Asian Summer Monsoon Response to Greenhouse Gases and Anthropogenic Aerosols

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ABSTRACT

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The Asian monsoon-affected area is one of the most vulnerable regions in the world facing hydroclimate changes. Anthropogenic climate change, particularly the emissions of greenhouse gases (GHGs) and aerosols, exerts significant impacts on monsoon rainfall and circulation. Understanding the effects of external forcing on monsoon rainfall is essential for improving the predictability, constraining the uncertainty, and assessing the climate risks. In this dissertation, I use a combination of observations, outputs from multiple Coupled Model Intercomparison Project - Phase 5 (CMIP5) models, and idealized atmospheric general circulation model (AGCM) experiments to examine the Asian summer monsoon variability and change. The main focus is understanding the responses to GHGs and anthropogenic aerosols and their differences for both the historical period and future projections.

The Asian monsoon is an interactive system influenced by multiple natural and anthropogenic factors. GHGs and aerosols induce significantly different changes in monsoon rainfall through both thermodynamical and dynamical processes. These changes can be further separated into the fast adjustments related to radiation and cloud processes and the slow response due to changes in sea surface temperature (SST). This thesis provides a detailed analysis of the multiple physical processes entangled in the total response, advancing our mechanistic understanding of the effects of external forcing on the Asian monsoon system and the associated uncertainties.

In Chapter 2, I first analyze the monsoon-ENSO (El Nino - Southern Oscillation) relationship in observations and CMIP5 models to determine the role of natural variability. Separating the natural and forced components shows that natural variability plays a dominant role in the 20th century, however enhanced monsoon rainfall associated with global
warming may contribute to a weakened ENSO-monsoon relation in the 21st century. In Chapter 3, I examine the physical mechanisms causing the changes of the Asian summer monsoon during the 20th and 21st century using observations and CMIP5 models, attributing the rainfall changes to the relative roles of thermodynamic and dynamic processes. CMIP5 models show a distinct drying of the Asian summer monsoon rainfall during the historical period but strong wetting for future projections, which can be explained by the strong aerosol-induced dynamical weakening during the 20th century and the thermodynamic enhancement due to GHGs in the 21st century.

In Chapters 4 and 5, I further use multiple AGCMs to separate the total monsoon response into a fast adjustment component independent of the sea surface temperature (SST) responses, and a slow response component associated with SST feedbacks. For GHGs (Chapter 4), the fast and slow monsoon circulation changes largely oppose each other, leading to an overall weak response and large inter-model spread. For aerosols (Chapter 5), the strongly weakened monsoon circulation over land due to aerosols is largely driven by the fast adjustments related to aerosol-radiation and aerosol-cloud interactions. Finally in Chapter 6, I design idealized AGCM experiments with prescribed SSTs using the Community Atmosphere Model (CAM5) and the Geophysical Fluid Dynamic Laboratory Model (GFDL-AM3) to investigate the relative roles of uniform SST warming/cooling as well as global and regional SST patterns in shaping the differing monsoon responses. While GHGs-induced SST changes affect the monsoon largely via the uniform warming effect, for aerosols the SST spatial pattern plays the dominant role through changes in atmospheric circulation.
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Chapter 1

Introduction

1.1 The Asian monsoon system

The most populated regions of the Asian continent are characterized by a distinct monsoon climate, spanning from the Indian subcontinent to the extratropics of eastern Asia. The “monsoon” describes the seasonal phenomenon that produces the majority of rainfall during the summer months, characterized by a “wet” summer and a “dry” winter, and a seasonal reversal of surface winds. The Asian monsoon is one of the major monsoon systems in the world, where the summer monsoon brings over 80% of the annual rainfall, with critical importance for various socio-economic sectors. In addition, as a major part of the summertime overturning circulation in the northern hemisphere tropics, the Asian monsoon has profound remote influences on the global-scale climate (Rodwell and Hoskins 1996; Lin and Wu 2012).

With its large population, increasing industrial development and severe water stresses, Asia is one of the most vulnerable regions in the world facing hydroclimate changes. Piao et al. (2010) assessed the impacts of climate change on water resources and agriculture in China, suggesting a 20% crop production decrease by 2050 under the worst-case scenario. However the overall impact is far from certain due to the high variability and uncertainty in projected climate, particularly precipitation, and the corresponding crop responses. Analyzing and understanding the characteristics of monsoon change has important implications for socioeconomic development and human well-being, including water resource manage-
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ment, agriculture, ecosystem service, food security, and public health (e.g. Kumar et al. 2004; Hong and Kim 2011; Mirza 2011).

The Asian monsoon is an interactive system influenced by both internal variability and external forcing (e.g. Lei et al. 2011; Turner and Annamalai 2012; Song et al. 2014). On interannual timescale, the Asian summer monsoon variability is strongly associated with the El Niño-Southern Oscillation (ENSO), as suggested by the abundance of observational, modeling and paleoclimate studies (e.g., Kumar et al. 2006; Cook et al. 2010; Mishra et al. 2012). During a warm ENSO event with anomalously high sea surface temperature (SST) over the eastern tropical Pacific, convection in the western Pacific shifts to the central and eastern Pacific, which then suppresses the convection in the Asian monsoon region as well as the western Pacific warm pool region, resulting in monsoon failure (e.g., Rasmusson and Carpenter 1983; Ropelewski and Halpert 1987; Webster and Yang 1992; Lau and Nath 2000; Wang et al. 2003). The ENSO signal has long been regarded as an important seasonal predictor for monsoon rainfall (e.g., Webster and Yang 1992; Ju and Slingo 1995).

