Seasonality and Regionality of ENSO Teleconnections and Impacts on North America

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ABSTRACT

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The El Niño – Southern Oscillation (ENSO) has far-reaching impacts across the globe and provides the most reliable source of seasonal to interannual climate prediction over North America. Though numerous studies have discussed the impacts of ENSO teleconnections on North America during boreal winter, it is becoming more and more apparent that the regional impacts of ENSO teleconnections are highly sensitive to the seasonal evolution of ENSO events. Also, the significant impacts of ENSO are not limited to the boreal winter seasons. To address these knowledge gaps, this thesis examines the seasonal dependence of ENSO teleconnections and impacts on North American surface climate, focusing on two examples.

Chapter 1 examines the relationship between El Niño – California winter precipitation. Results show that the probability of the anomalous statewide-wetness increases as El Niño intensity increases. Also, the influences of El Niño on California winter precipitation are statistically significant in late winter (Feb-Apr), but not in early winter even though that is when El Niño usually reaches its peak intensity. Chapter 2 further investigates why the strong 2015/16 El Niño failed to bring above normal winter precipitation to California, focusing on the role of westward shifted equatorial Pacific sea surface temperature anomalies (SSTAs) based on two reasons: the maximum equatorial Pacific SSTAs was located westward during the 2015/16 winter compared to those during the 1982/83 and 1997/98 winters, both of which brought extremely wet late winters to California. Also, the North American Multi-Model Ensemble (NMME) forecasts overestimated the eastern tropical Pacific SSTAs and California precipitation in the 2015/16 late
winter, compared to observations. The Atmospheric General Circulation Model (AGCM) experiments suggested that the SST forecast error in NMME contributed partially to the wet bias in California precipitation forecast in the 2015/16 late winter. However, the atmospheric internal variability could have also played a large role in the dry California winter during the event.

ENSO also exerts significant impacts on agricultural production over the Midwest during boreal summer. Chapter 3 examines the physical processes of the ENSO summer teleconnection, focusing on the summer when a La Niña is either transitioning from an earlier El Niño winter or persisting from an existing La Niña winter. The results demonstrate that the impacts are most significant during the summer when El Niño is transitioning to La Niña compared to that when La Niña is persisting, even though both can loosely be defined as developing La Niña summer. During the transitioning summer, both the decaying El Niño and the developing La Niña induce suppressed deep convection over the tropical Pacific and thereby the corresponding Rossby wave propagations toward North America, resulting in a statistically significant anomalous anticyclone over northeastern North America and, therefore, a robust warming signal over the Midwest. These features are unique to the developing La Niña transitioning from El Niño, but not the persistent La Niña.

In Chapter 4, we further evaluate the performance of NCAR CAM5 forced with historical SSTA in terms of the La Niña summer teleconnections. Though the model ensemble mean well reproduces the features in the preceding El Niño/La Niña winters, the model ensemble mean has very limited skill in simulating the tropical convection and extratropical teleconnections during both the transitioning and persisting summers. The weak responses in the model ensemble mean are attributed to large variability in both the tropical precipitation, especially over the western Pacific, and atmospheric circulation during summer season.
This thesis synthesizes the physical processes and assessments of climate models in different seasons to establish the sensitivity of regional climate to the seasonal dependence of ENSO teleconnections. We demonstrate that the strongest impacts of ENSO on North American regional climate might not be necessarily simultaneous with maximum tropical Pacific SST anomalies. We also emphasize the importance of the multi-year ENSO evolutions when addressing the seasonal impacts on North American summertime climate. The findings in this thesis could benefit the improvement of seasonal hydroclimate forecasting skills in the future.
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Introduction

The El Niño – Southern Oscillation (ENSO) refers to the natural year-to-year variations in the atmosphere and ocean coupled system in the tropical Pacific. It is the most dominant mode of seasonal climate variability (e.g., Shukla and Wallace 1983; Philander 1990; Shukla et al. 2000). Besides the tropical Pacific, ENSO can also lead to large-scale variations of the atmospheric circulation and subsequently affect regional-scale weather and climate around the globe (e.g., Ropelewski and Halpert 1987, 1996), through the so-called “teleconnections” (e.g., Horel and Wallace 1981; Trenberth et al. 1998; and many others). Their substantial effects on surface temperature and precipitation can persist over several seasons, leading to a wide range of impacts and natural disasters (e.g., flooding, landslides, droughts, and wild fires), crop failures, or disease outbreak, resulting in tremendous socioeconomic losses across the globe.

On the other hand, ENSO also provides the most reliable source for seasonal climate forecast skill in many regions, including North America (e.g., Tippett et al. 2012; L’Heureux et al. 2015). Tremendous progress has been made in the past three decades in our understanding of ENSO dynamics and its remote impacts on the seasonal climate over North America, especially during the boreal winter season. However, it is becoming more and more apparent that the regional impacts of ENSO teleconnections are highly sensitive to the seasonal evolution of ENSO events. Also, the significant impacts of ENSO are strongest in, but not limited to, the boreal winter seasons (e.g., Wang et al. 2007; Ding et al. 2011). Thus, the overarching goal of this thesis is to examine the seasonal dependence of ENSO teleconnections and impacts on the surface climate over North America. In this introduction, how ENSO affects remote regions via teleconnections will first be briefly discussed. Then, the importance to detail the seasonal evolutions of ENSO teleconnections
and impacts on North America will be demonstrated. The gaps in our understanding based on the previous literatures will also be discussed.

**What are ENSO teleconnections?**

In the tropical Pacific, the atmosphere and ocean are tightly coupled to each other. ENSO events periodically warm and cool the central to eastern equatorial Pacific sea surface temperature (SST), due to the relaxation and strengthening of the easterly trades, which lead to anomalous convective activity across the tropical Pacific region (Figure i.1). The anomalous tropical deep convection drives the bridge that connects the tropical SST anomalies (SSTA) with extratropical climate. During an El Niño, enhanced deep convection in the central to eastern Pacific triggered by the warm SST, induces anomalous upper tropospheric divergence (Figure i.1b). The anomalous stretching and outflow generates a quasi-stationary Rossby wave-train that propagates from the tropics across the Pacific-North America (PNA) region, to the extratropics approximately along a great circle route on the sphere but retracted by the mean flow (e.g., Hoskins and Karoly 1981; Webster 1981). This wave-train often leads to a deepened Aleutian Low which shifts the subtropical jet-stream and storm track equatorward, altering the weather patterns in the extratropical region (Figure i.2; e.g., Trenberth et al. 1998). On the contrary, during a La Niña, anomalously suppressed deep convection over the central to eastern tropical Pacific caused by the cold SST (Figure i.1c), shifts the subtropical jet-stream and storm track further poleward.

The modified storm track and the mid-latitude transient eddy activity can further feedback to change the mid-latitude mean-flow, which is impacted by the direct diabatic heating due to ENSO SST (Hoerling and Ting 1994). Therefore, besides the direct tropical influence via Rossby wave dispersion, midlatitude transient eddies also play an important role in maintaining
Figure i.1 Schematic diagrams of (a) normal, (b) El Niño, and (c) La Niña conditions in the tropical Pacific. Figures from the Australian Bureau of Meteorology [available: http://www.bom.gov.au/climate/enso/history/Ln-2010-12/three-phases-of-ENSO.shtml].
and modulating the extratropical response to the ENSO tropical forcing through an eddy-mean flow positive feedback (e.g., Harnik et al. 2010; Seager et al. 2010).

Both changes in the large-scale circulation (anomalous highs and lows) and the path of weather systems exert impacts on the surface climate, in both temperature and precipitation, across the globe, and notably over North America. These remote impacts of ENSO are the primary basis of seasonal climate prediction.

**Figure i.2** Schematic diagram of the dominant change in the upper troposphere in response to El Niño warm SSTA and the associated enhanced deep convection as well as anomalous upper tropospheric divergence near the equator (scalloped region). In the presence of mean westerlies, the anomalous divergent outflow results in an anomalous anticyclone in the subtropics, triggering a quasi-stationary Rossby wave-train, composed of alternating high and low geopotential anomalies, and propagating across the PNA region along a great circle route on the sphere. Figure from Trenberth et al. 1998 (Figure 4).
Seasonality of ENSO life-cycle and teleconnections

The evolution of the tropical ENSO SSTA has strong seasonality. Coupled with the seasonal cycle of mean-flow, the ENSO teleconnections and their regional impacts also have strong seasonality. A typical ENSO event develops in the late boreal spring, peaks at the end of the calendar year (denoted as El Niño(0) or La Niña(0) in Figure i.3), and decays in the following spring to early summer (Figure i.3; e.g., Rasmusson and Carpenter 1982; Okumura and Deser 2010). While El Niño and La Niña act like a mirror image of each other in many features, they do exhibit strong asymmetry in their temporal evolutions. After the mature phase of El Niño, the SSTA in the tropical Pacific tends to decay rapidly and often transition into the cold phase in the following spring to early summer (Figure i.3a). On the other hand, La Niña tends to persist through the following year and re-intensify in the subsequent winter, often becoming a multi-year La Niña (Figure i.3b; e.g., McPhaden and Zhang 2009; Okumura and Deser 2010; Okumura et al. 2017).

Besides the tropics, the extratropical response to ENSO also exhibits strong seasonality. The strength of the extratropical teleconnections depends not only on the amplitude of tropical SSTA, but also the mean-state, especially the subtropical jet-stream which tropical convection perturbs (e.g., Hoskins and Karoly 1981; Webster and Holton 1982). ENSO tropical forcing affects the extratropics through Rossby wave dispersion and mid-latitude eddy-mean flow feedback, both tightly linked to the intensity and location of the subtropical jet-stream (e.g., Hoskins and Ambrizzi 1993; Hoerling and Ting 1994). Thus, the teleconnections and their impacts on extratropical North America are strongest in the boreal winter when the ENSO tropical forcing reaches its peak and the jet-stream is strong and shifts closer to the tropics, allowing the Rossby wave source originating from tropical diabatic heating anomalies to extend into westerly flows and, hence, allowing Rossby wave propagation into the mid-latitudes (e.g., Webster 1982).
Figure i.3 Four-year evolutions of Oceanic Niño Index (three-month average SSTA in the Niño 3.4 region) for all the (a) El Niño and (b) La Niña events from 1950 to 2017. The calendar years corresponding to the ENSO events are denoted as year 0 (El Niño(0) & La Niña(0)). The calendar years prior to and following the events are denoted as year -1 and year 1, respectively. For multi-year ENSO events (e.g. 1998-2000 three-year La Niña), only the first ENSO events (i.e. 1998) are denoted as year 0. Grey lines are for each individual ENSO event. Thick lines indicate the average across all the El Niño and La Niña events. All the year 0 events are listed above each panel. The criteria of El Niño and La Niña events are described in Chapter 3.2.2.
Although the maximum SSTAs in the tropical Pacific and the associated teleconnections are strongest in the boreal winter, some features demonstrate a seasonal evolution of teleconnections throughout the winter half-year. For example, the strongest anomalous tropical circulation and thereby the teleconnections across the PNA region actually lag the maximum SSTA in the tropical Pacific about one to three months (Kumar and Hoerling 2003). The delayed response of atmospheric circulation is due to the dependence of deep convection on the total SST rather than SSTA. As the climatological SSTs over the central to eastern Pacific are warmer in the late boreal winter (February-April), the total SST and the associated anomalous convection are stronger after the peak of tropical SSTA which usually happens in November to January. Most previous studies have focused on the entire winter season (November-January, December-February, or even the entire winter half-year), thus ignoring the details of the sub-seasonally varying ENSO teleconnections. Yet, the regional responses to ENSO forcing can be highly sensitive to the subtle differences in the teleconnection patterns. Any slight difference in the configuration and location of the wave-train could differentiate between a dry and wet winter in a region (e.g., Hoerling and Kumar 1997; Chen and Kumar 2018). The relationship between California winter precipitation and El Niño is one of the examples. As the source of California winter precipitation can come from just a few heavy events due to extratropical cyclones (e.g., Dettinger et al. 2011), precise forecast of the timing and location of anomalous precipitation is critical to the water management in California. Therefore, improved knowledge of the seasonal evolution and regional details of the extratropical teleconnections in response to tropical SST forcing throughout the winter season is essential to advance our understanding of the regional impacts of ENSO. This regional and seasonal dependence of ENSO teleconnection over California will be discussed in Chapter 1.
Another feature of the seasonal dependence of teleconnections is that the anomalous tropical 200hPa geopotential height and tropospheric warming linger into the following summer even after the tropical Pacific El Niño SSTA dissipates. The primary source for this atmospheric thermal inertia is the delayed evolution of warm SSTA in the tropical Indian Ocean (e.g., Kumar and Hoerling 2003; Lau et al. 2005; Xie et al. 2009), caused by the El Niño tropical Pacific SSTA via the so-called atmospheric bridge (e.g., Alexander et al. 2002; Xie et al. 2009). The SST warming in the tropical Indian Ocean also induces anomalous atmospheric circulation, influencing remote area in the boreal spring to summer. Hence, the impacts of El Niño tropical forcing in the extratropics can persist into the following summer even after the dissipation of tropical Pacific SSTA. This delayed response to the ENSO tropical SST forcing also implies the different ENSO impacts in La Niña summers that follow an El Niño or are within a persisting La Niña, arguing that it is imperative to detail the seasonal evolution of ENSO teleconnections.

On the other hand, the ENSO teleconnections in boreal summer are less-established, compared to the ones in boreal winter. The primary reason is that, at the source, the intensity of teleconnections is weaker as the anomalous tropical SST and deep convection are in either the developing or decaying phases. Also, the dominance of tropical easterlies and the weaker and poleward-shifted North Pacific climatological jet stream limit the potential for Rossby wave propagation away from the tropics into the extratropical region (e.g., Hoskins and Karoly 1981; Webster and Holton 1982). Despite these unfavorable conditions, previous studies have demonstrated the possibility that ENSO tropical forcing can trigger Rossby waves propagating toward higher latitudes in the summer that impact North American summer climate (e.g., Lau and Peng 1992; Ding et al. 2011; Douville et al. 2011). For example, one previous study suggested that developing La Niña summers induce dry and hot conditions that depress US maize and soybean
yields (Anderson et al. 2017). Notwithstanding the strong impacts on agricultural production, the physical process underlying the ENSO summer teleconnections has not been established.

The above discussions establish the need to further examine seasonally evolving ENSO teleconnections and their regional impacts, both throughout the winter half-year and in other seasons, which can not only fill the gaps in our understanding of the complexity of ENSO teleconnections, but also improve seasonal climate prediction on regional scales. Chapters 3 and 4 will further discuss these issues regarding La Niña summer teleconnections and their sensitivity to the temporal evolution of the La Niña events in North America.

**The challenge of seasonal forecasting the regional impacts of ENSO teleconnections**

As discussed above, the ultimate goal of better understanding the seasonality of ENSO teleconnections is to improve the model forecasting skill of North American seasonal hydroclimate. However, strong event-to-event variability in ENSO regional impacts imposes appreciable challenges to operational seasonal forecast (e.g., Hoerling and Kumar 1997). The unexpectedly dry California winter during the strong 2015/16 El Niño is the most recent example. Most of the operational seasonal forecasts, including the North American Multi-Model Ensemble (NMME), predicted wetter-than-normal conditions in California in the 2015/16 late winter, consistent with the empirical relationship between El Niño and California (e.g., Wanders et al. 2017). The forecast failure demonstrates that it is necessary to identify the sources that causes the event-to-event variability (e.g., Siler et al. 2017; Chen and Kumar 2018; Quan et al. 2018; Singh et al. 2018).

There are several possible factors that can cause variations in North American climate between ENSO events. These include (i) random atmospheric internal variability unrelated to the forcing, (ii) sensitivity to differences in the detailed structure and longitudinal location of the SSTA
(e.g., Hoerling and Kumar 1997; Guo et al. 2017; Chen and Kumar 2018), and (iii) the influence of other oceans. Atmospheric internal variability is the dominant source of variability in the extratropical climate on seasonal time-scale (e.g., Hoerling and Kumar 1997; Deser et al. 2018). Hence, atmospheric internal variability can lead to large event-to-event variations in extratropical seasonal climate even when the tropical SSTA are similar among these ENSO events. On the other hand, the different longitudinal locations of the SSTA can cause shifts in the forced stationary waves, leading to different teleconnection patterns and impacts on North American climate (e.g., Mo and Higgins 1998a,b; Barsugli and Sardeshmukh 2002; Hoerling and Kumar 2002). Other possible sources of event-to-event variations include the variability in the Indian Ocean (e.g., Siler et al. 2017) and Arctic sea-ice (e.g., Singh et al. 2018). Also, the possibility that ENSO teleconnections are being modified by the warming climate cannot be ignored (e.g., Yeh et al. 2009; Cai et al. 2015; Quan et al. 2018).

Although it is difficult to quantify the relative contributions of each factor that caused the observed variability and the inaccurate seasonal forecast in 2015/16, it is important to acknowledge that, even with ENSO as the most dominant mode of seasonal climate variability, other atmospheric factors could lead to an appreciable range of event-to-event variations (e.g., Chen and Kumar 2018; Deser et al. 2018). Given the fact that the observations are just one single realization among many possible outcomes under the influence of internal atmospheric variability, it is essential to assess the performance of the entire ensemble of climate model forecasts, rather than simply the ensemble mean skill. Chapter 2 takes a close look at the prediction of the 2015/16 El Niño in this way and considers what we can learn from this important example.
The objective of this thesis

The overarching objective of this thesis is to address the gaps in our understanding regarding regional and seasonal dependence of ENSO teleconnections, as a necessary first step to better seasonal prediction skills. To accomplish this goal, this thesis focuses on two examples: the impacts of El Niño on California winter precipitation and the impacts of a developing La Niña on Midwest summer climate, through addressing the following specific questions:

The impacts of El Niño on California winter precipitation

- How do the impacts of El Niño on California precipitation change throughout the winter season?
- What is the role of tropical SSTA intensity in the relationship between El Niño and California winter precipitation?
- Why did California not receive the excess precipitation expected and predicted during the strong 2015/16 El Niño?
- What are the possible reasons that led to the inaccurate forecast for California precipitation during the El Niño 2015/16 winter? Did the forecast models really fail?

The impacts of a developing La Niña on Midwest summer climate

- What are the physical processes underlying different La Niña summer teleconnections?
- What are the possible physical mechanisms that lead to the significant drop in crop yields over the Midwest during the El Niño to La Niña transitioning summers?
- How well can an AGCM forced with historical SSTs reproduce the La Niña summer teleconnections?

