual precipitation trends seen over the Indian Ocean in this study. Using the HIST, eqPAC, and IO runs, we will identify the Indian Ocean precipitation response to the radiative, the tropical Pacific, and the Indian Ocean forcings.
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CHAPTER 5
CONCLUSION

This dissertation explored interannual-to-decadal climate variability through the perspective of ocean-atmosphere and inter-basin interactions, primarily focused on the tropics. Chapter 2 and 3 focused on inter-basin interactions between the Pacific and Atlantic Oceans, while Chapter 4 examined the Indo-Pacific sector precipitation trends. The general findings in this dissertation are that the three tropical oceans are tightly connected through the Walker circulation and that the Atlantic may have more influence on the Pacific than previously thought. This is vital information because of the scientific community’s efforts to understand and predict ENSO, given its socioeconomic impacts and remote teleconnections.

Chapter 2 examined interannual-to-decadal Australian precipitation variability and its relationship to SSTs in the Pacific and Atlantic. Using observational datasets and a suite of model experiments conducted with CESM and MIROC, we found that SST variability is the primary cause for Australian interannual-to-decadal precipitation variability. Specifically, the tropical Pacific is the main driver for Australian precipitation variability, but we also found a significant Atlantic component to Australian precipitation variability. The tropical Atlantic Ocean triggers atmospheric vertical motions and compensating vertical motions in the central tropical Pacific, which results in changes in the global Walker circulation. Once the Atlantic modulates the global Walker circulation, it triggers ocean-atmosphere interactions in the tropical Pacific through local Bjerknes feedback. Subsequently, Australian interannual-to-decadal precipitation changes occur since Australia is on the western flank of the Pacific Walker circulation. Although we did not directly examine the Indian Ocean, a zonal SST dipole pattern emerges in our analysis resembling the IOD and the Ningaloo Niño, which may also influence interannual-to-decadal Australian precipitation variability. The results from Chapter 2 indicate that about half of Australian interannual-to-decadal precipitation variability may have an Atlantic origin. Chapter 2 is published in Johnson
et al. (2018).

Chapter 3 also investigated the Atlantic’s influence on the Pacific through inter-basin interactions but went one step further to assess its impact on PDO variability. The PDO, the leading mode of North Pacific decadal SST variability, arises from regional air-sea interactions and remote forcing from the tropical Pacific. To better understand PDO variability from remote sources, we conducted experiments to assess how much PDO variability originates from the Atlantic and tropical Pacific basins. We found that in addition to the tropical Pacific’s influence on the PDO, the Atlantic can also impact the PDO through two proposed processes. One is the equatorial pathway, similar to the mechanism described in Chapter 2, highlighting changes in the global Walker circulation forced by the Atlantic Ocean. This triggers an ENSO-like response in the tropical Pacific and enhances local convection activity. The atmospheric forcing resulting from changes in tropical Pacific SSTs takes the form of a PNA-like pattern and contributes to the strengthening of Aleutian low variability. The second pathway reveals the AMO impact on the PDO through a Matsuno-Gill type atmospheric response in the Atlantic-Pacific sectors north of the equator. Warm AMO SSTs increase local precipitation and lead to upper-tropospheric high pressure while simultaneously, lower upper-tropospheric pressure and decreases in precipitation emerge over the northwestern subtropical Pacific. These Pacific changes trigger atmospheric Rossby wave propagation toward the north Pacific resembling the Western Pacific (WP) pattern rather than the PNA pattern. These teleconnections linking the tropical and North Pacific cause a thermodynamic ocean response in the North Pacific through changes in Aleutian low variability. Results in Chapter 3 indicate that about 12-20% of PDO variance originates from the Atlantic and 40-44% from the tropical Pacific. Chapter 3 is published in Johnson et al. (2020).

Chapter 4 explored how Indo-Pacific precipitation trends respond to the combination of global SST warming associated with external radiative forcings and La Niña-like tropical Pacific SST cooling associated with internal climate variability during 1982-2014. During this period, precipitation trend patterns are obscured because of the combined SST forcings.
We found a wetting trend in the western tropical Pacific and a drying trend in the eastern tropical Pacific associated with a La Niña-like cooling trend, thereby supporting the warmer-get-wetter hypothesis. However, over the Indian Ocean, we discovered an interhemispheric contrast in precipitation trends with wetting in the north and drying in the south, yet a monopole warming SST trend pattern. This interhemispheric contrast is unclear given the monopole SST warming; therefore, we recommend additional analyses to identify the physical mechanisms representing the precipitation trends over the Indian Ocean. These results illustrate the importance of considering internal climate variability combined with external radiative forcing. Chapter 4 will be submitted to the peer-reviewed journal, *Climate Dynamics*.

Despite deepening the understanding of inter-basin interactions in this dissertation, there are still many unknowns in this research topic as inter-basin interactions appear to be more influential on Earth’s climate than previously thought. One major challenge in identifying inter-basin interactions is that climate modes are intrinsically linked to one another. The interdependency of each ocean basin makes it almost impossible to isolate inter-basin interactions through observations alone, so we must use climate models. However, climate models are compromised by systematic errors, mean-state biases, and model climate sensitivity differences that may obscure inter-basin interactions. Considerable effort in the scientific community has helped reduce some of these errors, but systematic errors still plague the Atlantic. In this dissertation, we attempted to reduce some of these errors through a new assimilation method and strengthen our results using two separate climate models.

In addition to climate model issues needing to be resolved, a large unknown in inter-basin interactions relates to the considerable uncertainty in future climate projections. As seen in Chapter 4, the Indian Ocean is warming at an accelerated rate compared to the Pacific, adding to the unknown of Indo-Pacific inter-basin interactions. It is also unclear how ENSO will evolve in a changing climate. *Cai et al.* (2015) suggest that a slow-down in the Walker circulation is expected due to anthropogenic warming, which could lead to
extreme ENSO events. Will this slow-down in the Walker circulation lead to the Indian Ocean or Atlantic being more influential on the Pacific from the perspective of anthropogenic warming?

Lastly, ENSO is nonlinear, which has resulted in no single ENSO event being the same. Results in this dissertation suggest that the Atlantic Ocean may be more likely to trigger a central Pacific-type ENSO event rather than the eastern Pacific-type, but this may be related to errors simulating the spatial extent of ENSO warming in the Pacific. Although this dissertation gains confidence that Atlantic SSTs can impact the Pacific, the exact branch of the global Walker circulation modulated by Atlantic forcing is unknown, which requires further study. Despite these open questions, advances in understanding inter-basin interactions have provided the impetus for finding answers to the above research topics.
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