1.2 The human-induced climate change

1.2.1 Greenhouse gases

The climate system has warmed significantly since the preindustrial period, predominantly driven by the increasing greenhouse gas (GHG) emissions as a result of the rapid industrialization. The atmospheric concentrations of GHGs including carbon dioxide ($CO_2$), methane ($CH_4$), and nitrous oxide ($N_2O$) were 391 ppm, 1803 ppb, and 324 ppb in 2011, with increases by about 40%, 150%, and 20% from preindustrial levels, respectively (Hartmann et al. 2013). The Intergovernmental Panel for Climate Change (IPCC) Fifth Assessment Report (AR5) provided an estimate of the radiative forcing for 2011 relative to 1750, which quantifies the net change in the energy balance of the Earth system due to the imposed perturbations (Myhre et al. 2013). The radiative forcing from emissions of well-mixed GHGs ($CO_2$, $CH_4$, $N_2O$, and Halocarbons) is $3.00 \ W \ m^{-2}$, in which the emissions of $CO_2$ alone causing $1.68 \ W \ m^{-2}$, followed by $CH_4$ with $0.97 \ W \ m^{-2}$. This strong positive radiative forcing greatly contributes to surface warming, which based on the linear
trend of observed data, shows a warming of 0.85 °C for globally averaged land and ocean surface temperature over the period 1880 to 2012. The increase in GHGs is the largest contribution to the total radiative forcing since the preindustrial period, and is expected to continue in the 21st century. GHGs have long lifetimes and are well-mixed in the atmosphere, therefore the spatial distribution of the radiative forcing is largely homogeneous. As the dominant driver of anthropogenic climate change, GHG forcing has global scale impacts on multiple aspects of the climate system, including the atmosphere, ocean, land surface, the cryosphere, as well as the hydrological cycle.

1.2.2 Anthropogenic aerosols

Aerosols are minute solid particles or liquid droplets suspended in the atmosphere. Atmospheric aerosols can originate from both natural and human sources through two different pathways: 1) direct emissions of primary particulate matter, and 2) formation of secondary particulate matter from precursor gases. Natural sources of aerosols include sea salt, mineral dust by wind erosion of dry soils (e.g., deserts), volcanic eruptions, and primary biological aerosol particles (Boucher et al. 2013). Anthropogenic aerosols are mainly produced from combustion of fossil fuels and biomass burning, primarily consisting of sulfate, black carbon (BC), organic carbon (OC), nitrate, and ammonium.

Over the past few years, “Asian smog” has become one of the most pressing environmental threats around the world featuring unprecedentedly high pollution levels. The emission of anthropogenic aerosols has severe consequences on human health. According to a study conducted by the Global Burden of Disease from Major Air Pollution Sources (GBD MAPS) Working Group, ambient fine particular matter ($PM_{2.5}$) is a major contributor to mortality and disease burden in China, with an estimated contribution of 916,000 deaths in 2013, among which 40% is caused by coal combustion (GBD MAPS Working Group 2016). Apart from the direct health impacts, air pollution has been shown to affect meteorological conditions, as well as long-term climate variations (Gong et al. 2007; Boucher et al. 2013). The effects of aerosols on global and regional climate can be through two major physical pathways: 1) by altering the radiation budget through scattering or absorbing incoming solar radiation (direct effect), and 2) by interacting with clouds through microphysical processes
in which aerosols act as cloud condensation nuclei, therefore creating smaller cloud droplets and brighter clouds (1st aerosol indirect effect) as well as prolonging the cloud lifetime (2nd aerosol indirect effect). The direct and indirect effects of aerosols can cause subsequent changes in surface temperature as well as the hydrological cycle (Allen and Ingram 2002; Ming et al. 2010; Levy et al. 2013; Li et al. 2016).

The radiative forcing of the total aerosol effect, including the cloud adjustments, is $-0.9 \text{ W m}^{-2}$, which is a combination of a negative forcing from most aerosols (particularly sulfate), and a positive contribution from black carbon absorption of solar radiation (Myhre et al. 2013). Aerosols offset a substantial portion of the positive forcing from well-mixed GHGs during the historical period. As a result, the total radiative forcing for 2011 relative to 1750 is $2.29 \text{ W m}^{-2}$, with other forcing agents such as land use and solar irradiance contributing little. However, the aerosol radiative forcing is the largest uncertainty to the total radiative forcing estimate. Aerosols also have short lifetimes in the atmosphere from days to weeks and thus mostly reside closely to the emission source. As a result, compared to the relatively homogeneous spatial distribution of well-mixed GHGs, the aerosol radiative forcing displays large meridional asymmetry concentrated over the northern hemisphere continents (Shindell et al. 2013). For the 21st century, the projected decline in anthropogenic aerosol emissions may contribute to further warming alongside the increasing GHGs. Understanding the climate effects of aerosols is critical for constraining the uncertainty of future climate projections.

1.3 Monsoons in a changing climate: theoretical perspectives

This section consists of a brief overview of the current theoretical understanding for the large-scale changes of the hydrological cycle and monsoons to GHGs and aerosols. A schematic summarizing the major physical processes and mechanisms is provided in Fig. 1.1.

1.3.1 Thermodynamic and dynamic mechanisms

Previous studies have advanced our understanding of the physical mechanisms causing the changes of the hydrological cycle in response to global warming on a global scale. For
large-scale changes, the “wet-get-wetter” or “rich-get-richer” mechanism (Held and Soden 2006) emphasizes the thermodynamic effect due to the increase in lower tropospheric water vapor in a warming atmosphere. Based on the Clausius-Clapeyron equation, the amount of saturation vapor pressure increases at a rate of about 7% per 1 K of temperature rise. Consequently, horizontal moisture transport increases within the atmosphere, leading to enhanced precipitation minus evaporation ($P - E$) where mean moisture converges and reduced $P - E$ where moisture diverges. However precipitation increases are also controlled by the radiative (energy) constraints (e.g. Takahashi 2009; O’Gorman et al. 2012), making the rate of precipitation increase less than the 7% per 1 K. As a consequence, the atmospheric overturning circulation slows down as climate warms, especially for the Walker circulation in the tropics (Held and Soden 2006; Vecchi and Soden 2007). For spatial distribution of the tropical rainfall response, another fundamental view is the “warmer-get-wetter” mechanism (Xie et al. 2010) over the tropical ocean, suggesting that precipitation increases where the SST warming exceeds the tropical mean and vice versa.

On the regional scale, hydroclimate projections from state-of-the-art climate models show large uncertainty and model spread, particularly in the tropics and over the monsoon regions (Turner and Annamalai 2012; Christensen et al. 2013). The Asian summer monsoon precipitation is projected to enhance under greenhouse warming, dominated by the “wet-get-
wetter” thermodynamic mechanism (Kamae et al. 2014b; Wang et al. 2014). On the other hand, dynamical changes related to atmospheric circulation are relatively weak with a low model-agreement (Endo and Kitoh 2014). As highlighted by Xie et al. (2015), atmospheric circulation is the major source of uncertainty in regional rainfall projection.