The layout of this thesis are summarized in Table i.1.
**Table i.1** Thesis objectives and analysis structure.

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<th>Tasks</th>
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<tr>
<td>Chapter 1</td>
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<tr>
<td>• How do the impacts of El Niño on California precipitation change throughout the winter season?</td>
<td>El Niño’s Impacts on California winter precipitation: seasonality, regionality, and El Niño intensity (based on 1900-2010 historical events in observations)</td>
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<td>• What is the role of tropical SSTA intensity in the relationship between El Niño and California winter precipitation?</td>
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<tr>
<td>Chapter 2</td>
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<td>• Why did California not receive the excess precipitation expected and predicted during the strong 2015/16 El Niño?</td>
<td>The role of equatorial Pacific SST anomalies in the late winter California precipitation during the 2015/16 El Niño (using observations, NMME forecast, and AGCM experiments)</td>
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<tr>
<td>• What are the possible reasons that led to the inaccurate forecast for California precipitation during the El Niño 2015/16 winter? Did the forecast models really fail?</td>
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<td><strong>The impacts of a developing La Niña on Midwest summer climate</strong></td>
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<td>Chapter 3</td>
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<td>• What are the physical processes underlying different La Niña summer teleconnections?</td>
<td>ENSO teleconnections and impacts on US summertime temperature during multi-year La Niña life-cycle (using observations and a stationary wave model)</td>
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<td>• What are the possible physical mechanisms that lead to the significant drop in crop yields over the Midwest during the El Niño to La Niña transitioning summers?</td>
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<td>Chapter 4</td>
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<tr>
<td>• How well can an AGCM forced with historical SSTs reproduce the La Niña summer teleconnections?</td>
<td>ENSO summer teleconnections and impacts on North America as simulated in NCAR CAM5-GOGA</td>
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Chapter 1. El Niño’s Impact on California Precipitation: Seasonality, Regionality, and El Niño Intensity

Note: This chapter has been published in Environmental Research Letters (Jong et al. 2016)

1.1 Introduction

As California battles severe drought, it becomes increasingly important to understand the atmospheric and oceanic conditions that could interrupt or even end the drought that began in 2011. Recent research (Seager et al. 2015; Hartmann 2015; Davies 2015; Watson et al. 2016) indicates that the current drought is associated, to a significant degree, with warm sea surface temperature anomalies (SSTA) in the western tropical Pacific Ocean and, some argue, cool SSTA in the central to eastern equatorial Pacific. As the SSTA pattern evolved during the 2015/2016 El Niño event, an important question emerged as to the likelihood of El Niño moderating drought conditions. After all, El Niño, the most significant mode of climate variability and is a reliable source of seasonal to interannual prediction, imposes a major control on western North America climate (e.g., Schonher and Nicholson 1989; Cayan and Redmond 1994; Mo and Higgins 1998a; Andrews et al. 2004; Schubert et al. 2008; Seager and Hoerling 2014). Since California is one of the largest economies in the world and a world leader in agricultural production, improved understanding of El Niño’s impact on California precipitation has great economic and societal value.

During an El Niño, a low-frequency Rossby wave-train, forced by the positive SSTA and enhanced deep convection in the tropical Pacific, propagates from the equator to extratropical regions over the North Pacific and North America (e.g., Rasmusson and Wallace 1983; Trenberth et al. 1998), influencing climate in remote regions via well-known “teleconnections”. The large-
scale anomalous atmospheric circulation patterns impact the weather across all of North America, leading to wetness in the southeastern U.S. during El Niño winters (e.g., Ropelewski and Halpert 1986, 1996; Livezey et al. 1997; Mason and Goddard 2001; Chiodi and Harrison 2013) as well as a dry north-wet south pattern across western North America (e.g., Livezey et al. 1997; Dettinger et al. 1998). However, the influence of El Niño conditions on western North American rainfall, particularly California, is not robust and shows substantial variability among different historical events (Yarnal and Diaz 1986).

Two of the most prominent mid-latitude responses to El Niños are the deepening of the Aleutian Low (e.g., Bjerknes 1969; Schonher and Nicholson 1989) and the strengthening and southward shift of the subtropical jet over the North Pacific (Trenberth et al. 1998). Both conditions could lead to a southward shifted storm track and cause anomalous wetness at the U.S. southwest coast (Seager et al. 2010). On a regional scale, however, the precise location and timing of the precipitation increase can be extremely sensitive to the longitudinal and latitudinal position and strength of the low pressure anomaly, which determines how circulation anomalies may steer storms towards the U.S. west coast (Yarnal and Diaz 1986). For example, in California, if the low pressure anomaly is located right off the west coast, the entire state tends to be wetter during an El Niño winter; while if the low anomaly is located further to the west or north, precipitation may actually reduce in California (e.g., Schonher and Nicholson 1989; Ely et al. 1994).

Furthermore, the impact of El Niño on California rainfall varies from north to south (Schonher and Nicholson 1989; Cayan et al. 1999; Andrews et al. 2004; Schubert et al. 2008). About two-thirds of the total winter precipitation in California falls on the windward side of the northern California Sierra Nevada mountain range (Dettinger et al. 2011), but it is southern California, the relatively dry part of the state, that has the strongest relationship with El Niño.
Though the above studies pointed out the differences of El Niño’s impact in northern and southern California, most of them focused on the relationships by using winter half year or annual data. However, the change in El Niño impacts throughout the winter season and the role of El Niño intensity have not been fully examined and well-documented. In this chapter, we focus on quantifying El Niño’s impact on California precipitation based on historical observations of precipitation, SST, and atmospheric circulations. The goal is to determine the dependence of the El Niño - California precipitation relationship on timing (early versus late winter), region (northern versus southern California), and the strength of the El Niño SSTA. Such information will be of use in seasonal forecasting for California, including the alleviation and/or termination of drought conditions.

1.2 Data and Method

In this study, the relationships between SSTA in the Niño 3.4 region (120°W-170°W, 5°S-5°N) and northern and southern California winter precipitation (November - April) from 1900/01 to 2009/10 are examined. Sea surface temperature data are taken from the Extended Reconstructed Sea Surface Temperature (ERSST) version 3b from the National Oceanographic and Atmospheric Administration (NOAA) (Smith et al. 2008). ERSST provides monthly SST data from 1895 with 2°×2° spatial resolution. Here, the trend from 1900/01 to 2009/10 of Niño3.4 SSTA (0.065°C/10yr for NDJ; 0.053°C/10yr for FMA) is removed to isolate the interannually varying component. To be consistent, all the variables used in this research have been linearly de-trended and the trend is removed for each three-month season. The California precipitation data are taken from the Global Precipitation Climatology Centre (GPCC, Full Data Product version6) that provides monthly gridded precipitation from 1901 to 2010 with 0.5°×0.5° spatial resolution (Schneider et al. 2014).
The state of California is divided into northern and southern parts based on the characteristics of the climatological winter precipitation (Figure 1.1a) as well as the correlations between precipitation and de-trended Niño3.4 SSTA (Figure 1.1b). Here, northern California is defined as the region within 124°W-118°W, 36°N-42°N, while southern California is within 122°W-114°W, 32°N-36°N (see the two black boxes in Figure 1.1a and b).

The atmospheric circulation data for 200hPa geopotential height are taken from the NOAA 20th Century reanalysis version 2c (Compo et al. 2011). The monthly data are available from 1851 to 2014 with 2° × 2° spatial resolution. To understand the El Niño teleconnection, global precipitation, near-surface moist static energy (MSE) and convective available potential energy (CAPE) from 1979/80 to 2009/10 are also used in this research. Monthly global precipitation data are obtained from the Global Precipitation Climatology Project (GPCP, version 2) (Adler et al. 2003), from January 1979 to the present with a spatial resolution of 2.5° by 2.5°. Monthly MSE and CAPE data are derived from European Center for Medium-Range Weather Forecasts (ECMWF) interim reanalysis dataset (ERA-Interim) from January 1979 to November 2015 with a spatial resolution of 1.5° by 1.5° (Dee et al. 2011). The identification of El Niño and La Niña years is based on the “Oceanic Niño Index (ONI)”, derived from the 3-month running mean of Niño3.4 SSTA, relative to a centered 30-year climatology updated every 5 years. El Niño events are defined when the ONI reaches the threshold of +0.5°C for at least 5 consecutive 3 month means (see the NOAA Climate Prediction Center website: http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml for a complete description). In this study, the definition of weak and moderate-to-strong El Niños is based on the strength of the de-trended Niño3.4 SSTA. Weak (moderate-to-strong) El Niños are defined as the Niño3.4 SSTA between 0.5°C and 1°C (more than 1°C) for the corresponding season.
Figure 1.1 (a) California region climatological precipitation for the six-month cold season (NDJFMA, 1900/01-2009/10) in units of mm/month. The boxes indicate the areas of northern and southern California used in this research. (b) Correlation coefficients between precipitation and de-trended Niño3.4 SSTA (NDJFMA, 1900/01 – 2009/10). Stippling denotes 95% significance. (c) Monthly climatology of precipitation averaged for the California region (blue solid line), northern California (green dashed line), and southern California (orange dotted line) in units of mm/month. (d) Correlation coefficients between 3-month running mean de-trended Niño3.4 SSTA and 3-month running mean precipitation in California region (blue solid), northern (green dashed), and southern (orange dotted) California.
1.3 Results

The seasonal cycles of the monthly precipitation over the entire state and the northern and southern parts of California are shown in Figure 1.1c. There is a clear peak in total precipitation in winter months, from December to February. Northern California receives more than double the amount of precipitation of southern California. Three-month running mean correlations between California precipitation and the de-trended Niño3.4 SSTA are shown in Figure 1.1d. There is an almost linear increase in correlation from October, November and December (OND) to February, March, and April (FMA) for all three regions. The precipitation in northern California is only weakly influenced by ENSO (El Niño – Southern Oscillation, including El Niño and La Niña), as shown in Figure 1.1d, possibly due to large internal atmospheric variability in northern California caused by mid-latitude weather systems. Although the west portion of northern California shows a coherent region of significant correlation with ENSO, the relation is not as strong as in southern California, where the correlation is largely significant throughout the winter season.

To examine further the relationship between California winter precipitation and El Niño on a year-to-year basis, as well as the difference in the relationship for northern and southern California during early and late winters, Figure 1.2 shows scatter plots for precipitation anomalies in percent of climatology in northern and southern California as a function of the Niño3.4 SSTA. The percentages, instead of the absolute values, are used here because the distribution of precipitation is extremely uneven across the state, as shown in Figure 1.1a. The percent of climatology is defined here based on the area-averages of precipitation anomaly and climatology. We also calculated the precipitation percent anomaly at each grid point first and then area-averaged the percentages over northern and southern California. The results in Figure 1.2 are insensitive to how the percent of climatology is calculated. The colors in Figure 1.2 indicate El Niño (red), La
Niña (blue) and neutral (black) years according to the NOAA definition as explained in section 1.2. In early winter (NDJ), the Niño3.4-precipitation relationships are very weak for both northern and southern California, although the strongest El Niño events tend to have above normal precipitation in both regions. In contrast, in late winter (FMA), the relationships strengthen in both northern and southern California. While the correlation in northern California is weak (0.19), it is significant at the 5% level and, furthermore, precipitation anomalies are all above normal for the five most intense El Niño events. In southern California, the correlation is highly significant (0.43), with the strong El Niño events having between 80 and 160 percent above climatological normal precipitation. Figure 1.2 also indicates the asymmetry of ENSO’s impact on precipitation for northern and southern California. In late winter, the correlation between precipitation anomalies and Niño3.4 SSTA for El Niño events only (dashed lines in Figure 1.2) is 0.50 (p=0.0025) in northern California and 0.53 (p=0.0014) in southern California, both are highly significant. Compared to the correlations for all years (shown in Figure 1.2), northern California has a high tendency to be wet during an El Niño, but not necessarily dry during La Niña. The asymmetry also exists in southern California, although to a lesser extent. The lack of significant impact of the La Niña events on California precipitation may be related to the fact that suppressed convection and the associated atmospheric teleconnection patterns tend to be located further to the west, and away from the North American west coast, for La Niña compared to El Niño (Hoerling et al. 1997).

To understand the difference between El Niño’s impacts on early and late winter California precipitation, Figure 1.3 shows the composites of 200 hPa geopotential height anomalies (contours) and SSTA (shading) for moderate-to-strong El Niños and weak El Niños for early and late winter. The corresponding California precipitation anomalies in percent of climatology are also shown. In all four cases, there is a low-pressure anomaly over the northern North Pacific, a high anomaly
over Canada and another low over the Southeastern U.S., consistent with the well-known Pacific North American (PNA) teleconnection pattern (Horel and Wallace 1981). In the moderate-to-strong El Niño composites, however, the intensity of the PNA pattern increases in late winter even though the tropical Pacific SSTA decreases slightly (top panels in Figure 1.3). The differences of 200hPa height in late and early winter are statistically significant at 95% confident interval over the eastern North Pacific and the U.S. west coast (figure not shown). The anomalous Aleutian Low for late winter is almost double the amplitude of that for early winter while the Niño3.4 SSTA decreased from 1.52°C to 1.29°C (Table 1.1). Correspondingly, the precipitation anomaly in northern (southern) California is more than 10 (8) times larger in late winter than in early winter (see Table 1.1). For the weak events (lower panels), however, the seasonal dependence is less striking although late winter does show lower heights over California and a stronger precipitation response (Table 1.1).

Table 1.1 Average de-trended Niño3.4 SSTA and northern and southern California precipitation anomalies (in % of climatology) for (upper) moderate-to-strong and (lower) weak El Niños in early and late winters during 1900/01 to 2009/10. The anomalies for moderate-to-strong El Niños during 1979/80 to 2009/10 are shown in parentheses. Italic (asterisk) numbers indicate the 90% (95%) significance of variations.

<table>
<thead>
<tr>
<th>Niño3.4 SSTA (°C)</th>
<th>CA precip anomalies (%)</th>
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<tbody>
<tr>
<td></td>
<td>NDJ</td>
</tr>
<tr>
<td>NDJ</td>
<td>FMA</td>
</tr>
<tr>
<td>Moderate-Strong</td>
<td></td>
</tr>
<tr>
<td>1.52° (1.57°)</td>
<td>1.29° (1.39°)</td>
</tr>
<tr>
<td>Weak</td>
<td>0.79°</td>
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<td></td>
<td>N</td>
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<tr>
<td></td>
<td>S</td>
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<td></td>
<td>5.05%</td>
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</table>

Italic (asterisk) numbers indicate the 90% (95%) significance of variations.
Figure 1.2 De-trended precipitation anomaly (% of climatology) as a function of de-trended Niño3.4 SSTA for (upper) northern and (bottom) southern California in (left) early (NDJ) and (right) late (FMA) winter from 1900/01 to 2009/10. Red, black, and blue dots denote El Niño, neutral, and La Niña years, respectively. The two strongest El Niño events on record (1982/83 and 1997/98) are indicated as purple and yellow dots, respectively. Dashed lines are the best fits for El Niño events only.
Figure 1.3 Composites of (a, c, e and g) 200hPa height anomalies (contour, interval: 10m), SSTA (ocean) for and (b, d, f, h) precipitation anomalies (% of climatology) for (top) moderate-to-strong and (bottom) weak El Niños in (left) early and (right) late winters during 1900/01 to 2009/10. All the anomalies here are de-trended. The plotting region for b, d, f, h is within 124°W-114°W, 32°N-41°N, indicated by the boxes in a, c, e, g. Stippling regions in a, c, e, g indicate the 95% significance for 200hPa height variations. Stippling regions in b, d, f, h indicate the 95% significance for precipitation variations. The definition of weak and moderate-to-strong El Niños is described in Section 1.2. The numbers of El Niño events for each case are indicated in the titles.
The possible cause of the apparent nonlinear relationship between SSTA amplitude and teleconnection strength in Figure 1.3 for early and late winter may be the warmer equatorial Pacific SST basic state in late winter. That is, even though, during an El Niño event, the SSTA weakens from early to late winter, the smaller anomalies in late winter are imposed on a warmer climatological SST, which leads to a more favorable environment for deep convection. To investigate this, Figure 1.4 shows the composites of precipitation anomalies (shaded) and total SST (contours) in early and late winters for the moderate-to-strong El Niño events that have occurred since satellite observations became available. The SSTA composites for this period are similar to those in Figure 1.3 (top panels). While, the Niño3.4 SSTA amplitude is 1.57°C for early winter and 1.39°C for late winter (Table 1.1), the composites of precipitation anomalies show a much stronger late winter signal that also extends further to the east of the corresponding location in early winter. The total SST composites show that, in late winter, the warmest region (indicated by the 28°C isotherm, thick lines in Figure 1.4) extends further east, which enhances the deep convection and precipitation anomalies in the central and eastern Pacific.

To further examine the characteristics of the environment for deep convection, Figure 1.5 shows the latitudinally averaged 1000hPa MSE, CAPE and precipitation between 10°N and 15°S for early and late winters during moderate-to-strong El Niños. The near-surface entropy (MSE) and CAPE are measures of the strength of instability that deep convection removes. In the central to eastern Pacific (around 150°W to 80°W), near-surface MSE and CAPE (Figure 1.5a,b) are larger in late winter than in early winter, indicating a more unstable environment favorable to deep convection, consistent with the larger precipitation composites in late winter (Figure 1.5c). The results in Figure 1.4 and Figure 1.5 support the hypothesis that stronger teleconnection patterns in late winter occur due to stronger and more eastward-spread tropical heating anomalies. This is a
consequence of warmer climatological SST conditions in the eastern tropical Pacific in late winter than early winter allowing a smaller SSTA to cross the threshold for deep convection. Further idealized modeling experiments are needed to fully understand the differences in early and late winter teleconnections, the dynamical mechanisms behind it, and the possible additional contribution from the tropical-wide warming that follows the peaking of the El Niño.