A number of studies have addressed the possible physical mechanisms of aerosol-induced tropical rainfall and monsoon changes (Lau et al. 2006; Ganguly et al. 2012a; Guo et al. 2013, 2015; Hwang et al. 2013). Unlike GHGs that induce a strong atmospheric moistening, aerosols affect monsoon rainfall largely through changes in atmospheric circulation. Increased aerosols in the atmosphere could reduce the surface solar radiation (“dimming” effect) which reduces the local SST gradient in the Indian Ocean (Ramanathan et al. 2005; Chung and Ramanathan 2006) and introduces a hemispheric energy imbalance caused by the spatial inhomogeneity of aerosol distributions (Bollasina et al. 2011), as well as increase atmospheric stability through direct and indirect effects (Lau and Kim 2017), contributing to weakened monsoon circulation. Some other studies find that aerosols may cause an earlier onset and enhanced June rainfall over India due to absorbing aerosols such as black carbon through the “elevated heat pump” hypothesis (Lau et al. 2006) in which aerosols could enhance the meridional temperature gradient in the mid-to-upper troposphere, indicating high complexity and uncertainty associated with aerosol-monsoon interactions.

Although there is a general consensus on the large-scale thermodynamic mechanism of hydroclimate change, the dynamical response is not well understood. Furthermore, the effects of climate change on the regional scale, especially monsoon strength and variability, are complex and uncertain (Christensen et al. 2013). The question of how Asian monsoon rainfall and circulation may respond to future anthropogenic forcing, including both GHGs and aerosols, is far from conclusive.

1.3.2 Fast and slow responses

From an energetics perspective, the response of the climate system to an external forcing involves two components on different time scales: the fast response without the mediation of SST and the slow response due to SST changes (Allen and Ingram 2002; Andrews et al. 2009). The fast response refers to the rapid adjustment of the atmosphere to the forcing
before SST changes occur, which includes the perturbation to the radiation budget and for aerosols, also the interactions with clouds. The slow response refers to the feedbacks induced by subsequent SST changes. The fast and slow components may lead to differing responses in the hydrological cycle, often studied using idealized atmospheric general circulation model (AGCM) experiments with prescribed SSTs. Recent studies have shown that the direct radiative forcing of CO$_2$ and SST warming may cause different responses in tropical circulation (Ma et al. 2012; He and Soden 2015a), summertime Pacific anticyclone and the Asian monsoon cyclone (Shaw and Voigt 2015), and midlatitude jets (Grise and Polvani 2014). For tropical rainfall and circulation, Bony et al. (2013) find that a large fraction of the long-term regional precipitation change can be explained by the direct atmospheric radiative response that occurs shortly after an abrupt CO$_2$ increase, independent of surface warming. On the contrary, Chadwick et al. (2014) argue that the fast dynamical precipitation response as shown in Bony et al. (2013) is dominated by surface warming patterns rather than the direct radiative effect. Several studies have shown that the direct radiative forcing and SST change exert different effects onto the land-sea thermal contrast changes, which then influence the atmospheric thermodynamic structures and circulation patterns (e.g. Joshi et al. 2008; Kamae et al. 2014a). However for anthropogenic aerosols, it has been shown that the slow response due to SST change may dominate the total monsoon rainfall and circulation changes using a single AGCM (Ganguly et al. 2012b).

For the slow response, an external forcing (GHGs/aerosols) can alter precipitation and atmospheric circulation through changes on both the mean SST and the spatial pattern. For example, the asymmetric SST response to aerosols can lead to changes in the Hadley circulation and a southward shift of the intertropical convergence zone (ITCZ) (Kang et al. 2008; Ming and Ramaswamy 2009, 2011; Seo et al. 2014). Hill et al. (2015) show that the Hadley circulation weakens due to both mean and patterned SST changes driven by GHGs, while aerosols induce northward energy flux anomalies largely due to the spatial pattern. Furthermore, despite the difference in the spatial distribution of GHG and aerosol forcing, Xie et al. (2013) find that GHGs and aerosol induce similar responses in the spatial pattern of SSTs and oceanic rainfall. Studies have further shown that the SST spatial pattern dominates the uncertainty and intermodel spread of precipitation and atmospheric
CHAPTER 1. INTRODUCTION

circulation in global climate models (Ma and Xie 2013; Kent et al. 2015; Chen and Zhou 2015).

1.4 Evaluation of climate model simulations

Global climate models, or general circulation models (GCMs), are widely utilized in climate research. Current state-of-the-art GCMs - the Coupled Model Intercomparison Project Phase 5 (CMIP5) models for IPCC AR5 - have been shown to have improved in many aspects since the CMIP3 models for IPCC AR4. However, models generally perform less well for precipitation than for surface temperature (Flato et al. 2013). This section provides a brief evaluation of the CMIP5 models used in the subsequent chapters, focusing on monsoon characteristics.

Figure 1.2 shows 30-year climatology of summertime monsoon rainfall and low-level circulation in observations and the multi-model mean (MMM) of 35 CMIP5 models (109 total realizations). Observed rainfall data is from the Climate Research Unit (CRU) at the University of East Anglia (UEA) version 3.2 (Harris et al. 2014). Winds are from the 20th Century Reanalysis Project version 2 (Compo et al. 2011). The climatological monsoon circulation features the cross-equatorial flow over the Indian Ocean, the southwesterlies from the Arabian Sea to the Bay of Bengal, and the southerly winds over the South China Sea and eastern China. On the large scale, CMIP5 MMMs show reasonable skills in simulating the climatological monsoon features. Although as illustrated by previous studies (e.g. Sperber et al. 2013), CMIP5 models generally underestimate monsoon rainfall over India, associated with the weak low-level monsoon flow.

Figure 1.3 shows the Taylor diagram (Taylor 2001) of the rainfall climatology over the Asian monsoon region ($5^\circ N - 55^\circ N$, $60^\circ E - 150^\circ E$, land-only) in 35 CMIP5 coupled models and 11 models from the Atmospheric Model Intercomparison Project (AMIP). AMIP models are atmosphere-only models without a fully-coupled ocean. The Global Precipitation Climatology Centre (GPCC) Full Data Product version 6 from the World Climate Research Programme (WCRP) Global Climate Observing System (GCOS) (Schneider et al. 2011) is also used as a comparison to the CRU dataset. The spatial resolution is $0.5^\circ \times 0.5^\circ$ for both
Figure 1.2: 1976-2005 climatology of June-August (a, b) precipitation (mm day$^{-1}$) and (c, d) 850hPa wind (m s$^{-1}$) for (a) CRU, (c) 20CR, and (b, d) multi-model mean of 35 CMIP5 models. For (c, d), arrows show wind vectors and colors are the wind speed.

datasets. All observed and modeled data are interpolated into a 1° × 1° spatial resolution for direct comparison. The time period for the climatology is 1976-2005 for CMIP5 models and the 30 years available for the AMIP models. CRU is chosen as the reference.

The spatial correlations range from 0.6 to 0.8 for most of the coupled models, with comparable standard deviations as observations. The coupled (black) and atmosphere-only (red) models do not exhibit significant differences, suggesting that ocean coupling does not have much effect in simulating monsoon climatology. Only one AGCM (FGOALS-g2) produces an unrealistically large standard deviation due to an overestimation of rainfall over India and Indochina (not shown), thus this model has been eliminated from further analysis.

Although CMIP5 models are able to simulate the climatological characteristics of monsoon rainfall and circulation reasonably well, the biases and model spreads for the externally forced changes are much more prominent. The uncertainties of long-term monsoon trends and monsoon variability (ENSO-monsoon teleconnection) in global climate models will be
Figure 1.3: Taylor diagram showing (blue dot-dashed lines) the spatial pattern correlation coefficient, (black dotted contours) standard deviation, (green dashed contours) the root-mean-square difference (RMSD) for June-August area averaged ($5^\circ N - 55^\circ N, 60^\circ E - 150^\circ E$) land precipitation climatology in (black dots) 35 CMIP5 models (1976-2005) and (red dots) 11 AMIP models. Orange dots show observations (CRU, GPCC). CRU is used as the reference field. Rainfall is in mm day$^{-1}$. 
CHAPTER 1. INTRODUCTION

discussed in the subsequent chapters.

1.5 Scientific questions and outline

The objective of this dissertation is to obtain a thorough mechanistic understanding of the response of the Asian summer monsoon system to anthropogenic forcing - specifically GHGs and anthropogenic aerosols - for both the historical period and future projections. Understanding the effects of external forcing on monsoon rainfall is essential for improving the predictability, constraining the uncertainty, as well as climate adaptation and mitigation policies. I use observations, large ensemble sets from Coupled Model Intercomparison Project - Phase 5 (CMIP5) models, and a combination of statistical methods and idealized AGCM experiments to perform the analysis.

The main scientific questions addressed in this thesis include:

- What are the effects of external forcing on monsoon variability?

- What are the changes and mechanisms of the Asian summer monsoon in response to anthropogenic forcing for the historical period and future projections?

- What are the relative roles of the fast adjustments and the slow response (via uniform change and SST spatial pattern) for GHGs and aerosols?

- What are the possible factors contributing to the uncertainty of monsoon changes in climate model simulations?

The remaining chapters are outlined as follows. In Chapter 2, I analyze the ENSO-monsoon teleconnection in observations and CMIP5 models to examine the interannual variability of the Asian monsoon and determine the effects of external forcing. In Chapter 3, I examine the physical mechanisms causing the changes of the Asian summer monsoon during the 20th and 21st century, and attribute the changes to the relative roles of GHGs and aerosols. I then focus on quantifying the fast and slow responses of the monsoon to GHGs and aerosols using idealized AGCM experiments, presented in Chapters 4 and 5, respectively. Chapter 6 further addresses the relative roles of uniform SST change and the
CHAPTER 1. INTRODUCTION

SST spatial pattern on the Asian monsoon through designing a suite of AGCM experiments using the Community Atmosphere Model version 5.3 (CAM5) and the Geophysical Fluid Dynamics Atmospheric Model version 3 (GFDL-AM3). The main conclusions and future research directions are summarized in Chapter 7.
Chapter 2

Recent and future changes in the Asian monsoon - ENSO relationship: Natural or forced?


2.1 Introduction

The Asian monsoon variability and change have attracted tremendous attention in the climate research community (e.g., Kumar et al. 2006; Turner and Annamalai 2012; Wang et al. 2014) due to the system’s serious impacts on the region’s socio-economic development (e.g., Kumar et al. 2004; Mirza 2011). On interannual timescale, the Asian summer monsoon variability is strongly associated with the El Niño-Southern Oscillation (ENSO), as suggested by the abundance of observational, modeling and paleoclimate studies (e.g., Kumar et al. 2006; Cook et al. 2010; Mishra et al. 2012). During a warm ENSO event with anomalously high sea surface temperature (SST) over the eastern tropical Pacific, convection in the western Pacific shifts to the central and eastern Pacific, which then suppresses the convection in the Asian monsoon region as well as the western Pacific warm pool region, resulting in monsoon failure (e.g., Rasmusson and Carpenter 1983; Ropelewski and Halpert
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON -
ENSO RELATIONSHIP: NATURAL OR FORCED?

Recent observational analyses, however, suggest that the ENSO-monsoon relationship may have weakened or broken down for the recent decades (Kumar et al. 1999; Kinter et al. 2002). Some studies suggest it is likely that anthropogenic global warming may have been the cause of the weakening relationship. Kumar et al. (1999) examined observed data and proposed two possible reasons for the changing ENSO-monsoon relation due to global warming: first, the southeastward shift of the climatological Walker Circulation due to greenhouse warming disconnects the Indian monsoon from the region of ENSO impact; second, the increased winter and spring surface temperature and reduced snow cover over Eurasia due to greenhouse warming enhances land-sea thermal contrast and favors a stronger monsoon, thus reduces the negative impact of El Niño on monsoon rainfall. Ashrit et al. (2001), based on a single realization of the Max-Plank Institute climate model, confirmed that the increase in ground temperature over Eurasian continent due to greenhouse warming led to a stronger monsoon and thus a reduced impact from ENSO. Thus there are indications in both models and observations that a weakening of the ENSO-monsoon relationship could occur in a warmer future climate. On the contrary, several studies suggest that this weakening relationship is more likely due to natural climatic variability rather than a response to global warming (e.g., Kripalani et al. 2003; Kitoh 2007). Using multiple model simulations for the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) for the 20th century and global warming experiments in which the atmospheric CO₂ concentration was raised to twice the preindustrial value, Annamalai et al. (2007) showed that the ENSO-monsoon correlation in the warming runs is very similar to that in the historical simulations, countering the argument that the connection will weaken as climate warms.