Figure 1.4 Composites of GPCP precipitation anomalies (shaded, unit: mm/day) and SST (contour, interval: 1°C) for moderate-to-strong El Niño in (a) early and (b) late winters during 1979/80 to 2009/10. Thick contours indicate SST 28°C isotherm. Stippling regions indicate 95% significance for precipitation variations. All the variables used here are de-trended.
1.4 Discussions and Conclusions

The seasonality, regionality and dependence on El Niño intensity of California rainfall anomalies have important implications for seasonal prediction of El Niño’s impacts. Dividing the climatological precipitation distribution into terciles, Table 1.2 shows the number of moderate-to-strong and weak El Niño events that fell into each tercile, for northern and southern California and for early and late winter, as well as the associated precipitation anomaly, expressed as percent of climatology for each tercile. This provides overall information about the relationship between El Niño and California precipitation. During early winter, in both northern and southern California, there is no clear preference for El Niños to be in the wettest tercile. However, in late winter, 8 of 10 moderate-to-strong El Niño events put southern California in the wettest tercile. For northern California, none of these 10 events put northern California in the driest tercile, 6 were in the middle

**Figure 1.5** Composites of latitudinal average (a) 1000hPa moist static energy, (b) convective available potential energy (CAPE), and (c) GPCP precipitation between 10°N and 15°S for moderate-to-strong El Niño in early (dashed lines) and late (solid lines) winters during 1979/80 to 2009/10.
tercile and 4 in the upper tercile. In other words, with regard to season, El Niño’s impacts are likely to be stronger in late winter than in early winter; and, in terms of region, southern California has a greater chance of wet winters during an El Niño than northern California. Further, only a relatively strong El Niño is likely to bring heavy precipitation across the entire state. In summary, a moderate-to-strong El Niño in the late winter can make southern California precipitation very likely to be in the upper tercile and make northern California precipitation very unlikely to be in the lower tercile, while a weak El Niño or a moderate-to-strong El Niño in early winter cannot be relied upon to favor a wet winter in California.

**Table 1.2** Precipitation anomaly (% of climatology) of each precipitation tercile in (top) northern and (bottom) southern California during moderate-to-strong and weak El Niños in (left) early and (right) late winters. The number of events for each category is shown in parentheses. (For instance, for the past 10 moderate-to-strong El Niño during late winter, 1 had below normal precipitation southern California; 1 had a normal precipitation; 8 were associated with above normal precipitation. These 8 winters had precipitation 107% above climatology on average.) Italic (asterisk) numbers indicate statistically significant at 90% (95%) confidence using Monte Carlo bootstrapping method.

<table>
<thead>
<tr>
<th></th>
<th>NDJ</th>
<th>FMA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lower</td>
<td>Middle</td>
</tr>
<tr>
<td><strong>North</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Strong El Niño</td>
<td>43% (5)</td>
<td>2% (9)</td>
</tr>
<tr>
<td>Weak El Niño</td>
<td>46% (5)</td>
<td>0% (3)</td>
</tr>
<tr>
<td><strong>South</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Strong El Niño</td>
<td>42% (5)</td>
<td>10% (7)</td>
</tr>
<tr>
<td>Weak El Niño</td>
<td>48% (6)</td>
<td>9% (2)</td>
</tr>
</tbody>
</table>
El Niño’s impact on California late winter precipitation is associated with the strengthening of the teleconnection from early to late winter even though the tropical Pacific SSTA decreases slightly. This nonlinearity between SSTA and teleconnection response is possibly caused by a stronger and more eastward extended tropical diabatic heating in late winter due to a warmer climatological mean SST over the tropical eastern Pacific. Further modeling experiments are needed to quantitatively determine the differing atmospheric responses to early and late winter El Niño forcing.

During 2011/12 to 2014/15, California experienced the driest four successive winters since 1895 (Williams et al. 2015). The accumulated precipitation deficit for the 4-year period reached 148% in northern California and 195% in southern California of the winter precipitation climatology, which means a very strong El Niño like 1982/83 or 1997/98 might be able to remove the statewide accumulated precipitation deficit within one winter (Figure 1.2). The 2015/16 El Niño, as one of the strongest El Niño events in recent history, was thought to have contributed to several severe storms to California in December 2015 and January 2016, causing serious flash flooding and landslides in southern California. The impacts of the 2015/16 El Niño on California winter precipitation will be examined in the next chapter.
Chapter 2. Role of Equatorial Pacific SST forecast error in the late winter California precipitation forecast for the 2015/16 El Niño

Note: This chapter has been published in *Journal of Climate* (Jong et al. 2018)

2.1 Introduction

El Niño, the warm phase of the El Niño-Southern Oscillation (ENSO) cycle, has far-reaching impacts on seasonal weather anomalies and interannual climate variability across the globe (e.g., Ropelewski and Halpert 1987, 1996; Kiladis and Diaz 1989; Mason and Goddard 2001; Larkin and Harrison 2005; Chiodi and Harrison 2015, and many others). During an El Niño event, anomalously positive sea surface temperature (SST) and enhanced deep convection in the tropical Pacific force an upper level stationary wave. The stationary wave links the tropical forcing to extratropical climate, particularly in the Pacific/North American (PNA) region (e.g., Wallace and Gutzler 1981; Seager et al. 2010), imposing a major control on the weather across western North America, including California (e.g., Ropelewski and Halpert 1986; Harrison and Larkin 1998; Ely et al. 1994; Cayan et al. 1999; Schubert et al. 2008, and many others). The statistical link between El Niño and California winter precipitation, however, is less robust than elsewhere in southwest North America. The influences of El Niño vary from event to event (e.g., Schonher and Nicholson 1989; Hoerling and Kumar 1997) and are highly dependent on the region and time of year with, for southern California precipitation, the late winter impact showing the strongest signal (Jong et al. 2016; Chapter 1). The state of California, one of the largest economies in the world as well as a major state for agricultural production, has experienced one of its worst droughts in the past five
years (e.g., Seager et al. 2015a; Williams et al. 2015) and it continued, surprisingly, despite the strong 2015/16 El Niño. A better understanding of El Niño’s varying impact on California winter precipitation could potentially enhance the predictability of seasonal to interannual variability across the state, including drought onset and termination, and provide economic and societal benefit.

During an El Niño event, as the jet stream and extratropical storm track move southward, California, particularly relatively dry southern California, tends to get excessive amount of precipitation (e.g., Schonher and Nicholson 1989; Cayan et al. 1999; Andrews et al. 2004; Hoell et al. 2016; Jong et al. 2016; Chapter 1). The probability of anomalous statewide-wetness increases as El Niño intensity increases, according to both historical observations (Jong et al. 2016; Chapter 1) and model simulations (Hoell et al. 2016). Furthermore, the influences of El Niño on California winter precipitation are statistically significant in late winter (February-April), but not in late fall or early winter even though that is when an El Niño usually reaches its peak intensity (Lee et al. 2008, 2014; Jong et al. 2016; Chapter 1).

The 2015/16 El Niño, one of the strongest events in recent history, was comparable in strength to the 1982/83 and 1997/98 strong El Niño events. During the peak season of the event (from November 2015 to January 2016), climate models generally predicted wetter-than-normal conditions in the southern tier of the US, including California, from December-January-February (DJF) to late winter February-March-April (FMA) (Climate Prediction Center 2015a,b, 2016; Steinschneider and Lall 2016). However, the 2015/16 event, despite expectation of a good probability of excess precipitation and drought relief (Seager et al. 2015b), only brought an about average amount of precipitation to northern California (NoCal, 101%) and below average precipitation to southern California (SoCal, 81%) in FMA 2016 (Table 2.1). It is therefore
interesting to ask why California did not receive the excess precipitation expected and predicted during the 2015/16 El Niño.

There are several possible reasons that cause variations in North American climate between El Niño events. These include random atmospheric internal variability and sensitivity to differences in the detailed structure and longitudinal location of the SST anomalies (SSTA) (e.g., Hoerling and Kumar 1997; Guo et al. 2017). The different longitudinal locations of the SSTA can cause shifts in the forced stationary waves, leading to different teleconnection patterns and impact on North American climate (e.g., Mo and Higgins 1998a,b; Barsugli and Sardeshmukh 2002; Hoerling and Kumar 2002). For some local regions, such as the Pacific Coast, the precipitation anomalies can be extremely sensitive to small shifts in the teleconnection patterns. For example, a slight shift in the anomalous Aleutian Low during an El Niño event can differentiate between a dry and wet winter in California (e.g., Schonher and Nicholson 1989; Ely et al. 1994). Many researchers have discussed the sensitivity of North American climate to the diversity of El Niño based on the longitudinal locations of the SSTA (e.g., Larkin and Harrison 2005b; Weng et al. 2009; Yu and Zou 2013; Capotondi et al. 2015; Taschetto et al. 2016; Infanti and Kirtman 2016; Guo et al. 2017). Two specific El Niño types that have been discussed are the Eastern-Pacific (EP) El Niño and the Central-Pacific (CP) El Niño (also termed as Dateline El Niño in Larkin and Harrison 2005b and El Niño Modoki in Weng et al. 2009). A CP El Niño generally enhances the dry anomalies and weakens the wet anomalies across most of the US regions. In California, however, they found the precipitation anomalies tend to be similar or even wetter during CP El Niño events as compared to canonical EP El Niño, particularly in SoCal (Weng et al. 2009; Yu and Zou 2013).

In this chapter, we try to understand the 2015/16 California precipitation responses from
the perspectives of both the characteristics and time evolution of the SSTA and the anomalous circulation patterns using observations of the three most recent strong El Niño events (1982/83, 1997/98, and 2015/16). Then, we examine the coupled climate forecast models from the North-American Multi Model Ensemble (NMME) to determine to what extent the models are able to capture the SST and California precipitation anomalies in the 2015/16 late winter. We also conduct atmospheric general circulation model (AGCM) experiments to test our hypothesis, derived from these analyses, that the forecasts were too wet because of SST forecasts that were too warm in the equatorial east Pacific during late winter.

Table 2.1 Niño 4 (160°E-150°W, 5°N-5°S), Niño3.4 (170°W-120°W, 5°N-5°S), Niño3 (150°W-90°W, 5°N-5°S) SSTA and northern (NoCal, 124°W-118°W and 36°N-42°N) and southern (SoCal, 122°W-114°W and 32°N-36°N) California precipitation (% of climatology) for the strong El Niños (Niño3.4 SSTA > 2°C in NDJ) in (upper) early and (lower) late winters from 1981/82 to 2015/16.

<table>
<thead>
<tr>
<th></th>
<th>Niño4</th>
<th>Niño3.4</th>
<th>Niño3</th>
<th>NoCal P</th>
<th>SoCal P</th>
</tr>
</thead>
<tbody>
<tr>
<td>NDJ</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>82/83</td>
<td>0.64 °C</td>
<td>2.48 °C</td>
<td>3.01 °C</td>
<td>147 %</td>
<td>179 %</td>
</tr>
<tr>
<td>97/98</td>
<td>0.78 °C</td>
<td>2.58 °C</td>
<td>3.45 °C</td>
<td>152 %</td>
<td>177 %</td>
</tr>
<tr>
<td>15/16</td>
<td>1.54 °C</td>
<td>2.74 °C</td>
<td>2.74 °C</td>
<td>119 %</td>
<td>85 %</td>
</tr>
<tr>
<td>FMA</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>82/83</td>
<td>0.52 °C</td>
<td>1.75 °C</td>
<td>2.01 °C</td>
<td>212 %</td>
<td>243 %</td>
</tr>
<tr>
<td>97/98</td>
<td>0.40 °C</td>
<td>1.43 °C</td>
<td>2.01 °C</td>
<td>207 %</td>
<td>289 %</td>
</tr>
<tr>
<td>15/16</td>
<td>1.20 °C</td>
<td>1.70 °C</td>
<td>1.43 °C</td>
<td>101 %</td>
<td>81 %</td>
</tr>
</tbody>
</table>

2.2 Data

2.2.1 Observed Data

The SST data are taken from the NOAA Optimum Interpolation sea surface temperature analysis version 2 (OIv2). OIv2 provides monthly SST data from 1981 to present with 1°×1° spatial resolution (Reynolds et al. 2002). Precipitation data over North America from 1981 to 2016
are taken from the National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC) (Chen et al. 2002). The monthly global data with 0.5°×0.5° spatial resolution are reconstructed by interpolation of gauge observations from over 17,000 stations collected in the Global Historical Climatology Network (GHCN), version 2, and the Climate Anomaly Monitoring System (CAMS) datasets. Precipitation data over the Pacific are derived from NCEP CPC “Climate Anomaly Monitoring System (CAMS) and OLR Precipitation Index (OPI)” (CAMS_OPI) which merged observations from rain gauges with precipitation estimates from satellites (Janowiak and Xie 1999). CAMS_OPI provides monthly data from 1979 to the present with 2.5°×2.5° spatial resolution. Atmospheric circulation data (200hPa height) are taken from the NCEP/NCAR Reanalysis Project which is a joint project between NCEP and the National Center for Atmospheric Research (NCAR), providing monthly atmospheric analyses from 1948 to the present with 2.5°×2.5° spatial resolution (Kalnay et al. 1996). The monthly climatology for the winter months used in this study is consistently based on winter (November to April) 1981/82 to winter 2015/16.

2.2.2 Forecast Data

The SST and precipitation forecasts are derived from the North American Multi-Model Ensemble (NMME), an experimental multi-model seasonal forecasting system including coupled models from US NOAA/NCEP, NOAA Geophysical Fluid Dynamics Laboratory (GFDL), NCAR, National Aeronautics and Space Administration (NASA), and Canadian Meteorological Center (CMC) (Kirtman et al. 2014). Eight models from the NMME are used in this research: NCAR CCSM3, NCAR CCSM4, GFDL-CM2.1, GFDL-CM2p5, NASA-GMAO, NCEP CFSv2, CMC1-CanCM3, and CMC2-CanCM4. All the model data are provided with a spatial resolution of 1°×1°.
In this study, we examine the February-March-April (FMA) 2016 3-month average forecast initialized with February 01 2016 atmospheric and oceanic conditions. The NMME data is accessible at the IRI (International Research Institute for Climate and Society, Columbia University) Data Library (http://iridl.ldeo.columbia.edu/SOURCES/.Models/.NMME/).

2.2.3 AGCM Experiments

To test California precipitation sensitivity to the tropical Pacific SST anomaly (SSTA), we conduct AGCM experiments by prescribing the observed and forecast SSTA. The atmosphere model used is the NCAR Community Climate Model version 3 (CCM3, Kiehl et al. 1998), Community Atmosphere Model version 4 (CAM4, Neale et al. 2013), and version 5 (CAM5, Neale et al. 2012). CCM3 has horizontal resolution of triangular spectral truncation T42 with 128×62 longitude/latitude cells (approximately 2.8°×2.8°) and 18 levels with the model top at 2.9hPa. CAM4 and CAM5 both have 144×96 longitude/latitude cells as horizontal resolutions (approximately 1.9°×1.9°). The vertical resolution in CAM4 is 26 levels with the model top at 2.2hPa. CAM5 has 4 extra levels in the boundary layer (below 2200m) with a total of 30 vertical levels.

In each model, we conduct three sets of 100 member ensemble experiments. For each ensemble member, a random perturbation of the order of 10^{-14} was added to the initial 3-D temperature field on January 1st, 2016 and ran for a month to reach equilibrium. For the first set of experiments, the climatological FMA 3-month averaged SSTs from 1982 to 2016, derived from OIv2 were prescribed globally and generate the control runs. For the anomaly runs, FMA 2016 3-month averaged SSTA was added to the climatological SSTs only in the tropical Pacific from 30°S to 30°N with additional 5° width edges for tapering. Two SSTAs were applied, one is the OIv2
observed SSTA (OBS) and one is the NMME multimodel mean 3-month forecast made on Feb 01 2016 (FRCST). Thus, each model there are three experiments: control run with climatological FMA SST, observed FMA 2016 SSTA runs, and forecast FMA 2016 SSTA runs.

2.3 Results

2.3.1 Comparison of SST and circulation during the three strong El Niño events

The 2015/16 El Niño was among the strongest El Niño events since records began, with Niño3.4 (170°W-120°W, 5°N-5°S) SSTA reaching 2.74°C in the early winter (November-December-January, NDJ) and 1.70°C in the late winter (February-March-April, FMA) (Table 2.1). The other two comparably strong events, 1982/83 and 1997/98, were slightly weaker in terms of the NDJ Niño3.4 index but were stronger in terms of the FMA Niño3.4 index than the corresponding 15/16 event. Both these prior events brought excessive amounts of precipitation to California in the late winter: 212% and 207% of the climatology for NoCal and about 243% and 289% for SoCal in 1982/83 and 1997/98, respectively (Table 2.1). In contrast, the 2015/16 event only brought about average precipitation to NoCal and below average precipitation to SoCal in both early and late winter 2015/16. To examine the possibility that distinct ocean conditions induced different teleconnection patterns and impacts on California precipitation for these three events, here we compare the associated tropical Pacific SST anomalies and anomalous atmospheric circulation patterns (Figure 2.1). As previous studies suggested, El Niño’s impact changes throughout the winter (Lee et al. 2014; Jong et al. 2016; Chapter 1), so we separate into early and late winter in Figure 2.1.
In the early winter, all three events had warm SSTA patterns occupying the entire eastern tropical Pacific from the dateline to coastal South America (Figure 2.1a-c), with Niño3 (150°W-90°W, 5°N-5°S) SSTAs being warmer than Niño4 (160°E-150°W, 5°N-5°S) SSTAs (Table 2.1). However, in the 2015/16 early winter, the SSTA maximum was located slightly westward compared to the 1982/83 and 1997/98 early winters, resulting in a slightly smaller Niño3 SSTA, but stronger Niño4 SSTA, than the other two strong events (Table 2.1). Figure 2.2 shows that the anomalous tropical deep convection in the 2015/16 early winter was located closer to the dateline.
and extended less to the eastern tropical Pacific than the prior events. As a possible consequence, the longitudinal location of the teleconnections pattern was also shifted westward during the 2015/16 early winter. The anomalous Aleutian Low, a classic mid-latitude response to El Nino, was located just off the west coast of North America (Figure 2.1c), unlike the patterns during 1982/83 (Figure 2.1a) and 1997/98 (Figure 2.1b) when low-pressure anomalies extended across North America.

In the 1982/83 and 1997/98 late winters (Figure 2.1d and e), though the amplitude of the SSTA had weakened, the patterns remained similar to those in early winters, with maximum SSTA centered at around 120°W and Niño3 SSTAs of ~2°C. The precipitation anomalies over the tropical Pacific extended further eastward (Figure 2.2d and e), compared to the early winters, since the total SSTs in the eastern tropical Pacific were warmer in late winters (due to the climatological warming of the cold tongue region). The low-pressure anomalies extended zonally from the North Pacific across North America (Figure 2.1d and e), which would steer storms and precipitation to California, causing the extremely wet late winters in the state as shown in Table 2.1. However, in FMA 2016 (Figure 2.1f), the SSTA maximum retreated further westward, resulting in Niño4 SSTA of similar strength to Niño3 SSTA. The anomalous tropical deep convections (Figure 2.1f) was also shifted substantially westward, showing a lack of anomalous convection in the eastern Pacific especially as compared to the previous two strong events. We hypothesize that the westward shifted patterns of SSTA and tropical deep convections prevented the low-pressure anomalies in the extratropical North Pacific from extending eastward to reach the North American landmass (Figure 2.1f), resulting in near normal - below normal precipitation in California.
Figure 2.2 NCEP-CPC precipitation anomalies (unit: mm/day) for (upper) 1982/83, (middle) 1997/98, and (lower) 2015/16 (left) early and (right) late winters.

The time evolutions of the SSTA in the central (Niño4 region) and eastern (Niño3 region) tropical Pacific (Figure 2.3) illustrate more clearly the similarity and disparity of the SSTA patterns for these three events. During the 2015/16 event, the Niño4 SSTA was warmer than the previous two events throughout the time period of the events (Figure 2.3a), while, after the peak of the events, the Niño3 SSTA was weaker (Figure 2.3b). In the 2015/16 late winter, as Niño3 SSTA dropped faster than Niño4 SSTA, Niño3 and Niño4 SSTAs reached comparable strength, and the SSTA pattern shifted further to the west, compared to the 1982/83 and 1997/98 events. The SSTA
Figure 2.3 The evolution of (a) Niño4 (160°E-150°W, 5°S-5°N) and (b) Niño3 (150°W-90°W, 5°S-5°N) three-months running average SSTAs from June-July-August (J) to March-April-May (A) during 1982/83 (dotted line), 1997/98 (dashed line), and 2015/16 (solid grey line), based on OIv2 observed data. NMME model-ensemble mean forecasted Niño4 and Niño3 SSTAs for 2015/16 period is shown in red thick line, where every three-months running average forecast is initialized with the first day of the first month. (e.g. June-July-Aug (J) is initialized with June 01 2015.) Also, the time-longitude plots of SSTA along the equatorial Pacific (5°S-5°N averaged) during (c) 1982/83, (d) 1997/98, and (e) 2015/16 are shown. NMME model-ensemble mean forecasted equatorial Pacific SSTA is shown in (f).
Hovmöller plots (Figure 2.3c-e) show that in all three events SSTAs weakened from their peak in NDJ to FMA. They also show that the maximum SSTA in FMA was further west for 2016 than for 1983 and 1998. Indeed, for FMA 1998 the maximum SSTA was at the coast of the Americas. The reasons for these SSTA evolutions and differences are beyond the scope of this research.

The knowledge that El Niño’s impact on California precipitation occurs mainly in late winters (Jong et al. 2016; Chapter 1) and the fact that California was extremely wet during FMA 1983 and 1998 but dry during FMA 2016 justifies the hypothesis that the weak Niño3 SSTA during the 2015/16 late winter may have played a role in suppressing the impact of El Niño 2015/16 on California precipitation. We begin exploring this question by first examining, in the next section, the forecast models from the North-American Multi Model Ensemble (NMME) to determine if there is any similar relationship between forecast SSTA and California precipitation.

2.3.2 Forecast of winter 2015/16 from the NMME

Here we examine the coupled climate forecast models from the NMME to determine the agreements and discrepancies between forecasts and observations in winter 2015/16, focusing on the FMA 2016 3-month average and the NMME forecast initialized on February 1 2016.

Figure 2.4 shows the spatial patterns of the forecast SSTA for FMA 2016 from the eight NMME models (a-h), the multi-model ensemble mean (i) and observations (j). Four of the eight models (COLA-RSMAS-CCSM3 (a), GFDL-CM2p5-FLOR-B01 (d), NASA-GMAO (e), and CMC2-CanCM4 (g)), as well as the model ensemble mean forecast a pattern of SSTAs that are much stronger than observations (j) in the eastern tropical Pacific. The forecast SSTAs from the COLA-RSMAS-CCSM4 (b) and NCEP-CFSv2 (h) models have about the right strength but the patterns were too far to the east compared to observations. The GFDL-CM2p1-aer04 (c) model
forecast shows a spatial pattern that is similar to observations with the maximum SSTA near the dateline, even though the strength of SSTA is slightly stronger than the observations. The CMC1-CanCM3 (f) produces an SSTA that is similar in both pattern and amplitude to observations. Figure 2.4k presents the scatter plot of Niño4 versus Niño3 SSTAs for each model, the model-ensemble mean and the observations, as a measure of the relative location of the SSTA. Except for CMC1-CanCM3 (light blue) and GFDL-CM2p1-aer04 (orange), the rest of the eight models and model-ensemble mean all overestimate the Niño3 SSTA which means they forecast warmer than observed conditions in the eastern equatorial Pacific in late winter. In summary, for most of the models, their ensemble mean did not capture the longitudinal location of the SSTA pattern during the late winter of this event. CMC1-CanCM3 and GFDL-CM2p1-aer04 are the two exceptions. This situation may be relevant to the NMME model bias of systematically producing too-warm SSTA in the eastern tropical Pacific for events that have more local warming in the central tropical Pacific (Kirtman et al. 2013; Infanti and Kirtman 2016).

We next examine forecast precipitation over the California region for FMA 2016 (Figure 2.5). Most of the models predict wetter than normal conditions: COLA-RSMAS-CCSM3 (a), NASA-GMAO (e), CMC2-CanCM4 (g), and NCEP-CFSv2 (h) predict a very wet late winter over the entire state, especially in SoCal; while COLA-RSMAS-CCSM4 (b) and GFDL-CM2p5-FLOR-B01 (d) predict a normal to drier winter in NoCal, but overestimate the amount of precipitation in SoCal. The two models that had their SSTA closest to observations, the GFDL-CM2p1-aer04 (c) and the CMC1-CanCM3 (f), both show drier than normal conditions in the north, and normal or dry conditions in the south, respectively, much closer to the observations than the other six models. The scatter plot for precipitation in NoCal versus SoCal (Figure 2.5k) also shows
that CMC1-CanCM3 (light blue) and GFDL-CM2p1-aer04 (orange) have the closest match to the observations, with neither NoCal nor SoCal being too wet.

Overall, CMC1-CanCM3 and GFDL-CM2p1-aer04 are the only two models that did not overestimate the Niño3 SSTA in the late winter. They were also the only two models that did not forecast an anomalously wet late winter in California. There is less consistency when comparing anomalous tropical convection between observations and the NMME model forecast, on the other hand. The forecast precipitation anomalies over the tropical Pacific (Figure 2.6a-h) suggest that all NMME models, including CMC1-CanCM3 and GFDL-CM2p1-aer04, overestimate the tropical eastern Pacific convection as compared to observations (Figure 2.6j). Thus, it is difficult to determine whether the two NMME models simulated realistic California precipitation for the right reason.

The time evolutions of Niño3 and Niño4 SSTA from the NMME model-ensemble 3 month forecast are also shown in Figure 2.3. The model ensemble was correctly forecasting a Niño4 SSTA for the 2015/2016 event warmer than 1982/1983 and 1997/1998 (Figure 2.3a), but did not capture the fast decay in Niño3 SSTA in FMA, resulting in Niño3 SSTA in the late winter warmer than observations (Figure 2.3b and f). The ensemble mean FMA forecast of the Niño3 SSTA for 15/16 was closer to what happened in 1982/1983 and 1997/1998. Therefore, to add to our hypothesis, we surmise that the majority of the state-of-the-art forecast models forecast too-high California precipitation because they overestimated the SSTA strength in the eastern tropical Pacific in late winter.
Figure 2.4 FMA 2016 forecast SSTA (unit: °C) from (a)-(h) each model and (i) model-ensemble mean. Observed FMA 2016 SSTA from OIv2 is shown in (j). Niño4 SSTA as a function of Niño3 SSTA for each model (color dots), model-ensemble (black diamond), and observation (red star) are shown in (k).
Figure 2.5 FMA 2016 forecast precipitation (% of climatology) over North America from (a)-(h) each model and (i) model-ensemble mean. Observed FMA 2016 precipitation from NCEP-CPC is shown in (j). Scatter plot for precipitation in northern California versus precipitation in southern California for each model (color dots), model-ensemble (black diamond), and observation (red star) are shown in (k).
Figure 2.6 FMA 2016 forecast precipitation (unit in mm/day) over the tropical Pacific from (a)-(h) each model and (i) model-ensemble mean. Observed FMA 2016 precipitation from NCEP-CPC is shown in (j).
2.3.3 AGCM experiments

In order to test if the too-warm SSTA forecast in late winter, especially in the eastern tropical Pacific, influenced the California precipitation forecast, we turn to the AGCM experiments forced with climatological, observed and forecast SSTAs. The precipitation and circulation anomalies shown in the Figure 2.7 and Figure 2.8 are the differences between the ensemble mean of the observed/forecast SSTA runs and control runs.

Figure 2.7 presents the 100-member ensemble mean precipitation anomalies over the US region from the observed (OBS) and forecast (FRCST) SSTA runs. All the experiments produce normal-to-wetter conditions in NoCal and wetter-than-average conditions in SoCal, but the differences between OBS and FRCST differ by model:

- In CCM3 (Figure 2.7a-c), FRCST predicts the entire state to be wetter than the climatology. In the OBS SSTA runs, while the southern part of the state is still wetter than the climatology, NoCal has close to normal conditions. Thus, differences between the OBS and FRCST show that the observed SSTA drives drier conditions over California region than does the forecast SSTA.

- In CAM4 (Figure 2.7d-f), both OBS and FRCST produce wetter-than-average conditions across the entire state with the OBS slightly drier in NoCal (at 90% statistical confidence level) and slightly wetter in SoCal (not statistically confident at 90% level).

- CAM5 is similar to CAM4, with both OBS and FRCST predicting wetter-than-average conditions across California. However, the observed SSTA, compared to the forecast SSTA, makes SoCal much wetter, with an extra 20% to 40% of the climatological precipitation. The differences in NoCal precipitation between these
two SSTA runs are not statistically significant, though the ensemble mean from the OBS is slightly drier than the FRCST.

**Figure 2.7** FMA 2016 US ensemble mean precipitation anomalies from AGCM (left) observed SSTA runs and (middle) forecast SSTA runs using (upper) CCM3, (middle) CAM4, and (lower) CAM5. The observed SSTA runs (OBS) minus forecast SSTA runs (FRCST) differences are shown in the right panels. Stippling denotes 90% significance using a two-tailed Student t-test. Red boxes indicate California region. (unit: % of climatology)
Figure 2.8 The observed SSTA runs (OBS) minus forecast SSTA runs (FRCST) differences of FMA 2016 ensemble mean precipitation (shaded, unit in mm/day over ocean and % of the climatology over land), 200hPa height (left panels, contour, interval:5m), 700hPa height (right panels, contour, interval:5m), and 850hPa wind (right panels, vectors).
Therefore, based on these experiments, the observed SSTA in all models tends to make NoCal drier than does the forecast SSTA. However, the models have no consensus on the variations in SoCal precipitation: CCM3, CAM4, and CAM5 have drier, about the same, and much wetter SoCal in the OBS, compared to the FRCST, respectively.

The differences in California precipitation between these two SSTA experiments and the differences among these three models could be explained by the differences in their teleconnection patterns. Figure 2.8 presents the differences in anomalous precipitation and circulations over the Pacific/North American region between the OBS and FRCST. In all of these three models, the observed, colder, SSTA, drives weaker convection in the eastern tropical Pacific compared to the FRCST (shaded area over the ocean in Figure 2.8). As the tropical convection in the eastern equatorial Pacific weakens, the deepening Aleutian Low is mitigated and shifted westward over the North Pacific across North America in the OBS, compared to the FRCST. The changes in the circulations can be identified in both low (Figure 2.8d-f) and high (Figure 2.8a-c) levels given the barotropic structure of the deepening Aleutian Low. The weakening and westward shift of the low-pressure anomalies in the OBS could subsequently result in drier conditions in California, as happens in the CCM3 experiments (Figure 2.8a-c). In contrast, in CAM4 and CAM5 SoCal gets wetter as the low-pressure anomalies weaken over western North America in the OBS runs, implying that California precipitation in these models are affected by other factors.

In the CAM4 and CAM5 experiments, besides the differences off the west coast of North America, the teleconnections also respond to the differences between OBS and FRCST in tropical SSTA in the western Pacific (Figure 2.8b and c). Unlike the dipole pattern in the CCM3 experiments, which is a direct response to SSTA in the eastern tropical Pacific, CAM4/5 show a wave-train-like response from the western tropical Pacific. Why the models have differing
precipitation anomalies over California can be seen by looking at the low level flow (Figure 2.8d-f). Wetting in OBS relative to FRCST in California and Mexico is related to, in the OBS minus FRCST model difference, southerly flow on the eastern flank of a low level cyclone over the North Pacific. In CCM3, the southerly flow difference is weak and over Mexico (and California is drier) while in CAM4 and CAM5 it is stronger and over California, wetting the state. It is not clear whether the differences between models in terms of the circulation anomalies over the coast of southwest North America are related to the different teleconnections patterns from tropical SSTs or not.

To examine how internal atmospheric dynamics may have contributed to the drier than expected California in 2015/16 winter, we examine the probability of NoCal and SoCal being wet or dry given the observed and forecast SSTA using the AGCM experiments. Figure 2.9 shows the histograms of California precipitation from the 100 OBS SSTA runs and 100 FRCST SSTA runs for each model. In NoCal (upper panels in Figure 2.9), in all three models, more members in the FRCST are wetter than the observations, compared to the OBS. In the CCM3 experiments (Figure 2.9a), 58% of the FRCST are wetter than the observation, compared to only 37% of the OBS. Also, in the OBS, the distribution is more centered towards the observations. However, in CAM4 and CAM5 (Figure 2.9c and e) the number of members drier than the observation is quite similar for OBS and FRCST (CAM4: 27% of OBS versus 23% of FRCST; CAM5: 25% of OBS versus 21% of FRCST) and OBS even has more members at the wet end of the distributions. The ensemble mean for OBS in CAM4 and CAM5 are about the same to slightly drier in NoCal compared to the FRCST (Figure 2.7f and i). Nevertheless, the observed precipitation is within the range of uncertainties of the prescribed SSTA for all models, indicating that internal variability could drive a near normal NoCal even in the presence of a strong El Niño.
Figure 2.9 Histograms of FMA 2016 (upper) NoCal and (lower) SoCal ensemble member precipitation (mm/day) from the OBS SSTA runs (black-outlined bars) and FRCST SSTA runs (grey bars) using (left) CCM3, (middle) CAM4, and (right) CAM5. Dotted lines indicate the observations from NCEP-CPC (3.31 mm/day in NoCal and 0.91 mm/day in SoCal), indicated as percent of climatology equivalent to each model’s value (i.e. for NoCal, 3.31 mm/day is equivalent to 101% of the observed climatology, the corresponding rainfall amount equivalent to that percentage in each model is shown as dotted line, in order to remove bias in model climatology). The percentage numbers indicate the percent of the runs in each model that are below (left) and above (right) the observations (black: OBS ; grey: FRCST).
The simulated precipitation in SoCal (lower panels in Figure 2.9) has even larger spreads compared to that in the North. In all three models, the distributions of precipitation anomalies from both SSTA runs show large spreads with long tails at the wet ends. Also, as discussed in the previous paragraphs, the responses in SoCal precipitation are inconsistent among these models. In the CCM3 experiments (Figure 2.9b), 83% of the FRCST and 76% of the OBS are wetter than the observations. However, the probability of being extremely wet (precipitation anomalies > 100% of the climatology) drops in the OBS (12%), compared to the FRCST (23%). In the CAM4 experiments, both OBS and FRCST show large variances in SoCal precipitation. The percentage of the runs drier than the observations and the variances are similar in both experiments. In CAM5, the observed SSTA increases the variance of SoCal precipitation compared to FRCST and the probability of being extremely wet (precipitation anomalies > 100% of the climatology) is enhanced from 35% in the FRCST to 40% in the OBS with 10% chance to be larger than 300% of the climatology. Thus, as shown in Figure 2.7i, the CAM5 ensemble mean of SoCal precipitation is much wetter in the OBS. However, once more, the observations are within the ensemble spread in all distributions shown in Figure 2.9, implying that internal dynamics alone can cause the dry SoCal in the presence of a strong El Niño.

2.4 Conclusions and Discussions

The 2015/16 El Niño event was one of the strongest ever, comparable to 1982/83 and 1997/98, both of which brought extremely wet late winters to all of California. In the late 2015/16 winter, this event, however, only brought about average precipitation to northern California while southern California was drier than normal allowing the multiyear drought to persist. The purpose of this paper is to examine a possible explanation of why this event did not bring excessive
precipitation to California in the late winter, as was forecast by most prediction models and expected based on observational and model-based analyses (Seager et al. 2015b).

We first compared the three strongest El Niños since records began (1982/83, 1997/98, and 2015/16) based on observations. In the 2015/16 winter, the maximum equatorial Pacific SSTA was located westward compared to that during 1982/83 and 1997/98 winters. This was particularly the case in the 2015/16 late winter, when the maximum SSTA weakened and retreated further to the west. The North Pacific low-pressure anomaly was located away from the North American coast in 2015/16, unlike the patterns during 1982/83 and 1997/98 when low pressure anomalies extended zonally across North America. These observations raised the question of whether the colder observed SSTA in the eastern tropical Pacific in the 2015/16 winter prevented the teleconnections from extending from the North Pacific across North America and bringing extra precipitation to California.

We then examined the forecast of SSTA and precipitation for February-March-April 2016 from the North-America Multi-Model Ensemble (NMME). The NMME model-ensemble overestimated the eastern tropical Pacific SSTA in the late winter, as the models did not well capture the fast drop of the Niño3 SSTA from its December 2015 peak. Consistently, the anomalous deep convection in the model-ensemble over the tropical Pacific extended further to the east than the observations. The model-ensemble also predicted a wetter late winter in California, especially in southern California, than the observations. Thus, consistent with the comparison among the three strongest events, we hypothesized that in FMA 2016 the too-warm Niño3 SSTA forecast drove too-strong deep convection anomalies in the eastern tropical Pacific, triggering a too far east teleconnection and a wet bias in the forecast California precipitation.
To test this hypothesis, we conducted two SSTA-forced experimental runs in three NCAR GCMs (CCM3, CAM4, and CAM5): one forced by the observed FMA 2016 SSTA and the other forced by the NMME model-ensemble mean forecast FMA 2016 SSTA. The observed SSTAs were colder in the central-eastern equatorial Pacific and slightly warmer in the westernmost tropical Pacific than the forecast SSTAs. In response, all three models had a weaker and westward shifted low height anomaly over the North Pacific and west coast of North America when the observed SSTA was prescribed. As the result, precipitation in northern California was either about the same (CAM5) or drier (CCM3 and CAM4) in the observed SSTA runs than in the forecast SSTA runs. However, over southern California the response of precipitation varied across models. One of the possible explanations was that in CCM3 the teleconnections responded mainly to the SSTA differences in the central-eastern equatorial Pacific which caused weaker low-pressure anomalies and a drier southern California in the observed SSTA runs. On the other hand, the teleconnection patterns in CAM4 and CAM5 were also sensitive to the small SSTA differences in the westernmost tropical Pacific, which influenced the teleconnected height response over North America. From a local perspective, subtle differences in the observed-minus-forecast model low level height difference meant that southerly anomalies at the west coast were weak and located over Mexico in CCM3 but stronger and over California in CAM4 and CAM5 creating wet anomalies in the latter two models. Thus, based on these experiments, we tentatively claim that the too-warm Niño3 SSTA forecast might be partly responsible for the too-wet northern California forecast but it is difficult to claim it influenced the too-wet precipitation forecast for southern California in FMA 2016.