It is noted that the ENSO-monsoon system has undergone significant interdecadal changes in observations (Torrence and Webster 1999), paleoclimate reconstructions (Berkelhammer et al. 2014), and long-term control simulations of coupled general circulation models (GCMs) (Kitoh 2007; Hunt 2014). Krishnamurthy and Goswami (2000) proposed the hypothesis that the interdecadal variations of the Indian summer monsoon and tropical SST are part of a tropical ocean-atmosphere coupled mode. Chowdary et al. (2012) suggest that
the interdecadal variations of ENSO teleconnections over the Indo-western Pacific is mainly governed by ENSO variance itself. Several studies have proposed possible mechanisms linking the decadal change of the ENSO-monsoon system to other large-scale climate variability. For example, Kinter et al. (2002) related the change to north Pacific SST and atmospheric circulation. Krishnamurthy and Krishnamurthy (2014) showed that the Pacific Decadal Oscillation (PDO) can enhance (counteract) the ENSO-monsoon relation when ENSO and PDO are in (out of) phase. Chang et al. (2001) suggest that the weakening ENSO-monsoon relation is most likely due to the strengthening and poleward shift of the jet stream over the North Atlantic. Using a coupled GCM, Chen et al. (2010) found that the Atlantic Multidecadal Oscillation (AMO) could induce coupled feedbacks in the tropical Pacific and modulate the multidecadal variation of the ENSO-monsoon connection. Some studies also address the influence of statistical stochasticism in analyzing the relationship between ENSO and monsoon rainfall (Hunt 2014). Gershunov et al. (2001) point out that running correlations between pairs of stochastic time series are typically characterized by low-frequency evolution, thus the decadal fluctuations in the observed ENSO-monsoon relationship can be explained by sampling variability alone.

The ENSO signal has long been regarded as an important seasonal predictor for monsoon rainfall (e.g., Webster and Yang 1992; Ju and Slingo 1995). Thus understanding how the ENSO-monsoon relationship might change on longer timescales has strong implications to successful monsoon forecasts. In particular, it is vital to understand whether the recent weakening of the ENSO-monsoon relationship is related to anthropogenic climate change, and if so, how would one expect it to change in the future. The lack of consensus in previous studies reveals substantial uncertainties in observations and model simulations, as well as analysis methods. Many questions remain to be addressed, such as the robustness of the change in ENSO-monsoon relationship in different models, and how the anthropogenically forced component can be separated from the naturally varying ENSO-monsoon relationship.

We examined the ENSO-monsoon relationship for the 20th and 21st centuries using observations and the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project - Phase 5 (CMIP5) model simulations. We address the following question in this chapter: what are the forced and natural patterns of the ENSO-monsoon
teleconnection and how is this relationship changing in the future? The chapter is organized as follows. Section 2 describes observational datasets and model simulations used in the study, as well as analysis methodology. Section 3 presents results regarding the natural and forced SST-monsoon relationship. The main conclusions are summarized in section 4.

2.2 Data and Methods

For observed precipitation, we used monthly data from the Climate Research Unit (CRU) at the University of East Anglia (UEA) version 3.2 (Harris et al. 2014). The spatial resolution is \( 0.5^\circ \times 0.5^\circ \), interpolated to \( 1^\circ \times 1^\circ \) to allow for a higher rain gauge count within each grid box. Only the grid boxes where at least one rain gauge existed in any month of the June-August (JJA) season for at least 80 years during the 1901-2005 period were used in the analysis in order to improve the data reliability. We used observed monthly SST from the National Oceanic and Atmospheric Administration (NOAA) National Climate Data Center (NCDC) Extended Reconstructed Sea Surface Temperatures (ERSST) version 3b (Smith et al. 2008), and winds from the 20th Century Reanalysis (20CR) Project version 2 (Compo et al. 2011), both with a \( 2^\circ \times 2^\circ \) spatial resolution.

Model simulations in this study include a multi-model ensemble of CMIP5 models (Taylor et al. 2012) used in IPCC AR5. We used monthly output of 34 models for preindustrial control simulations, as well as all realizations for historical simulations and future projections under the high-end representative concentration pathway 8.5 (rcp8.5) emission scenario. Altogether 103 realizations were analyzed for the historical period and 71 for rcp8.5 (listed in Table 2.1). All model outputs were interpolated to a \( 1^\circ \times 1^\circ \) spatial resolution for precipitation and \( 2^\circ \times 2^\circ \) for SST.

In order to separate the naturally varying ENSO-monsoon relation and the forced response of rainfall to SST warming, we applied signal-to-noise (S/N) maximizing empirical orthogonal function (EOF) analysis (Allen and Smith 1997; Venzke et al. 1999; Chang et al. 2000) to JJA seasonal averaged global SST of the CMIP5 multi-model ensemble to extract the externally forced signal, as in Ting et al. (2009, 2011). Given the multi-model, multi-realization ensemble, the total covariance matrix of the ensemble mean can be assumed as
Table 2.1: List of CMIP5 models and the number of realizations used in Chapter 2.