Nevertheless, the influences of atmospheric internal variability on California precipitation cannot be neglected, in spite of the strong tropical forcing (e.g., Hoerling and Kumar 1997). The
observed FMA 2016 California precipitation amounts are within the range of ensemble members in the AGCM experiments. Hence, a near-normal NoCal and a dry SoCal could have been driven by internal variability even in the presence of a strong El Niño. In the NMME forecast, the observed California precipitation amounts were also within the range of uncertainties even though more ensemble members prefer wetter-than-observed conditions (63% in NoCal, 74% in SoCal), resulting in the biased ensemble mean forecast. The 2015/16 California precipitation forecast was only “failed” in terms of the ensemble mean, not the ensemble spread.

Although we have not been able to prove our hypothesis that a forecast of too warm water in the eastern equatorial Pacific led to a forecast of too wet conditions across California in late winter, the observational and modeling work does show that such SST differences matter for atmospheric circulation and precipitation over North America. However, models disagree on the details of the circulation response, which can actually cause different models to respond to the same SST differences with opposite signs of precipitation differences over southern California. This makes clear that improved prediction over California will require improved SST forecasts and improved simulation of the atmospheric response to forecast SST anomalies.
Chapter 3. ENSO teleconnections and impacts on US summertime temperature during multi-year La Niña life-cycle

3.1 Introduction

The El Niño – Southern Oscillation (ENSO) influences the interannual variability of North American hydroclimate not only in winter (e.g., Ropelewski and Halpert 1986, 1987; Mason and Goddard 2001; Larkin and Harrison 2005; Jong et al. 2016; Chapter 1; many others) but also in summer (e.g., Ropelewski and Halpert 1986; Ting and Wang 1997; Wang et al. 2007). Previous studies have suggested that ENSO can exert significant impacts on crop yields over North America during the summer growing season (e.g. Handler 1984; Iizumi et al. 2014; Anderson et al. 2017). However, the less-established understanding of ENSO summer teleconnections might be leading to poor forecasting skill in the Northern Hemisphere summer extratropical circulations, in sharp contrasts to the demonstrated skill of boreal winter ENSO-based seasonal climate forecasts (e.g., Wang et al. 2009; Ding et al. 2011). To address the knowledge gap in ENSO summer teleconnections, this study focuses on the different physical mechanisms of summer teleconnections and characteristics of remote impacts on the US in the summer due to the different temporal evolution of the multi-year evolution of ENSO.

A typical ENSO event develops in late boreal spring, peaks at the end of the calendar year, and decays in the following spring to early summer (e.g., Rasmusson and Carpenter 1982; Okumura and Deser 2010). During an ENSO event, anomalous tropical deep convection induced by sea surface temperature (SST) anomalies triggers an upper-level Rossby wave propagating from the equator to the extratropics across the Pacific-North America (PNA) region (e.g., Hoskins and Karoly 1981; Webster 1981). The low-frequency Rossby wave shifts the subtropical jet stream
and storm track equatorward (poleward) during an El Niño (La Niña), subsequently influencing climate in remote regions including North America (e.g., Trenberth et al. 1998). Besides the direct tropical influence via Rossby wave dispersion, midlatitude transient eddies also play an important role in maintaining and modulating the extratropical response to the ENSO tropical forcing through an eddy-mean flow positive feedback (e.g., Hoerling and Ting 1994; Harnik et al. 2010; Seager et al. 2010). Both mechanisms are tightly linked to the intensity and location of the subtropical jet stream (e.g., Hoskins and Ambrizzi 1993; Hoerling and Ting 1994). Thus, the teleconnections and their impacts on extratropical North America are the strongest in the boreal winter when the ENSO tropical forcing reaches its peak and the jet-stream is strong and shifts closer to the tropics, allowing the Rossby wave source originating from tropical diabatic heating anomalies to extend into westerly flows and, hence, allowing Rossby wave propagation into mid-latitudes (e.g., Webster 1982).

These typical features of boreal winter climate, including both the ENSO tropical forcing and the mean locations of jet-stream and storm track, differ in the summer season. The intensity of teleconnections is much weaker as the anomalous tropical SST and deep convection are in either the developing or decaying phases of ENSO. Further, the dominance of tropical easterlies and the weaker and poleward-shifted North Pacific jet stream limit the potential for Rossby wave propagation out of the tropics into the extratropical region (Hoskins and Karoly 1981; Webster and Holton 1982). The difficulties in establishing the regional impacts of ENSO summer teleconnections are also aggravated by stronger land-atmosphere interactions in the summer season, which, over North America, can be comparable to the impact of remote SST forcing (e.g., Koster et al. 2000; Douville 2010). These factors constrain our knowledge of ENSO teleconnections and potentially limit the model forecasting skill of seasonal regional impacts on North America.
Despite the limitations, the previous literature has demonstrated the possibility that ENSO tropical forcing can trigger Rossby waves propagating toward higher latitudes in the summer season (e.g., Lau and Peng 1992; Ding et al. 2011; Douville et al. 2011) and impact US summer climate such as variability in Great Plains rainfall (Ting and Wang 1997; Hu and Feng 2001) and the Great Plains low-level jet (Weaver and Nigam 2008; Liang et al. 2015). In particular, a continental-scale anomalous anticyclone typically sits over North America in the summer of a developing La Niña and thereby leads to hot and dry summers over the central US (Wang et al. 2007). The strong rise in maximum temperature and decrease in precipitation over major crop-producing area of the US in the developing La Niña summer were found to negatively affect maize and soybean yields (Anderson et al. 2017). The 1988 US drought, when the strongest La Niña event since 1980 was underway in the tropical Pacific, was one of the examples (e.g., Trenberth et al. 1988; Trenberth and Branstator 1992). The tremendous agricultural and socioeconomic losses, fatalities, and damage from forest fires during the 1988 drought (Trenberth and Branstator 1992) highlight the importance and urgency of better understanding the physical mechanisms that control the extratropical teleconnections in the developing La Niña summers. In establishing the physical processes of ENSO summer teleconnections, however, the multi-year evolution of ENSO was rarely considered in the previous literature.

The importance of the multi-year ENSO evolution originates from the nonlinearity and asymmetry in the evolution and duration of El Niño and La Niña events. A La Niña tends to persist through the following summer and often re-intensifies in the subsequent winter, leading to a multi-year La Niña event, especially since the 1980s (Rasmusson and Carpenter 1982; McPhaden and Zhang 2009; Okumura and Deser 2010). Unlike La Niña, an El Niño tends to decay rapidly in the tropical Pacific in the boreal spring, but El Niño-induced warming in the Indian Ocean can persist
into the following summer and impact the global circulation, especially in the PNA region (e.g., Lau et al. 2005; Xie et al. 2009). In fact, all the first-year La Niñas since 1950 transitioned from El Niño winters (https://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php).

Therefore, La Niña summers can be when an El Niño is transitioning to a La Niña or when a La Niña is persisting. These two different cases were both loosely defined as “developing La Niña” in most of the previous studies despite the distinct prior ENSO conditions. The difference in the prior El Niño or La Niña conditions may also lead to distinct teleconnections in these two different La Niña summers, one transitioning from El Niño and one persisting from La Niña. For example, the aforementioned drops in the US maize and soybean yields are uniquely observed in the developing summer of a first-year La Niña, that is when an El Niño is transitioning to a La Niña, but not in the developing summer of second- or third-year La Niñas, that is when a La Niña is persisting (Anderson et al. 2017). The different agricultural impacts imply that these summer teleconnections may involve different dynamics, which has not been explored in any prior work.

In this chapter, we focus on distinguishing the features of teleconnections between the two different La Niña summers (transitioning versus persisting) based on observations. The goal is to understand the physical processes that lead to the strong anomalous anticyclone which is unique in the developing summer of a first-year La Niña, that is when El Niño is transitioning to La Niña. A stationary wave model (SWM) is used to characterize the relationships between ENSO tropical forcings and teleconnections in the two types of La Niña summers. In section 3.2, we detail the observational data and the stationary wave model used. In section 3.3.1, we compare the evolutions of the two types of La Niña cases from the preceding winters to the following La Niña winters based on the observations. We also identify the sources that lead to the different teleconnections
in the two developing La Niña summers. In section 3.3.2, we use the SWM as a diagnostic tool to test the hypothesis derived from the observational analyses. Conclusions and discussions are provided in section 3.4.

3.2 Data and Method

3.2.1 Observed Data

SST data are taken from the Extended Reconstructed Sea Surface Temperature version 5 (ERSSTv5, Huang et al. 2017). ERSSTv5 provides monthly SST data from 1895 with 2° × 2° spatial resolution. Atmospheric circulation (200hPa geopotential height and wind) and global precipitation data are taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Reanalysis I (Kalnay et al. 1996). This dataset provides monthly values from 1948 to the present with 2.5° × 2.5° spatial resolution for pressure level data and T64 Gaussian grid for surface data. For monthly surface temperature over land area, we use the 0.5° × 0.5° spaced Climate Research Unit TS3.26 (Harris et al. 2014) available from 1901 to 2016. The monthly climatology used in this study is consistently based on averages from January 1950 to December 2014. The SST and surface temperature over land area are both linearly de-trended and the trend is removed for each 3-month season separately.

3.2.2 Definition of El Niño and La Niña events

El Niño and La Niña events are selected based on the Oceanic Niño Index (ONI), a 3-month running mean of SST anomalies in the Niño3.4 region (5°N-5°S, 170°-120°W) from ERSSTv5, relative to a 30-year climatology. The 30-year base period is updated every 5 years and
centered to the first year of these 5 years (see the NOAA Climate Prediction Center (CPC) website: https://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_change.shtml for a complete description). El Niño and La Niña events are defined when the October-November-December ONI reaches the threshold of +0.5°C and -0.5°C (corresponding to about half standard deviation), respectively.

Figures 3.1 Evolutions of the Oceanic Niño Index for the first-year La Niña during 1950 to 2014 from the previous year to the following two years. Purple, orange, and blue lines are for the evolutions of single-year, two-year, and three-year La Niñas, respectively. Circles, triangles, and diamonds indicate the first-year, second-year, and third-year La Niña winters, respectively. Dotted line indicates the -0.5°C threshold used to define La Niña events.

Based on this criteria, we identified 4 single-year La Niña events from 1950 to 2014 (1964, 1988, 1995, 2005, indicated by purple lines in Figure 3.1), 5 two-year La Niña events (1954-55, 1970-71, 1983-84, 2007-08, 2010-11, blue lines in Figure 3.1), and 2 three-year La Niña events (1973-75, 1998-2000, orange lines in Figure 3.1). Therefore, there are 11 first-year La Niña winters.
(indicated by the black dots in Figure 3.1). The preceding winters of these first-year La Niña were all identified as El Niño winters (Figure 3.1). We categorize the summers in the first-year La Niña developing phase as “transitioning summer” (denoted as JJA(0)T in all the figures). On the other hand, there are 7 second-year La Niña winters (triangles in Figure 3.1) and 2 third-year La Niña winters (diamonds in Figure 3.1). We categorize the summers prior to these La Niña winters as “persisting summer” (denoted as JJA(0)P).

3.2.3 Stationary wave model (SWM)

The time-dependent baroclinic model used in this study is based on the three-dimensional nonlinear primitive equations in sigma (σ) coordinates. The basic variables in this model are the deviations from a prescribed zonally varying climatological mean state in response to imposed zonally asymmetric forcings. In order to find a steady state solution, we damp out the transient eddies with a 15-day interior Rayleigh drag and a 15-day Newtonian relaxation along with a scale-selective biharmonic diffusion with the coefficient of $1 \times 10^{17}$. The model includes 24 vertical σ levels and a rhomboidal truncation at wavenumber 30 in the horizontal (R30, roughly 2.25° latitude × 3.75° longitude). We run the model for 80 days and the average from days 30 to 80 is shown. The SWM has been widely used as a diagnostic tool to examine the mechanisms of ENSO stationary waves in both boreal winter seasons (e.g., Ting and Hoerling 1993; Hoerling and Ting 1994) and summer (e.g., Liu et al. 1998). More details are described in Ting and Yu (1998) and Simpson et al. (2015).

The basic state is the observed three-dimensional June-July-August (JJA) 3-month averaged climatology (1950-2014), including temperature, horizontal wind, and surface pressure fields, obtained from the NCEP-NCAR R1. The diabatic forcings are derived from the composites
of anomalous diabatic heating for transitioning La Niña summer (JJA(0)ₜ) and persisting La Niña summer (JJA(0)ₚ). Diabatic heating is calculated as a residual from the three-dimensional thermodynamic equation, constructed by monthly temperature and wind fields from NCEP-NCAR R1 and the transient eddy sensible heat flux convergences. As the SWM does not explicitly simulate transient eddies, the effects of midlatitude transient eddies are included by adding them as an additional forcing term. Both the transient heat and vorticity flux convergences are computed from the NCEP-NCAR R1 daily temperature and wind fields.

**3.3 Results**

**3.3.1 Observations**

3.3.1.1 Evolution of SST anomalies

The fundamental difference between the transitioning and persisting La Niña summers originates from the evolutions of oceanic conditions. Figure 3.2 illustrates the evolutions of the SST anomalies from the preceding winters to the La Niña winters. For the transitioning La Niña, SST anomalies over the tropical Pacific evolve from an El Niño state (Figure 3.2a) to a La Niña state (Figure 3.2d). During the preceding El Niño winter, warm SST anomalies extend from the tropical central Pacific (CP) to the eastern Pacific (EP) and these decay rapidly in the following spring (Niño3.4 SST anomalies drops from 1.45°C to 0.62°C, Figure 3.2b). By the transitioning summer JJA(0)ₜ (Figure 3.2c), the tropical Pacific has turned into a La Niña state with negative SST anomalies from the tropical CP to EP. Contrary to the rapidly evolving tropical CP and EP, the warming SST anomalies over the Indo-western Pacific and the tropical Atlantic persist from
Figure 3.2 Composites of de-trended ERSSTv5 SST anomalies (shaded over the ocean; °C), NCEP-NCAR R1 de-trended surface temperature (shaded over the land; °C) and 850hPa geopotential height anomalies with the zonal-mean removed (contours; interval: 5m) for the (left) transitioning and (right) persisting La Niña summers from (a,e) the preceding winters December-January-February (D(-1)JF(0)), (b,f) the preceding springs March-April-May (MAM(0)), (c,g) the developing La Niña summers JJA(0), to (d,h) the La Niña winters D(0)JF(1). Stippling denotes the 90% significance for de-trended SST anomalies using a two-tailed Student’s t-test. Thick lines indicate the 90% significance for 850hPa height variations. For surface temperature over the land area, only statistically significant values (at 90% level) are present.
the preceding winter to the transitioning summer. The warming over the Indo-western Pacific in the boreal spring to summer is a classic delayed response to a decaying El Niño (e.g., Lau et al. 2005; Xie et al. 2009) In other words, the tropical Indian and Pacific Oceans during the transitioning summer possesses the anomalies from both the decaying El Niño and the developing La Niña.

On the other hand, the oceanic conditions during a persistent La Niña evolve differently (Figure 3.2e-h). In the first-year La Niña winter, cold SST anomalies extend from the tropical CP to EP, as well as the Indian Ocean and the tropical Atlantic (Figure 3.2e). Following the peak season, unlike El Niño events, the tropical Pacific SST anomalies decay slowly, with Niño3.4 SST anomalies changing from -1.24°C in the winter to -0.81°C in the spring, showing the asymmetry in the duration between El Niño and La Niña evolutions (Figure 3.2f). In the persisting summer JJA(0)P (Figure 3.2g), the negative SST anomalies over the tropical Pacific remain with slightly weaker intensity compared to the preceding winter and spring. Compared to the transitioning summer (Figure 3.2c), the spatial distribution of the tropical Pacific SST anomalies is more meridionally extended. Also, the entire tropics are colder than normal, distinct from the transitioning summer in which the developing La Niña in the tropical Pacific was surrounded by warm anomalies in the Indian Ocean and tropical Atlantic persisting from the decaying El Niño.

3.3.1.2 Tropical rainfall anomalies

The distinct oceanic characteristics lead to different atmospheric responses over the tropical Pacific. For transitioning La Niña events, over the tropical CP, enhanced rainfall triggered by the El Niño warm SST anomalies (Figure 3.3a) evolves into weak reduced rainfall anomalies triggered by the developing La Niña SST anomalies (Figure 3.3c). During the transitioning summer, besides the suppressed deep convection over the CP, another significant region of
suppressed deep convection appears in the subtropical western Pacific (WP; Figure 3.3c). In term suppressed deep convection in the subtropical WP is likely caused by the baroclinic Kelvin wave forced by enhanced precipitation over the warm Indian Ocean (Figure 3.2c) which triggers low level divergence and upper level convergence in the subtropical WP (Xie et al. 2009). Therefore, during the transitioning summer, there is suppressed deep convection over the CP, due to the developing La Niña, and over the WP, due to the decaying El Niño.

On the other hand, the warming in the Indian Ocean and the suppressed rainfall over the subtropical WP are absent in the persisting summer preceded by a La Niña winter (Figure 3.2g & Figure 3.3g). Instead, only the suppressed deep convection induced by the negative La Niña SST anomalies is present over the tropical CP (Figure 3.3g). Accordingly, the primary difference in the anomalous rainfall field is the suppressed rainfall over the subtropical WP caused by the preceding El Niño, unique feature to the transitioning La Niña summer.

3.3.1.3 Anomalous 200hPa atmospheric circulations

Since ENSO teleconnections are forced by anomalous tropical convection, the distinct tropical rainfall patterns between the transitioning and persisting La Niña summers will lead to different teleconnection patterns. In the transitioning summer, significant anomalous atmospheric circulations extend from the tropics to the extratropics, with a significant anomalous anticyclone over northeastern North America (Figure 3.3c). The anomalous circulation pattern over the PNA region appears to be composed of two wave-trains: one from the suppressed convection over the tropical CP following the Great Circle Rossby wave route (Hoskins and Karoly 1981), with an anticyclone in the central North Pacific, a deepened Aleutian Low and the anticyclone over northeastern North America; and another originates from the suppressed convection over the
Figure 3.3 As in Figure 3.2, but for composites of precipitation anomalies (shaded; mm/day) and 200hPa geopotential height anomalies with the zonal-mean removed (contours; interval: 5m). Purple boxes in (c) and (g) indicate the subtropical WP and eastern North America regions used in Figure 3.5.
subtropical WP which triggers an additional Rossby wave propagating across the PNA region. This second wave-train is composed of an anomalous low near the suppressed convection, an high anomaly in the mid-latitude North Pacific (centered at around 40°N & 165°W and separate from the main high center caused by the CP cooling), a deepened Aleutian Low and the anomalous anticyclone over North America. We hypothesize that the two wave-trains superimpose on each other and constructively contribute to the anomalous anticyclone over North America. The extratropical teleconnections are essentially barotropic, extending down to the lower level and affecting the surface climate over the US (Figure 3.2c), as will discussed in the next sub-section.

For the persisting summer, however, statistically significant anomalous atmospheric circulations are confined in the tropics, although there are indications of a single wave-train emanating from the tropical CP (Figure 3.3g). This teleconnection, triggered by the weak suppressed convection in the tropical CP, is weak and is not augmented by a wave-train from the subtropical WP. Therefore, the teleconnection patterns over extratropical North America behave differently in these two La Niña summers: a superposition of teleconnections influence North America in the transitioning summer, but only a weak tropics to extratropics teleconnection exists in the persisting summer.

3.3.1.4 US surface temperature

The atmospheric teleconnections are the bridge connecting tropical forcing and extratropical meteorological conditions. Hence, the regional impacts of ENSO on the US surface climate are substantially different in these two developing La Niña summers. The evolution of the US surface temperature (Ts) for the transitioning La Niña presents the classic distribution of Ts anomalies during ENSO winters, warm (cold) north – cold (warm) south dipole pattern during
Figure 3.4 As in Figure 3.2, but for composites of CRU de-trended surface temperature. Stippling denotes the 90% significance for de-trended surface temperature anomalies using a two-tailed Student’s t-test. Boxes in (c) and (g) indicate the Midwest area used in Figure 3.5.
El Niño (La Niña) winters (e.g., Ropelewski and Halpert 1986; Figure 3.4a and d). For the transitioning summer (Figure 3.2c and Figure 3.4c), when the teleconnections reach extratropical North America, the anomalous anticyclone, with barotropic structure, exerts significant warm anomalies on most of the area east of the Rocky Mountain, especially over the Midwest region where the anomalies are more than 1 degree Celsius. The warming over the Midwest (box area in Figure 3.4c) is robust, as it happened in nine of the eleven historical transitioning summers from 1950 to 2014 (orange dots in Figure 3.5a). Also, the warming has been identified in the both the CRU (Figure 3.4c) and NCEP-NCAR R1 (Figure 3.2c) data, implying the warm anomaly is not sensitive to the particular data used. In addition, the anomalous anticyclone also leads to a dry tendency over the Midwest region, eight of the eleven historical transitioning summers brought drier-than-normal condition to the Midwest (Figure 3.5b).

For the persisting summer, the statistically significant parts of the teleconnections are mostly confined in the tropics and the remote impacts on extratropical North America are weak and insignificant (Figure 3.4g). Also, unlike in the transitioning summer, Ts anomalies over the Midwest shows no consistency among the historical persisting summers (blue dots in Figure 3.5a), with half of the events showing warm anomalies and half showing cold anomalies. The strong warming over the Midwest in the transitioning summer and the much weaker response in the persisting summer reinforce the substantial differences between these two types of La Niña summers and indicate the need for better understanding the dynamics underlying the different teleconnection patterns.
Figure 3.5 Midwest CRUv3p25 de-trended Ts and anomalous rainfall for all developing La Niña summers are present in (a) and (b), respectively. Scatter plots for JJA (c) subtropical WP rainfall versus 200hPa geopotential height anomalies over eastern North America, (d) subtropical WP rainfall versus Midwest de-trended surface temperature (Ts), and (e) 200hPa geopotential height anomalies over eastern North America versus Midwest de-trended Ts. Grey solid lines in (c)-(e) are the linear regression lines. The regions of subtropical WP and eastern North America are indicated in the Figure 3.3c and g. The region of Midwest is presented in Figure 3.4c and g. Black dots represent all the JJA in 1950-2014. Orange (blue) dots indicate transitioning La Niña summers JJA_T(0) (persisting La Niña summers JJA_T(0)).
3.3.1.5 The role of the WP suppressed convection

The differences in these two developing La Niña summers originate from the different ENSO temporal evolutions. The transitioning La Niña summer possesses the characteristics of both a decaying El Niño and a developing La Niña with suppressed convection over the both the subtropical WP and the tropical CP. On the contrary, the persisting La Niña summer only possesses the La Niña characteristics with suppressed convection over the tropical CP only. The primary difference is therefore the suppressed convection over the subtropical WP. This WP suppressed convection is a robust feature during the transitioning summer: 10 out of 11 historical transitioning summers experienced drier-than-normal rainfall over the subtropical WP (Figure 3.5c and d, orange dots). At the same time, positive 200hPa geopotential anomalies over eastern North America and the anomalously warm Midwest Ts tend to be associated with the suppressed convection in the subtropical WP (Figure 3.5c-e orange dots). Yet these features are not as connected to the subtropical WP in the persisting summer (Figure 3.5c and d, blue dots). Therefore, we hypothesize that this El Niño-induced WP suppressed convection and the associated Rossby wave strengthen the extratropical teleconnection patterns induced by the developing La Niña SST forcing, resulting in a strong anomalous anticyclone over the US during the transitioning summer.

To test the hypothesis, we first calculate the Rossby wave source (RWS) which represents the anomalous vorticity source produced by upper-level divergence due to anomalous convective activities in the tropics (e.g., Sardeshmukh and Hoskins 1988).

The RWS is defined as

\[ RWS = -\nabla \cdot \nabla \left( \tilde{\zeta} + f \right) - \left( \tilde{\zeta} + f \right) \nabla \cdot \tilde{V} \]

where \((\cdot)\) and \((\cdot)’\) represent the climatological three-month mean and perturbation, respectively, and \(\tilde{V}_x\) is the divergent component of the wind field, \(\zeta\) is the relative vorticity, and \(f\) is the Coriolis
parameter. The first term on the right-hand side represents the vorticity advection by anomalous divergent flow and the second term is the vorticity stretching term due to anomalous divergence.

Figure 3.6 presents the contribution to the RWS through the vorticity advection by the anomalous divergent flow (first term; upper panels in Figure 3.6) and through the stretching term due to anomalous divergence (second term; middle panels in Figure 3.6) during the transitioning and persisting La Niña summers. During the transitioning summer, significant positive vorticity forcing due to stretching is found near the suppressed convections in both the subtropical WP and tropical CP (Figure 3.6). This is expected from the local response to tropical thermal forcing: anomalous suppressed convection triggers anomalous convergence in the upper-levels which leads to low pressure anomalies north of the convergence and subsequently a Rossby wave propagation further downstream. In particular, the suppressed convection over the subtropical WP during the transitioning summer provides an anomalous vorticity source that induces Rossby wave propagation towards extratropical North America. On the other hand, during the persisting summer, the RWS due to anomalous upper-level convergence is only significant over the tropical CP where the suppressed convection triggered by the developing La Niña SST anomalies is located.

The RWS associated with vorticity advection by the anomalous divergent flow (Figure 3.6; upper panels) are rather similar between the transitioning and persisting summers. Therefore, the primary difference in RWS between the two cases stems from the stretching effect due to the suppressed convection in the subtropical WP caused by the decaying El Niño. In the next section, we use the stationary wave model to further examine the role of the suppressed convection in the subtropical WP in strengthening the extratropical teleconnections in the transitioning summer.
Figure 3.6 Composites of precipitation anomalies (shaded; mm/day) and 200hPa (upper) vorticity advection by anomalous divergent flow, (middle) stretching term due to anomalous divergence, and (lower) the sum of the previous two terms (contours) during the (left) transitioning and (right) persisting La Niña summers. The contour interval is $0.2 \times 10^{-10} \text{ s}^{-2}$. The zero contour is omitted for simplicity. Stippling denotes the 90% significance for RWS terms using a two-tailed Student’s t-test.
3.3.2 Stationary wave model results

3.3.2.1 Global anomalous diabatic heating

We first force the SWM with the observed anomalous diabatic heating globally from both the transitioning and persisting summers to examine ENSO summer tropical forcing of extratropical teleconnections. The composites of anomalous diabatic heating at 400hPa, where the strongest mean diabatic heating happens, (Figure 3.7) are largely similar to the anomalous rainfall patterns (Figure 3.3c and g) in the tropics. During the transitioning summer, two areas of significant anomalous cooling at 400hPa are observed over the tropical CP and subtropical WP, representing the two areas of suppressed convection. The vertical profiles of the anomalous diabatic heating also show the anomalous cooling throughout the troposphere over both the tropical CP and subtropical WP (Figure 3.7c, orange lines), indicating the suppression of these two deep convection areas. In contrast, during the persisting summer, anomalous cooling is only observed in the tropical CP, and not in the subtropical WP (Figure 3.7b and d).

Figure 3.8b and e show the model anomalous streamfunction in response to the global anomalous diabatic heating forcing (Figure 3.7) during the two developing La Niña summers. During the transitioning summer, there is a quadruple pattern of anomalous streamfunction in the tropics that resembles the Gill-Matsuno response to a tropical heat source centered off the equator (Ting and Yu 1998) and similar to the observations (Figure 3.8a). The quadruple pattern is centered at around 120°W and extends westward to reach East Asia and Australia in both the model and the observations. The pattern correlations for the anomalous streamfunction between the observations and the model response are 0.84 for the global area and 0.87 for the PNA area (0°-75°N, 120°E-60°W). This suggests that tropical diabatic forcing is able to cause anomalous circulations outside of the tropics including North America, even though the basic-state westerlies are weak in the
boreal summer. In the persisting summer (Figure 3.8e), the quadruple pattern of anomalous streamfunction is weaker in amplitude and shifted further to the east compared to the transitioning summer, though it is also similar to the observations (Figure 3.8d). Unlike in the transitioning summer, the western part of the quadruple pattern only extends to around 150°E, not reaching East Asia and Australia. The pattern correlations between the observation and the model response are 0.67 for the global area and 0.73 for the PNA area.

Tropical diabatic heating is the dominant driver of the ENSO teleconnection pattern, but the teleconnections are also influenced by midlatitude transient eddy vorticity and sensible heat fluxes (e.g. Hoerling and Ting 1994). Figure 3.8c and f show the streamfunction responses to the combination of diabatic heating and transient heat and vorticity flux convergences during the two types of La Niña summers. The primary effect of midlatitude transient eddies is to shape the details of the teleconnection patterns in the extratropics. For example, the anomalous anticyclone over the US during the transitioning summer (Figure 3.8c) becomes more distinct and like the observations in the presence of transient eddy forcing, compared to the case forced with only the diabatic heating (Figure 3.8b). Similarly, the anomalous anticyclone in North America during the persisting summer shifts northeastward and compares better with the observations (the pattern correlation in the PNA region increases from 0.73 to 0.77) when the transient eddy effects are added. These results suggest that the SWM forced with diabatic heating and transient eddy forcing has the ability to reproduce the ENSO teleconnections as well as to distinguish the difference in circulation responses between the two different developing La Niña summers.
**Figure 3.7** Composites of anomalous diabatic heating during the (a,b) transitioning and (c,d) persisting La Niña summers using NCEP-NCAR R1 data. Upper panels are the anomalous diabatic heating at 400hPa with an $0.2 \times 10^{-5}$K/s interval. Stippling denotes the 90% significance using a two-tailed Student’s t-test. Purple boxes indicate the subtropical WP and tropical CP regions used to force the SWM in Figure 3.9. Lower panels are the vertical profiles of anomalous diabatic heating over the subtropical WP (dashed) and tropical CP (solid). Black lines indicate the climatological diabatic heating over these two regions.
Figure 3.8 200hPa streamfunction anomalies from (upper) observed composites using NCEP-NCAR R1, (middle) the SWM forced with observed diabatic heating anomalies, and (lower) the SWM forced with observed diabatic heating and transient vorticity flux anomalies in the (left) transitioning and (right) persisting La Niña summers (interval: 10^6 m^2/s). Numbers in (b), (c), (e), and (f) indicate the pattern correlations with the observations (a, d) for the global and PNA (0°-75°N, 120°E-60°W) regions.
3.3.2.2 Regional anomalous diabatic heating effect

To focus on the role of diabatic cooling in the subtropical WP in the transitioning summer, we next examine the model responses to the regional diabatic heating (Figure 3.9). We force the stationary wave model with the global anomalous diabatic heating and the transient vorticity forcing over (1) both the subtropical WP and tropical CP (EXP-WP+CP, Figure 3.9a,d), (2) the tropical CP (EXP-CP, Figure 3.9b,e), and (3) the subtropical WP (EXP-WP, Figure 3.9c,f) for both the transitioning and persisting summers.

In the transitioning summer (denoted as EXP\(_T\)), the diabatic cooling over the subtropical WP and the tropical CP dominate the anomalous circulations. The anomalous circulations from EXP\(_T\)-WP+CP (Figure 3.9a) are highly similar to the anomalous circulations forced by the global diabatic heating field (Figure 3.8c) with a pattern correlation of 0.90 for the global domain and 0.96 for the PNA region. The streamfunction pattern in Figure 3.9a also resembles the observations shown in Figure 3.8a, with a pattern correlation of 0.81 for the global domain and 0.84 for the PNA region. When only the tropical CP diabatic cooling is prescribed to force the model (Figure 3.9b), the quadruple pattern of anomalous streamfunction is much weaker in amplitude and does not extend as far to the west as in Figure 3.9a when both the WP and CP diabatic cooling are included. This is also reflected in the spatial pattern correlation with the anomalous circulations forced by the global diabatic heating (Figure 3.8c), which drops to 0.69 for the global domain and 0.65 for the PNA region. The intensity of the extratropical teleconnections is weakened, but an anomalous anticyclone is still found over North America, consistent with the classic wave-train in response to the La Niña tropical forcing.

On the other hand, when only the subtropical WP diabatic cooling is applied to the model, the quadruple pattern shifts westward with the center near the dateline (Figure 3.9c), suggesting
that the WP diabatic cooling contributes to the westward extension of the tropical response associated with the La Niña tropical CP forcing. Furthermore, the subtropical WP diabatic cooling also contributes to the anomalous anticyclone over North America with a similar amplitude as that due to the tropical CP cooling (Figure 3.9b). The pattern correlations with the anomalous circulations forced by the global diabatic heating (Figure 3.8c) are 0.61 for the global domain and 0.68 for the PNA region, comparable to the ones in EXP\(_T\)-CP, justifying the important role played by the subtropical WP cooling in the overall teleconnection in the transitioning La Niña summer. These results support our hypothesis that the suppressed convection over the subtropical WP can trigger stationary wave propagation towards extratropical North America and strengthening the ENSO extratropical teleconnections during the transitioning summer.

In the persisting summer (denoted as EXP\(_P\)), the major difference, compared to in the transitioning summer, is that the anomalous circulations in EXP\(_P\)-CP (Figure 3.9e) are similar to the ones in EXP\(_P\)-WP+CP (Figure 3.9d). The quadruple patterns in these two experiments are both similar to the anomalous circulations forced by the global diabatic heating (Figure 3.8f) as well as the observations (Figure 3.8d) with the center around 120°W and extending westward to around 150°E. This implies that the diabatic heating over the subtropical WP is not influential in this case. Figure 3.9f shows the anomalous circulations from EXP\(_P\)-WP. This shows no similarity with the observations (pattern correlation is 0.07 for the global domain and 0.01 for the PNA region). Hence, in the persisting summer, diabatic cooling over the tropical CP dominates the ENSO teleconnection patterns, unlike during the transitioning summer when diabatic coolings over both the tropical CP and the subtropical WP play substantial roles.
Figure 3.9 200hPa streamfunction anomalies from the SWM forced with regional observed diabatic heating from (upper) both the subtropical WP and the tropical CP, (middle) the tropical CP, and (lower) the subtropical WP together with global transient vorticity flux anomalies in the (left) transitioning and (right) persisting La Niña summers (interval: \(10^6 \, m^2/s\)). Dashed (solid) lines indicate the area where diabatic heating anomalies are smaller than \(-0.4 \times 10^5 \, K/s\) (larger than \(0.4 \times 10^5 \, K/s\)). Numbers indicate the pattern correlations with the observations (Figure 3.8a and d, denoted as OBS) and the streamfunction anomalies in response to global diabatic heating anomalies in the SWM (Figure 3.8c and f, denoted as SWM) for the global and PNA regions. The area of regional diabatic heating anomalies are indicated in Figure 3.7a and c.
Therefore, the SWM experiments confirm the hypothesis that the suppressed convection over the subtropical WP in the transitioning summer is the reason why the teleconnections differ from those in the persisting summer. Furthermore, the WP suppressed convection contributes substantially to the strong anomalous anticyclone over North America which is closely linked to the robust warming signal over the Midwest. Although the tropical diabatic heating plays a major role in shaping the extratropical teleconnections, the mid-latitude transient eddies also contribute to the details of the extratropical teleconnections and, particularly, the anomalous anticyclone over North America that subsequently impacts the surface climate in North America.

3.4 Conclusions and Discussions

3.4.1 Conclusions

Here we have examined the physical mechanisms of teleconnections in developing La Niña summers when ENSO tropical forcing negatively affects soybean and maize yields in the US. Since 1950, a developing La Niña summer is either when an El Niño is transitioning to a La Niña (transitioning summer) or a La Niña is persisting (persisting summer). We have focused on distinguishing the dynamics of these two developing La Niña summers based on observations and using a stationary wave model as a diagnostic tool.

- According to the observations, transitioning and persisting summers have different SST anomaly patterns across the tropics since they evolved differently from the preceding winters. During the transitioning summer, although the tropical Pacific has transitioned into La Niña state, the Indian Ocean and the tropical Atlantic are still in the El Niño decaying phase. In contrast, during the persisting summer, the La Niña signal alone spans the tropics.
- Different oceanic anomalies lead to different atmospheric responses. During the transitioning summer, two suppressed deep convection areas dominate the anomalous rainfall field over the tropical Pacific: one is over the central Pacific (CP) due to the developing La Niña, and another one over the western Pacific (WP) due to the decaying El Niño. On the other hand, during the persisting summer, only the suppressed deep convection induced by the La Niña SST forcing is present over the tropical CP.