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a sum of two linearly independent matrices, one for forced signal and one for internal variability (“climate noise”). We first determined the spatial structure of the internal modes of variability through an EOF analysis on the “noise” matrix, formed using the second century of the simulated SST in the preindustrial control run for each of the corresponding CMIP5 model in the analysis. These noise EOFs (retaining 80% of the total variance) were used to form a spatial pre-whitening transformation matrix to filter the internal variability that was not removed by the multi-model ensemble averaging, so that the spatial covariance in the ensemble average is largely due to forced change. The leading EOF mode of the ensemble mean gives the dominant forced signal (Fig. 2.1). We used the leading principle component (S/N PC1) as the forced index. For observations and each realization of model simulations, we regressed JJA seasonal average SST anomalies \((SST_{\text{total}})\) onto S/N PC1 to obtain the forced component:

\[
k(x, y) = r(x, y) \frac{\sigma_{SST_{\text{total}}}(x, y, t)}{\sigma_{\text{PC1}}(t)},
\]

\[
SST_{\text{forced}}(x, y, t) = k(x, y) \cdot \text{PC1}(t),
\]

where \(r(x, y)\) is the correlation coefficient between grid point SST and S/N PC1 and \(\sigma\) the corresponding standard deviation. We calculated the difference between \(SST_{\text{total}}\) and \(SST_{\text{forced}}\) to obtain the natural component \((SST_{\text{natural}})\):

\[
SST_{\text{natural}}(x, y, t) = SST_{\text{total}}(x, y, t) - SST_{\text{forced}}(x, y, t),
\]

as in Ting et al. (2009, 2011); Kelley et al. (2012). We then averaged \(SST_{\text{total}}, SST_{\text{natural}}\) and \(SST_{\text{forced}}\) over the Niño 3.4 box \((5^{\circ}S - 5^{\circ}N, 170^{\circ}W - 120^{\circ}W)\) (Trenberth 1997) to obtain the total, natural and forced Niño 3.4 SST indices, respectively.

2.3 Results

We first examined the spatial structure of the ENSO-monsoon relationship in observations and CMIP5 models. Figure 2.2 shows regressions of JJA seasonal average monsoon rainfall and global SST onto the total Niño 3.4 SST index for observations (a, b) and CMIP5 historical simulations (c, d), for 1901-2005, and future projections under the rcp8.5 scenario for 2006-2099 (e, f). The CMIP5 results (Figs. 2.2c-f) are the multi-model mean (MMM)
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON - ENSO RELATIONSHIP: NATURAL OR FORCED?

Figure 2.1: First EOF mode of S/N maximizing EOF analysis of JJA global SST for CMIP5 historical 1901-2005 (a, c) and rcp8.5 2006-2099 (b, d) simulations using 34 CMIP5 models with 103 (historical) and 71 (rcp8.5) realizations. (a, b) Spatial structures, shown as regressions of SST anomalies onto standardized S/N PC1, stipping denotes 26/34 model-agreement; and (c, d) standardized leading principle components (S/N PC1).

patterns of the regression coefficients. In calculating the MMM, we first computed each model’s ensemble mean regression pattern, and then performed multi-model ensemble average using each model’s ensemble mean. The observed (Fig. 2.2a) and modeled (Fig. 2.2c) rainfall patterns both show the expected strong inverse relationship with ENSO over the Indian subcontinent. The robustness is indicated by the statistical significance in Fig. 2.2a and the high degree of model-agreement in Fig. 2.2c. The CMIP5 MMM captures the ENSO signal in monsoon rainfall variability reasonably well. The global SST regressions show the canonical ENSO patterns in both the observation (Fig. 2.2b) and the CMIP5
Figure 2.2: Regressions of JJA precipitation (a, c, e) and global SST (b, d, f) onto total Niño 3.4 SST index for observations (CRU, ERSST) 1901-2005 (a, b), CMIP5 34-model MMM historical 1901-2005 (c, d), and rcp8.5 2006-2099 (e, f) simulations. For observed precipitation (a), only the grid boxes where at least one rain gauge existed in any month of the JJA season for at least 80 years are plotted. Stippling denotes 5% significance based on 2-sided Student’s t test in (a-b) and 26/34 model-agreement in (c-f). Units are mm day$^{-1}$ °C$^{-1}$ for (a, c, e), and °C °C$^{-1}$ for (b, d, f).
MM (Fig. 2.2d). The ENSO-related SST anomaly in the CMIP5 MMM extends further west compared to the observed pattern. This has been noted in previous studies to be a systematic bias of CMIP5 models, driven by unrealistic westward displacement and overestimation of the equatorial wind stress in the western Pacific (Taschetto et al. 2014). For rcp8.5, the rainfall pattern (Fig. 2.2e) exhibits clearly weakened ENSO-monsoon connection with low model-agreement in the Indian subcontinent. The SST pattern (Fig. 2.2f) displays mixed signals of both ENSO and global warming trend. Thus the change in rainfall pattern in Fig. 2.2e may be a combination of natural variability and anthropogenic forcing.

To illustrate the temporal evolution of the ENSO-monsoon relation, we computed running correlations with a 31-year sliding window between the total Niño 3.4 SST index and the all-India average rainfall (weighted area average of 5°N – 30°N, 70°E – 90°E), for the JJA season. We used the second century of preindustrial control runs for the 34 models to examine the natural fluctuations without any external forcing, as shown in Fig. 2.3a. The correlation values are plotted against the midpoint years of the 31-year window and box-whiskers of the model spread are plotted every five years. The box edges give the 25th to 75th percentile, and the whiskers extend from each end of the box to the maximum values within 1.5 times ($w = 1.5$) of the interquartile range, which corresponds to approximately $+/-2.7$ standard deviation and 99.3% coverage if the data were normally distributed. The correlation coefficients of each model are plotted in colored dots. We ranked the models based on the correlation values for the first 31-year period and assigned each model a corresponding color. The green line shows the MMM. The corresponding results for the historical and rcp8.5 simulations are shown in Figs. 2.3b and 2.3c, using the first realization for each model. The black thick line in Fig. 2.3b is the observed 31-year running correlation based on ERSST and CRU precipitation, for a slightly extended period to 2011.

The observed ENSO-monsoon relationship in Fig. 2.3b shows a weakening trend since the 1970s, with a $\sim 0.25$ maximum decrease in correlation value from below $-0.6$ to slightly above $-0.4$. However there are prominent decadal fluctuations throughout the 20th century, with a range of $\sim 0.35$ in correlation value. The first half of the century shows a similar weakening trend with a maximum decrease ($\sim 0.3$) slightly exceeding the recent period. Additionally, the correlation trend reversed after 1990, suggesting a possible strengthen-
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON - ENSO RELATIONSHIP: NATURAL OR FORCED?