- During the transitioning summer, the suppressed convection over the tropical CP and the subtropical WP both provide anomalous vorticity sources via the stretching effect and subsequently induce Rossby wave propagation extending to North America. These two wave-trains superimpose on each other, leading to statistically significant teleconnections in the extratropics with a significant anomalous anticyclone over northeastern North America and subsequently a robust warming over the Midwest. In contrast, during the persisting summer, without the augmentation by a wave-train from the subtropical WP, the teleconnection is weak and only statistically significant in the tropics with no significant temperature anomalies over the US.

- The contribution of the suppressed convection over the subtropical WP to the extratropical teleconnections during the transitioning summer is confirmed by the stationary wave model (SWM) which is able to well-reproduce the observed anomalous circulation when forced by global diabatic heating in the developing La Niña summers. According to the SWM experiments, the diabatic cooling over the subtropical WP contributes substantially to the intensities of the extratropical teleconnections and the anomalous anticyclone over North America, comparable to the contribution from the diabatic cooling over the tropical CP. On the contrary, during the persisting summer, diabatic cooling over the tropical CP
dominates the ENSO teleconnection pattern. Anomalous diabatic heating over the subtropical WP is not influential in this case.

- Therefore, the suppressed convection over the subtropical WP in the transitioning summer distinguishes the teleconnections from those in the persisting summer. This El Niño-induced WP suppressed convection and the associated Rossby wave strengthen the extratropical teleconnection induced by the developing La Niña SST forcing, leading to a strong anomalous anticyclone and robust warm signals over the Midwest during the transitioning summer.

3.4.2 Discussions

Although the model experiments decently reproduced the observations in many aspects, the observed difference in the intensity of anomalous anticyclone between transitioning and persisting summers (Figure 3.8a and d) is much larger than in the SWM results (Figure 3.8c and f). A plausible explanation for this discrepancy is that the intensity of the anomalous anticyclone in the observations is also affected by several other factors not included in the SWM. These possible factors include:

- Land-atmosphere feedback is strong in the summer and its influence on circulation is comparable to that of remote SST forcing according to some previous studies (e.g., Koster et al. 2000; Douville 2010). This could amplify the impacts on atmospheric circulations of tropical surface temperatures. This land-atmosphere feedback is not included in the SWM.

- Atmospheric internal variability could contribute appreciably to the amplitude of the observed anomalies, augmenting the forced response (e.g., Hoerling and Kumar 1997; Chen and Kumar 2018; Jong et al. 2018; Chapter 2).
The transient eddy flux anomalies are caused by changes in the mid-latitude mean flow, but also feedback on the mid-latitude mean flow. However, this eddy-mean flow interaction is not allowed in the model as transient eddies are treated as forcing and this could lead to errors in amplitude of the forced response.

To summarize, the different oceanic states of different La Niña summers result in different atmospheric convection and circulation anomalies. Hence, it is necessary to separately consider the transitioning and persisting La Niña events as their teleconnections and, therefore, impacts on crop yields are significantly different. This demonstrates that improved understanding of ENSO summer teleconnections and seasonal prediction of US summertime hydroclimate will require further study of the seasonal evolution of ENSO characteristics within a multi-year ENSO lifecycle.
Chapter 4. ENSO summer teleconnections and impacts on North America as simulated in NCAR CAM5

4.1 Introduction

In Chapter 3, we showed that ENSO summer teleconnections can impose significant impacts on North American surface climate even though both ENSO tropical forcing and basic-state westerlies are weak in boreal summer. However, due to the different prior ENSO conditions, the physical mechanisms of teleconnections and characteristics of remote impacts on the US during different ENSO summers can be appreciably different. Two different types of developing La Niña summers, the one transitioning from a peak El Niño winter and another persisting from a prior La Niña winter, are found to exhibit substantial differences in their impacts on North American summer climate.

During the transitioning summer, the decaying El Niño and the developing La Niña trigger suppressed deep convection over the subtropical western Pacific (WP) and the tropical central Pacific (CP), respectively. Both regions of suppressed convection trigger Rossby wave propagation across the Pacific-North America (PNA) region, resulting in statistically significant extratropical teleconnections and an anomalous anticyclone over eastern North America which leads to a robust warming over the Midwest. On the other hand, during the persisting summer, only one region of suppressed convection is present and is over the tropical CP forced by the developing La Niña. The teleconnection is correspondingly weak and only statistically significant in the tropics. We further demonstrated the role of the suppressed convection over the subtropical WP in the transitioning summer in distinguishing the teleconnections from those in the persisting summer using a stationary wave model. The stationary wave model results suggested that the
wave-train from the subtropical WP augments that from the central Pacific and is crucial to the extratropical teleconnections and the strong anomalous anticyclone over North America during the transitioning summer.

Current climate models have relatively poor forecasting skill for Northern Hemisphere summer extratropical circulation, in sharp contrast to the demonstrated skill for boreal winter ENSO-based seasonal climate forecasts (e.g. Wang et al. 2009; Ding et al. 2011). It is therefore interesting to examine the performance of an Atmospheric Global Circulation Model (AGCM) forced by historical SST in terms of ENSO summer teleconnections and whether the model is capable of distinguishing the different characteristics of transitioning and persisting La Niña summers as shown in observations in Chapter 3.

### 4.2 Model data and Method

In this chapter, we use the National Center for Atmospheric Research (NCAR) Community Atmospheric Model, version 5 (CAM5, Neale et al. 2012). We use the 16-member Global Ocean Global Atmosphere (GOGA) experiments where monthly historical SST observations and sea ice from the UK Met Office’s Hadley Centre (Rayner et al. 2003) are prescribed over the global ocean for the period 1856 to 2016. Each ensemble member differs only in their initial atmospheric state. To be consistent with observations (Chapter 3.2.1), we use the model output from January 1950 to February 2015 in this chapter. The monthly climatology is based on averages from January 1950 to December 2014. The surface temperature over land is computed by the model and, for analysis, is linearly de-trended with the trend removed for each 3-month season separately, consistent with the observational analyses.
The selection of El Niño and La Niña events is the same as in Chapter 3 but using the SST anomalies from the Hadley Center (Chapter 3.2.2). We use the Oceanic Niño Index (ONI) as the criteria. ONI is a 3-month running mean of SST anomalies in the Niño3.4 region (5°N-5°S, 170°-120°W), relative to a 30-year climatology. The 30-year base period is updated every 5 years and centered at the first year of these 5-year-increments. El Niño and La Niña events are defined when the October-November-December ONI reaches the threshold of +0.5 °C and -0.5 °C (corresponding to about half standard deviation), respectively.

Based on these criteria, we identified 4 single-year La Niña events from 1950 to 2014 (1964, 1988, 1995, 2005), 5 two-year La Niña events (1954-55, 1970-71, 1983-84, 2007-08, 2010-11), and 2 three-year La Niña events (1973-75, 1998-2000). Therefore, there are 11 first-year La Niña winters. The preceding winters of these first-year La Niña were all identified as El Niño winters. We categorize the summers in the first-year La Niña developing phase as “transitioning summer” (denoted as JJA(0)T in all the figures). On the other hand, there are 7 second-year La Niña winters and 2 third-year La Niña winters, a total of 9 persisting La Niña summers. We categorize the summers prior to these La Niña winters as “persisting summer” (denoted as JJA(0)p).

4.3 Results

4.3.1 Comparison of CAM5-GOGA ensemble mean and NCEP-NCAR Reanalysis

We first examine the CAM5-GOGA ensemble mean evolutions of atmospheric conditions from the preceding winters to the developing La Niña summers (Figure 4.1 and Figure 4.2). For the transitioning La Niña, the anomalous tropical precipitation shows the evolution from the El Niño state (Figure 4.1a) to a La Niña state (Figure 4.1c). During the preceding El Niño winter, the model ensemble mean well reproduces both the observed anomalous tropical precipitation and
teleconnection patterns (compare Figure 3.3a to Figure 4.1a). The pattern correlation for precipitation over the tropical Indo-Pacific region (30°S-30°N, 60°E-60°W) is 0.77 between the model ensemble mean and National Centers for Environmental Prediction–National Center for Atmospheric Research Reanalysis 1 (NCEP-NCAR R1 hereafter; see Table 4.1), while the pattern correlation for the 200hPa geopotential height anomalies over the PNA region (0°-60°N, 120°E-60°W, PNA Z200 hereafter) is as high as 0.95 (Table 4.1). The classic warm north–cold south dipole over North America during the observed El Niño winters (Figure 3.4a) is also well reproduced by the model ensemble mean (Figure 4.2a). These results illustrate that the NCAR CAM5 is skillful in simulating the El Niño winter PNA teleconnections.

The similarity between the model ensemble mean and NCEP-NCAR R1, however, decreases as ENSO evolves from winter to summer (Table 4.1, top). By the transitioning summer JJA(0), the pattern correlations for PNA Z200 and tropical Indo-Pacific precipitation drop to 0.59 and 0.57, respectively (Table 4.1). Over the tropical Pacific, only the suppressed convection over the tropical CP is present in the model ensemble mean, leading to a single wave-train propagating from the tropical CP across the PNA region and reach North America (Figure 4.1c). The suppressed convection over the subtropical WP, a significant feature in the observations (Figure 3.3c), is not present in the model ensemble mean. Furthermore, the extratropical teleconnection and the anomalous anticyclone over northeastern North America are much weaker compared to the ones in NCEP-NCAR R1 (Figure 3.3c). The discrepancy in the anomalous circulation between the model ensemble mean and observations is also reflected in the surface temperature over the US (Figure 4.2c compared to Figure 3.4c): only weak warm anomalies are present in the northeastern US, unlike in the observations where significant warm anomalies cover most of the area east of the Rocky Mountains. Therefore, the model ensemble mean poorly reproduces the
transitional La Niña summer teleconnections, even with the successful simulation during the preceding El Niño winter.

Table 4.1 Pattern correlations between CAM5-GOGA and NCEP-NCAR R1 composites for 200hPa geopotential height anomalies over the PNA region (0°-60°N, 120°E-60°W, PNA Z200) and precipitation anomalies over the tropical Indo-Pacific (30°S-30°N, 60°E-60°W), Indo-western Pacific (30°S-30°N, 60°E-180°), and central-eastern Pacific (30°S-30°N, 180°-60°W) regions for the transitioning La Niña summers from the preceding winters D(-1)JF(0), the preceding springs MAM(0) to the developing La Niña summers JJA(0)T. The CAM5-GOGA composites are constructed based on (top) ensemble mean, ensemble member which has the (middle) highest and (bottom) lowest PNA Z200 pattern correlation with NCEP-NCAR R1 among all the members during each season.

<table>
<thead>
<tr>
<th>Composite</th>
<th>Season</th>
<th>PNA Z200</th>
<th>Indo-Pacific Precip</th>
<th>Indo-WP Precip</th>
<th>CP-EP Precip</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ensemble mean</td>
<td>D(-1)JF(0)</td>
<td>0.95</td>
<td>0.77</td>
<td>0.63</td>
<td>0.87</td>
</tr>
<tr>
<td></td>
<td>MAM(0)</td>
<td>0.79</td>
<td>0.51</td>
<td>0.50</td>
<td>0.54</td>
</tr>
<tr>
<td></td>
<td>JJA(0)T</td>
<td>0.59</td>
<td>0.57</td>
<td>0.38</td>
<td>0.71</td>
</tr>
<tr>
<td>Best Ensemble</td>
<td>D(-1)JF(0)</td>
<td>0.97</td>
<td>0.79</td>
<td>0.67</td>
<td>0.87</td>
</tr>
<tr>
<td></td>
<td>MAM(0)</td>
<td>0.87</td>
<td>0.32</td>
<td>0.28</td>
<td>0.49</td>
</tr>
<tr>
<td></td>
<td>JJA(0)T</td>
<td>0.77</td>
<td>0.63</td>
<td>0.49</td>
<td>0.74</td>
</tr>
<tr>
<td>Worst Ensemble</td>
<td>D(-1)JF(0)</td>
<td>0.11</td>
<td>0.71</td>
<td>0.52</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>MAM(0)</td>
<td>-0.03</td>
<td>0.49</td>
<td>0.50</td>
<td>0.49</td>
</tr>
<tr>
<td></td>
<td>JJA(0)T</td>
<td>-0.19</td>
<td>0.36</td>
<td>-0.05</td>
<td>0.68</td>
</tr>
</tbody>
</table>
Figure 4.1 Composites of CAM5-GOGA ensemble mean precipitation anomalies (shaded, mm/day) and 200hPa geopotential height anomalies with the zonal-mean removed (contours, interval: 5m) for the (left) transitioning and (right) persisting La Niña summers from (a,d) the preceding winters D(-1)JF(0), (b,e) the preceding springs MAM(0) to (c,f) the developing La Niña summers JJA(0). Stippling denotes the 90% significance for precipitation anomalies using a two-tailed Student’s t-test. Thick lines indicate the 90% significance for 200hPa height variations.
Figure 4.2 As in Figure 4.1, but for composites of de-trended surface temperature. Stippling denotes the 90% significance for de-trended surface temperature anomalies using a two-tailed Student’s t-test.
The decrease in pattern correlations for circulation and tropical precipitation from ENSO winter to following summer also happens in the persisting La Niña (Table 4.2, top). For the persisting La Niña (Figure 4.1d-f), the anomalous tropical precipitation shows a slow decay from the preceding La Niña winter through the spring to the persisting La Niña summer, similar to the observations (Figure 3.3e-g). During the preceding La Niña winter, the pattern correlations for the PNA Z200 and tropical Indo-Pacific precipitation are 0.72 and 0.73, respectively, between the model ensemble mean and NCEP-NCAR R1. The pattern correlation for the PNA Z200 is not as high as in the one during the El Niño winter (0.95). This is also reflected in the impacts on the US surface temperature. Compared to the observations (Figure 3.4e), the model ensemble mean reproduces the warm anomalies over the Southeast in the winter (Figure 4.2d), but it simulates a strong cooling in the northern US which is not present in the observation. Hence, the model ensemble mean can reproduce the La Niña winter teleconnections with some skills, albeit not as high as for the El Niño winter.

Table 4.2 As in Table 4.1, but for the persisting La Niña summers.

<table>
<thead>
<tr>
<th>Composite</th>
<th>Season</th>
<th>PNA Z200</th>
<th>Indo-Pacific Precip</th>
<th>Indo-WP Precip</th>
<th>CP-EP Precip</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ensemble mean</td>
<td>$D(-1)JF(0)$</td>
<td>0.72</td>
<td>0.73</td>
<td>0.66</td>
<td>0.78</td>
</tr>
<tr>
<td></td>
<td>$MAM(0)$</td>
<td>0.22</td>
<td>0.65</td>
<td>0.63</td>
<td>0.67</td>
</tr>
<tr>
<td></td>
<td>$JJA(0)_{P}$</td>
<td>0.57</td>
<td>0.49</td>
<td>0.42</td>
<td>0.57</td>
</tr>
<tr>
<td>Best Ensemble</td>
<td>$D(-1)JF(0)$</td>
<td>0.76</td>
<td>0.72</td>
<td>0.58</td>
<td>0.80</td>
</tr>
<tr>
<td></td>
<td>$MAM(0)$</td>
<td>0.48</td>
<td>0.60</td>
<td>0.57</td>
<td>0.61</td>
</tr>
<tr>
<td></td>
<td>$JJA(0)_{P}$</td>
<td>0.69</td>
<td>0.48</td>
<td>0.40</td>
<td>0.56</td>
</tr>
<tr>
<td>Worst Ensemble</td>
<td>$D(-1)JF(0)$</td>
<td>0.57</td>
<td>0.66</td>
<td>0.62</td>
<td>0.70</td>
</tr>
<tr>
<td></td>
<td>$MAM(0)$</td>
<td>-0.09</td>
<td>0.58</td>
<td>0.56</td>
<td>0.59</td>
</tr>
<tr>
<td></td>
<td>$JJA(0)_{P}$</td>
<td>-0.22</td>
<td>0.48</td>
<td>0.39</td>
<td>0.59</td>
</tr>
</tbody>
</table>
The AGCM’s ability to simulate the tropical precipitation and circulation in the model ensemble mean drops substantially for the persisting summer JJA(0). The pattern correlation for the PNA Z200 is only 0.57 (Table 4.2), and even then it is mostly statistically significant in the tropics where it is most similar to the NCEP-NCAR R1 PNA Z200 (Figure 3.3g). This decrease in the similarity between the model and observations also appears in the US surface temperature (Figure 4.2f). In the model ensemble mean, there are significant warm anomalies over the central US, which is not so in the observations (Figure 3.4g).

These comparisons between the CAM5-GOGA ensemble mean and NCEP-NCAR R1 demonstrate that the model forced with historical SSTs can well reproduce ENSO winter features, especially during El Niño winters, but has relatively poor performance in simulating La Niña summer teleconnections. There exists, however, the possibility that atmospheric internal variability can overwhelm the summer SST-forced circulation in summer (which is weaker than in winter) and associated surface climate features. Therefore, a wide range of variability across the ensemble members could lead to a weak response in the ensemble mean, and a low signal-to-noise ratio for both the atmospheric circulation and tropical western Pacific precipitation in the summer season, compared to the winter season (Figure 4.3).

4.3.2 Examination of the variability in atmospheric circulation and tropical precipitation in CAM5-GOGA

To further examine whether the model has the ability to correctly simulate the observed ENSO atmospheric responses during La Niña summers, among the 16 ensemble members we select the best ensemble member for each La Niña summer that has the highest pattern correlation for the PNA Z200 with NCEP-NCAR R1 to construct a new ENSO composite (Figure 4.4 and
Figure 4.3 The ratio of signal to noise of (upper) 200hPa geopotential height anomalies and (lower) precipitation anomalies during (left) December-January-February (DJF) and (right) June-July-August (JJA). The signal is calculated from the variance of the ensemble mean; while the noise is the variance of all ensemble member.