Figure 2.3: 31-year running correlation between JJA India area averaged precipitation and total Niño 3.4 SST index for CMIP5 pre-industrial control runs (a), historical (b) and rcp8.5 (c), using the first realization for each model. The colored dots are correlation coefficients for individual models, plotted every 5 years. The box edges give the 25th to 75th percentile, the whiskers extend from each end of the box to the maximum values within 1.5 times the interquartile range, the black line in each box shows the median. Green thick lines show CMIP5 MMMs, black thick line in (b) shows the result of observations using CRU and ERSST. Dashed lines denote 5% significance based on 2-sided Student’s t test.
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ing of the correlation in the most recent years. A significance test using the method in Gershunov et al. (2001) shows that the variation of the running correlation in observations during the 20th century is not significantly different from what one would expect for correlated Gaussian noise processes at the 5% level. Thus one cannot rule out the possibility that stochastic noise may contribute to a low-frequency correlation change as large as in observations. Comparing to model-simulated results, the observed fluctuations in correlation in Fig. 2.3b is within the spread of the CMIP5 historical runs as well as the preindustrial simulations (Fig. 2.3a). The modeled ENSO-monsoon relation ranges from weakly positive to highly negative values, while the MMMs in preindustrial and historical simulations are both significantly weaker than that for observations. This implies that although the MMM is able to capture the spatial pattern of the ENSO-monsoon connection (Figs. 2.2a, c), there are large uncertainties within the model ensemble in the magnitude and, in some cases, even sign of the ENSO-monsoon relationship. The correlation values for individual models (colored dots) show that the spread comes from discrepancy from model to model rather than fluctuations within any single model, as the group of models with higher (lower) correlations remains rather consistent with time. In other words, models with higher correlation values tend to remain high throughout the century and vice versa. The MMM of historical simulations (Fig. 2.3b) has little change compared to the preindustrial runs (Fig. 2.3a). In the rcp8.5 case (Fig. 2.3c), the MMM shows a slightly decreased correlation, dropping below the statistical significance value denoted by the dashed line.

Figure 2.3 suggests that anthropogenic forcing has a minor effect in the rcp8.5 case with strong anthropogenic warming, but contributes little to the ENSO-monsoon relation in the historical period. Thus the weakening ENSO-monsoon relation in the recent decades is more likely due to decadal variability rather than the global warming trend. It is possible that greenhouse warming could contribute to a slight weakening of the ENSO-monsoon relationship in the future. However internal variability tends to dominate, and there is substantial uncertainty with low agreement among the models. Comparing our results to earlier studies, Ashrit et al. (2005) showed a much stronger weakening of the ENSO-monsoon relation in the 21st century, particularly after 2050. Ashrit et al. (2003), on the other hand, reported no systematic change of the ENSO-monsoon relation due to greenhouse
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warning. These diverse results also indicate a high degree of uncertainty possibly due to model discrepancies in terms of ENSO-monsoon simulations (Annamalai et al. 2007).

To separate the natural and forced ENSO-monsoon components, we applied regression analysis onto the natural and forced Niño 3.4 SST indices. Figure 2.4 shows the rainfall (a, c, e) and global SST (b, d, f) regression fields associated with the natural Niño 3.4 SST index for observations (a, b), CMIP5 historical (c, d) and rcp8.5 (e, f). Both the observed (Fig. 2.4a) and CMIP5 historical rainfall regressions (Fig. 2.4c) are very similar to that for the total Niño 3.4 index in Figs. 2.2a and 2.2c, indicating that natural variability plays the dominant role in the varied ENSO-monsoon relationship during the 20th century. The natural component of the SST regressions (b, d) are also similar to the total fields (Figs. 2.2b, d), with weaker positive correlations over regions such as western tropical Pacific, tropical and northern Atlantic, and the Southern Ocean. The natural components of the regression patterns for rcp8.5 (Figs. 2.4e, f), on the other hand, show distinctly different features from the total regressions (Figs. 2.2e, f). The rainfall pattern (Fig. 2.4e) gives a clear negative ENSO signal over India, and is remarkably similar to the historical case (Fig. 2.4c). The SST regression (Fig. 2.4f) displays the typical ENSO pattern almost undistinguishable from the historical SST regression in Fig. 2.4d, and, significantly different from that in Fig. 2.2f when the forced component is included. The high degree of similarity between Figs. 2.4c and 2.4e, and between Figs. 2.4d and 2.4f, indicates that the unforced ENSO-monsoon relationship as simulated in CMIP5 MMM is insensitive to the external radiative forcing. In all three cases, the 850hPa wind regressions (arrows) show westerly anomalies in the western Pacific, indicating a shift of the Walker circulation towards the central Pacific and the Asian monsoon is part of that shift. In the northern Indian Ocean, there are easterly wind anomalies that oppose the climatological monsoon circulation. This anticyclonic feature over the Arabian Sea and the surrounding land areas has been addressed in Lau and Nath (2000); Lau and Wang (2006) to be associated with the anomalous diabatic cooling in the western tropical Pacific during El Niño events. These circulation changes are consistent with the reduced monsoon rainfall during a warm ENSO phase.

The forced SST-monsoon relationship can be seen from the rainfall regressions to the forced Niño 3.4 SST time series, as shown in Fig. 2.5 for observations (a), CMIP5 historical
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Figure 2.4: Shading and stippling as in Fig. 2.2, but using natural Niño 3.4 SST index. Arrows show regressions of 850hPa winds onto natural Niño 3.4 SST index, units are $m \, s^{-1} \, ^oC^{-1}$. 

25
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON - ENSO RELATIONSHIP: NATURAL OR FORCED?

Figure 2.5: As in Fig. 2.4, but using forced Niño 3.4 SST index.
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON - ENSO RELATIONSHIP: NATURAL OR FORCED?