Figure 4.5; Table 4.1 and Table 4.2, middle). For the transitioning summer, the composite of anomalous circulation based on the best ensemble members is more similar to the NCEP-NCAR R1 (Figure 4.4a, pattern correlation is 0.77), compared to the one based on all ensemble members (Figure 4.1c, pattern correlation is 0.59). The extratropical teleconnections and the anomalous anticyclone over northeastern North America are stronger and more distinct, leading to a stronger and southward-expanded warming signal over the Midwest (Figure 4.5a). The improved performance is also present in the persisting La Niña summers (Figure 4.4c). The pattern correlation for the PNA Z200 increases from 0.57 using the all ensemble mean to 0.69 in the best ensemble composite (Table 4.2). This result indicates that the model forced with historical SST is able to simulate a circulation pattern that is similar to that in the observations in the La Niña.
summers. Hence, the weak response in the ensemble mean is the result of the SST-forced signal being small compared to the large spread of internal variability across the ensemble members.

The internal variability could originate from pure atmospheric internal variability or from variability in tropical precipitation. To identify the major source of the variability, we first compare the composites based on best ensembles (Figure 4.4, upper panels; Table 4.1 and Table 4.2, middle) and worst ensembles (Figure 4.4 lower panels; Table 4.1 and Table 4.2, bottom) in the preceding winter and developing La Niña summers. For the transitioning La Niña, during the preceding winter, the pattern correlation for the PNA Z200 drops from 0.97 using the best ensemble members to 0.11 using the worst ensemble members (Table 4.1). However, the change in the pattern correlation for tropical precipitation is small (0.79 to 0.71). Therefore, this indicates that atmospheric internal variability dominates the spread in the anomalous circulations across all the ensemble members during the El Niño winters. On the other hand, during the transitioning summer JJA(0)_T, the pattern correlations for the PNA Z200 and tropical Indo-Pacific precipitation both drop when using the worst ensemble members (Table 4.1, bottom). In particular, the variability in tropical precipitation mainly originates from the tropical Indo-western Pacific: the pattern correlation drops from 0.49 using the best ensembles to -0.05 using the worst ensembles, while the pattern correlations for the central to eastern Pacific precipitation are similar in both cases (0.74 versus 0.68). Hence, in the transitioning summer, both the PNA Z200 and tropical Indo-western Pacific convection show large variability across ensemble members.
Figure 4.4 Composites of precipitation anomalies (shaded, mm/day) and 200hPa geopotential height anomalies with the zonal-mean removed (contours, interval: 5m) during the (left) transitioning and (right) persisting La Niña summers. For each La Niña summer, only the ensemble member, whose 200hPa geopotential height anomalies over the PNA region has (top) the highest and (bottom) the lowest pattern correlation with NCEP-NCAR R1, is selected to do the composite. Stippling denotes the 90% significance for precipitation anomalies using a two-tailed Student’s t-test. Thick lines indicate the 90% significance for 200hPa height variations.
Figure 4.5 As in Figure 4.4, but for composites of de-trended surface temperature. Stippling denotes the 90% significance for de-trended surface temperature anomalies using a two-tailed Student’s t-test.

Since in Chapter 3 we demonstrated the importance of anomalous precipitation in the WP to the extratropical teleconnections during the transitioning summer, the relation of convection in this area and the PNA Z200 deserves more attention. Therefore, we compare the pattern correlation for the PNA Z200 in the transitioning summer in the three cases: ensemble mean composite (0.59), best ensemble composite (0.77), and worst ensemble composite (-0.19). The pattern correlation for the PNA Z200 varies consistently with the pattern correlation for the tropical Indo-western Pacific precipitation (0.38 for the ensemble mean, 0.49 for the best ensemble, and -0.05 for the worst ensemble). This relationship is not present in the persisting La Niña summer in which the WP does not have significant anomalous precipitation and plays no role in the extratropical teleconnections based on the observations. Instead, the pattern correlation for the tropical Indo-
western Pacific precipitation is similar across the three cases (Table 4.2, 0.42, 0.40, and 0.39). Therefore, these results reinforce the argument about the importance of precipitation anomalies over the WP to the teleconnections during the transitioning summer. Moreover, the results from CAM5-GOGA suggest that the strong variability in the PNA Z200 in the transitioning summer might be largely contributed by the variability in the tropical Indo-western Pacific convection.

4.4 Conclusions and Discussions

4.4.1 Conclusions

Here we have examined the performance of the NCAR CAM5 forced with historical SST for ENSO summer teleconnections. In particular, we have focused on two types of developing La Niña summers: transitioning and persisting summers. These have distinct characteristics of teleconnections and impacts on North America in the observations. We aimed to examine whether CAM5-GOGA is able to reproduce these distinct features of the transitioning and persisting summers. The main findings are summarized below:

- The model ensemble mean has limited skill in simulating the tropical convection and teleconnections during both the transitioning and persisting summers, even though the model ensemble mean well reproduces the features in the preceding ENSO winters. In particular, during the transitioning summers, the suppressed convection in the subtropical WP, important to the extratropical teleconnections in the observations, is not present in the model ensemble mean. Also, the strong anomalous anticyclone and the robust warming signal over the Midwest U.S. are not reproduced in the model ensemble mean.
- The large variability across the ensemble members contributes to the weak response in the model ensemble mean. During the transitioning summers, large variability is present in
both precipitation in the tropical Indo-western Pacific and 200hPa geopotential height over the PNA region. Moreover, the variability of 200hPa geopotential height over the PNA region varies with the variability of precipitation over the tropical Indo-western Pacific. In contrast, during the persisting summer, while 200hPa geopotential height over the PNA region also has large variability, precipitation over the tropical Indo-western Pacific shows little variability across all the ensemble members.

- Therefore, the presence of the WP precipitation is critical to reasonably simulating the extratropical teleconnection and the anomalous anticyclone over North America during the transitioning summer in the model. These results suggest the importance of WP precipitation to the extratropical teleconnections during the transitioning La Niña summer, supporting our results based on observations and the stationary wave model presented in Chapter 3.

4.4.2 Discussions

The analyses based on CAM5 simulations in this chapter also raise some interesting questions that we are not able to fully address. These are listed as follows:

- According to previous studies (e.g., Xie et al. 2009), the drier-than-normal subtropical WP is likely caused by the baroclinic Kelvin wave forced by enhanced precipitation over the warm Indian Ocean during the early summer when an El Niño is decaying. In other words, the WP suppressed convection is a response to the warm SST anomalies over the India Ocean. However, when CAM5 was forced by historical SST, the precipitation response over the tropical Indo-western Pacific shows large variability across the ensemble members during the transitioning summer with little SST-forced signal in the ensemble mean. This
raises the question whether the dry anomalies over the WP during summer when the El Niño is decaying or transitioning to La Niña are driven by the SST anomalies over the Indian Ocean. Further AGCM experiments and other analyses are needed to address this question.

- Table 4.1 and Table 4.2 show that the CAM5-GOGA ensemble mean has good skill in reproducing the ENSO winter anomalous circulation. However, the performance for El Niño winters (pattern correlation for the PNA Z200 is 0.95) is substantially better than for La Niña winters (pattern correlation for the PNA Z200 is 0.72). It is interesting to further ask whether this asymmetric performance in El Niño and La Niña winters is a general feature across all climate models and if so, what are the possible physical reasons behind this asymmetry.

- The El Niño and La Niña events have asymmetric duration: an El Niño tends to decay rapidly in the boreal spring, while a La Niña tends to decay slowly and persist through the following summer. Therefore, the amplitude of SST anomalies over the tropical Pacific is larger in the persisting spring (Figure 3.2f) compared to that during the El Niño decaying spring (Figure 3.2b). However, the CAM5-GOGA ensemble mean shows good performance in the PNA Z200 during the decaying spring of El Niño (Table 4.1, pattern correlation is 0.79), while for the persisting La Niña spring, the performance is even poorer than in the persisting summer (Table 4.2, pattern correlation is 0.22). What causes the poor performance of the model in reproducing the persisting La Niña spring feature is also an interesting question.

In summary, a weak SST forced signal-to-noise ratio in the boreal summer leads to large variability in both the tropical precipitation, especially over the western Pacific, and atmospheric
circulation (Figure 4.3), which challenges the current seasonal forecast models. Although the NCAR CAM5 forced with historical SST can reasonably simulate El Niño winter and spring teleconnections, the model has limited skill in other seasons during the multi-year ENSO life-cycle. More studies are needed to better understand the physical processes, both oceanic and atmospheric, that determine the seasonal forecast skills for the summer season, and thereby improve the forecast skill of ENSO-based climate model prediction skills in all seasons.
Conclusion

ENSO has far-reaching impacts on seasonal climate and weather anomalies across the globe and provides the most reliable source for seasonal hydroclimate forecast skill in many regions, including North America. Tremendous progress has been made in the past three decades in understanding the impacts of ENSO on seasonal climate over North America, especially during the boreal winter season. However, among these previous studies, relatively few examined the seasonal evolutions of ENSO teleconnections and their impacts across the winter half-year. Also, previous studies on ENSO’s impacts on North America disproportionately focused on the boreal winter season, focusing less on the impacts during the boreal summer in which crop yields are sensitive to even small variations in temperature and precipitation. Therefore, we built on existing literature to address these knowledge gaps and establish the seasonal dependence of ENSO teleconnections as well as impacts on the surface climate over North America. We focused on two examples: the impacts of El Niño on California winter precipitation (Chapters 1 and 2) and the impacts of developing La Niña on Midwest summer climate (Chapters 3 and 4).

The impacts of El Niño on California winter precipitation (Chapters 1 and 2)

California is one of the largest economies in the world and a primary agricultural production state in the US. It receives most of its precipitation during the winter months, from November to March. To improve the seasonal forecast skill, numerous studies have established that El Niño tends to bring excessive amount of winter precipitation to California. However, none of these studies examined the change in El Niño’s impacts throughout the winter half-year. Also, the role of El Niño intensity had not been fully examined and well-documented. Therefore, in Chapter 1, the dependence of the El Niño – California precipitation relationship on the timing,
strength of El Niño SST anomalies, and the regionality was examined based on 110-years of observations. This chapter reached the following conclusions:

- The influences of El Niño on California precipitation are statistically significant in late winter (Feb-Apr, FMA), but not in early winter (Nov-Jan) even though that is when El Niño usually reaches its peak intensity.
- A moderate-to-strong El Niño is more likely to cause a statewide wet winter in California, compared to a weak El Niño.
- The relationship between El Niño and precipitation is stronger in relatively dry southern California.
- The delayed teleconnection response in late winter is largely caused by a stronger and more eastward extended tropical diabatic heating in late winter due to a warmer climatological SST over the tropical eastern Pacific.

The novel aspect of this study is that we specifically identify that the strongest impacts of El Niño on California occur in late winter, after the peak of tropical SST anomalies. Together with other details of the California precipitation dependence on El Niño intensity and region, this study provides information that can further benefit seasonal forecasts of the timing and location of anomalous precipitation in California.

California, however, was surprisingly dry during the 2015/16 strong El Niño winter, which led to the work in Chapter 2. We synthesized observations, NMME forecasts, and AGCM experiments to examine one of the possible reasons why this strong El Niño did not bring expected precipitation to California in the late winter. The following summarizes the main findings in Chapter 2:
Through comparison of the 15/16 event to the two past strong El Niño events (82/83 and 97/98), it appears that the maximum equatorial Pacific SSTAs were located further west during the 15/16 late winter compared to those during the 82/83 and 97/98 late winters, both of which brought extremely wet late winters to California.

The NMME forecast models overestimated both the Niño3 SSTAs and California precipitation in the late winter, providing additional support for the idea that a westward shifted SSTA was responsible for the lack of a wet California precipitation anomaly during the strong 15/16 El Niño.

To test this hypothesis, AGCM experiments were conducted by prescribing observed FMA 2016 SSTA and NMME forecast FMA 2016 SSTA in three NCAR climate models. While the model results indicate a role for the SST bias in NMME contributing to the forecast wet bias in California precipitation, especially in northern California, it is not strong enough to explain what actually happened in the 15/16 winter over California.

Nevertheless, the observed California precipitation anomalies in the 15/16 winter were within the ensemble spread of the model experiments with prescribed SSTA from both observations and the NMME model forecast. Therefore, atmospheric internal variability could have played a considerable role in offsetting the SST-forced signal and generating the dry California winter during the 15/16 strong El Niño event.

The 2015/16 El Niño led to a number of studies focusing on the different possibilities for what caused the surprisingly dry winter in California during this event. Among these studies, Chapter 2 of this thesis uniquely synthesized both the physical mechanisms and possible errors from model forecasts, focusing on the role of SSTA spatial pattern over the tropical Pacific. This work not only helps us better understand the physical mechanism of ENSO regional
teleconnections, but also diagnoses the role of SST forecast errors in operational seasonal forecasting models.

As the regional climate response to ENSO forcing is highly sensitive to subtle differences in the teleconnection patterns, detailed studies of the relationship between regional climate and teleconnections, such as the precise timing of anomalous precipitation, are essential to improve seasonal forecast skill. The work in these two chapters aimed to complement the knowledge gaps in terms of the seasonal evolving ENSO winter teleconnections and impacts on California precipitation.

*The impacts of developing La Niñas on Midwest summer climate (Chapters 3 and 4)*

ENSO also exerts significant impacts on agricultural production over the Midwest during the boreal summer season. One example is that the soybean and maize yields in the US drop significantly during the summer when a La Niña is developing. While there are strong impacts on agricultural production, the physical process underlying the summer teleconnections has not been well established. In Chapter 3, we specifically examined the different physical mechanisms of ENSO teleconnections and their impacts on US summertime temperature during multi-year La Niña life-cycles based on observations and a stationary wave model. This chapter reached the following conclusions:

- A developing La Niña summer is either when an El Niño is transitioning to a La Niña or a La Niña is persisting. The oceanic and atmospheric characteristics during these two developing La Niña summers are distinct.

- During the transitioning summer, two suppressed deep convection areas dominate the anomalous rainfall field over the tropical Pacific: one is over the central Pacific due to the
developing La Niña, and another one is over the western Pacific due to the decaying El Niño. On the other hand, during the persisting summer, only the suppressed deep convection induced by the La Niña SST forcing is present over the tropical CP.

- During the transitioning summer, the suppressed convection over the tropical CP and the subtropical WP both induce Rossby wave propagation extending to North America. These two wave-trains superimpose on each other, leading to statistically significant teleconnections in the extratropics with a significant anomalous anticyclone over northeastern North America and subsequently a robust warming over the Midwest. In contrast, during the persisting summer, without the augmentation by a wave-train from the subtropical WP, the teleconnection is weak and only statistically significant in the tropics with no significant temperature anomalies over the US.

The research in this chapter demonstrates the significant impacts on North American summertime temperature from ENSO teleconnections, especially during the developing La Niña summer. In particular, it is the first one to indicate that, from the point of view of teleconnections and regional climate impacts, there are two different developing La Niña summers, transitioning and persisting. Therefore, it is necessary to separately consider the transitioning and persisting La Niña events because their teleconnections and impacts on crop yields are significantly different. This work also emphasizes that it is important to consider the seasonal evolution within a multi-year ENSO life-cycle when addressing the seasonal impacts on North America.

In Chapter 4, we further examined the NCAR CAM5 forced with historical SST to see whether the model is capable of distinguishing the different observed characteristics in transitioning and persisting La Niña summers as shown in Chapter 3. The following summarizes the main results in this chapter:
• The model ensemble mean has limited skill in simulating the tropical convection and teleconnections during both the transitioning and persisting summers, even though the model ensemble mean well reproduces the features in the preceding ENSO winters.

• A weak SST forced signal-to-noise ratio in the boreal summer leads to large variability in both the tropical precipitation, especially over the western Pacific, and atmospheric circulation and therefore contributes to the weak response in the model ensemble mean.

• In spite of the limitations, our results suggested that the presence of the WP precipitation is critical to reasonably simulating the extratropical teleconnection and the anomalous anticyclone over North America during the transitioning summer in the model.

The results in Chapters 3 and 4 demonstrate the challenge to study and predict ENSO summer teleconnections. Our research aimed to better understand the ENSO summer teleconnections by examining not only the physical processes, but also assessing the performance of a state-of-the-art climate model. The ultimate goal of this thesis is to improve the model’s forecast skill in North American seasonal hydroclimate. The results in these two chapters raise some interesting questions that are worth further studies in the future. For example:

• In the summer, land-atmosphere feedback strongly influences atmospheric circulation over North America. It is worthwhile to further examine to what extent the antecedent soil moisture anomalies contribute to the strong anomalous anticyclone over North America during the transitioning La Niña summer via land-atmosphere feedback. It is the anomalous anticyclone that imposes significant threats on crop yields.

• In Chapter 3, we demonstrated the substantially different characteristics between transitioning and persisting La Niña summers. Previous studies indicated that ENSO tropical forcing can impact the variability in the Great Plains low-level jet and rainfall
during the decaying summer of El Niño. It is therefore interesting to ask whether the teleconnections, and thereby impacts on the Great Plains, are substantially different between the summers when El Niño is decaying to the neutral state and the one when El Niño is transitioning to La Niña.

- We emphasized the importance of the suppressed convection over the subtropical western Pacific to the extratropical teleconnections during the transitioning summer. The drier-than-normal subtropical WP is caused by the warm SST anomalies over the Indian Ocean according to previous studies (e.g. Xie et al. 2009). However, when CAM5 was forced by historical SST, the precipitation response over the tropical Indo-western Pacific shows large variability across the ensemble members during the transitioning summer. This raises the question whether the dry anomalies over the WP during summer when an El Niño is decaying or transitioning to La Niña are driven by the SST anomalies over the Indian Ocean. Further AGCM experiments and other analyses are needed to address this question.

- In Chapter 4, we used CAM5-GOGA to examine the evolutions of atmospheric characteristics for transitioning and persisting La Niñas progressing from the preceding winters to the developing La Niña summers. We noticed that there exists an asymmetry in model performance simulating El Niño and La Niña winters. The asymmetry in performance is also there in the springs when ENSO is decaying. Further examination of the physical processes behind these asymmetries and assessment of model performance could improve the skill of seasonal forecast in current climate models.

The objective of this thesis was to examine the seasonal dependence of ENSO teleconnections and regional impacts on North American surface climate. As examples, we
focused on the impacts of El Niño on California precipitation across the winter half-year and the impacts of developing La Niñas on Midwest summertime climate. The work built on the existing literature to provide further information about the precise timing and location of anomalous precipitation in California. We also extended the study of ENSO teleconnections to the boreal summer season, establishing the importance of ENSO teleconnections during developing La Niña summers to North American summer climate. We also emphasized the need to consider the seasonal evolution within a multi-year ENSO life-cycle. This thesis synthesized both physical processes, and assessment of seasonal forecasts, based on observations, a diagnostic model, and AGCM experiments. It lays out the foundations for where attention must be paid in order to improve model forecast skill for ENSO impacts on North American seasonal hydroclimate.
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