(b) and rcp8.5 (c). The global SST regression to the forced Niño3.4 index is clearly a global warming pattern in all three cases (Figs. 2.5b, d, f), consistent with the S/N EOF pattern (Fig. 2.1). The corresponding rainfall regression pattern associated with the forced SST for the observation (Fig. 2.5a) displays a drying trend in the southwest and northeast and wetting over central India. The CMIP5 historical (Fig. 2.5c) shares some similarity with the observed pattern in India, in sharp contrast to that in east China, where the observed and modeled patterns are opposite to each other. Note that both the statistical significance in Fig. 2.5a and the model-agreement in Fig. 2.5c are relatively low, indicating a high degree of uncertainty for the forced response in the historical period. The rcp8.5 (Fig. 2.5e) gives a clearly enhanced rainfall trend across the Asian monsoon domain, with relatively high model-agreement for most of Asia. The wetting signal associated with SST warming in the rcp8.5 scenario has contributed to the weakening of the negative Indian monsoon pattern in the 21st century (Fig. 2.2e), given the fact that the natural component associated with ENSO generally remains unchanged (Fig. 2.4e). The enhanced monsoon rainfall associated with forced SST is consistent with the results in Chapter 3 in this thesis (Li et al. 2015), in which the radiatively forced monsoon rainfall signal is extracted directly using the S/N EOF analysis on rainfall rather than SST. Based on moisture budget analysis, Chapter 3 (Li et al. 2015) will further show that this future wetting trend is dominated by the thermodynamic contribution to the total mean moisture convergence. The 850hPa winds regressions (arrows) show distinct differences for 20CR (top panel), CMIP5 historical (middle panel) and rcp8.5 (bottom panel) for both the monsoon region and the Pacific, which may have contributed to the discrepancies in the forced rainfall responses. These circulation changes are consistent with Li et al. (2015), attributing the differences to the relative roles of aerosol and greenhouse gas (GHG) forcing during the historical period.

2.4 Summary

In this chapter, we have examined the ENSO-monsoon relationship for the 20th and 21st centuries using observations and CMIP5 model simulations. Running correlations between all-India rainfall and Niño 3.4 SST index show prominent decadal variability of the ENSO-
CHAPTER 2. RECENT AND FUTURE CHANGES IN THE ASIAN MONSOON-ENSO RELATIONSHIP: NATURAL OR FORCED?

monsoon relationship in observations. It is likely that the weakening in ENSO-monsoon relation in the recent decades is dominated by natural decadal variability rather than the global warming trend. The modeled ENSO-monsoon temporal correlation shows large inter-model spread, with the MMM significantly weaker than that for the observation. Although CMIP5 models tend to simulate well the ENSO-monsoon spatial structure when using the MMM, there is large uncertainty in the strength of the correlation within the model ensemble, ranging from slightly positive correlation in some models to strongly negative in others. In the rcp8.5 case, CMIP5 MMM shows a slightly weaker correlation than that in preindustrial and historical simulations, suggesting that the ENSO-monsoon relationship may change in the future.

We have applied S/N maximizing EOF analysis onto JJA seasonal averaged global SST of the CMIP5 ensemble to extract the externally forced signal, and separated the anthropogenically forced component from the naturally varying component of ENSO variance. Results show that the natural component of the ENSO-monsoon relationship in the observed, CMIP5 historical, and the rcp8.5 scenario simulations are very similar in their spatial structure, indicating that the unforced ENSO-monsoon relationship is insensitive to the strength of the radiative forcing. When the radiative forced component is included, on the other hand, the ENSO-monsoon relationship in observations and CMIP5 historical simulations are relatively unchanged from the unforced component, but the rcp8.5 simulations show a slightly weaker negative relationship. The results suggest that natural variability is the dominant factor in determining the ENSO-monsoon relationship in observations and CMIP5 historical simulations. In the 21st century, the wetting signal associated with SST warming may contribute to a weakening ENSO-monsoon relationship.

Our results have revealed that the negative ENSO-monsoon correlations are overall weak in CMIP5 MMM compared to observations and the variability among the models is rather large. Furthermore, there are large uncertainties in the forced rainfall response to SST warming during the historical period. It has been shown that GCMs have some difficulties in simulating and predicting regional monsoon variability and change, with aerosols being the major source of uncertainty (Turner and Annamalai 2012; Li et al. 2015). Although it is noted that CMIP5 models are more skillful than CMIP3 models (Sperber et al. 2013), there
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are still significant inter-model spreads (Wang et al. 2014). The ENSO-monsoon relationship could be a possible basis to evaluate the models and separate models into different categories for further mechanism-based studies. For example, if a model has a weak or opposite sign (positive) relationship between ENSO and Asian monsoon compared to observations, how would that difference affect the forced monsoon response to SST change? How would this type of models differ in their anthropogenically forced responses compared to the models that have stronger (negative) ENSO-monsoon relationship? Further study will be conducted to study the mechanisms contributing to the differences in the models, as well as their possible implications for future monsoon change.
Chapter 3

Mechanisms of Asian summer monsoon changes in response to anthropogenic forcing in CMIP5 models


3.1 Introduction

With its large population, increasing industrial development and severe water stresses, Asia is one of the most vulnerable regions in the world facing hydroclimate changes. Piao et al. (2010) assessed the impacts of climate change on water resources and agriculture in China, suggesting a 20% crop production decrease by 2050 under the worst-case scenario. However the overall impact is far from certain due to the high variability and uncertainty in projected climate, particularly precipitation, and the corresponding crop responses. Asian monsoon is one of the major monsoon systems in the world, with critical importance in terms
of climate impacts in Asia and globally (e.g. Rodwell and Hoskins 1996; Liu and Yanai 2001; Lin and Wu 2012). Analyzing and understanding the characteristics of monsoon change has important implications for various socioeconomic sectors and human well-being, including water resource management, agriculture, ecosystem service, food security, and public health (e.g. Kumar et al. 2004; Hong and Kim 2011; Mirza 2011).

The Asian monsoon is an interactive system influenced by both internal variability and external forcing (e.g. Lei et al. 2011; Turner and Annamalai 2012; Song et al. 2014). There have been extensive observational, modeling, as well as paleoclimate studies on Asian monsoon variability, particularly on interannual timescale, revealing strong associations with the El Niño-Southern Oscillation (e.g. Wang et al. 2000; Kumar et al. 2006; Lau and Nath 2006; Cook et al. 2010; Mishra et al. 2012). However anthropogenic factors, particularly the increasing concentration of greenhouse gases (GHGs) and changing aerosol emissions in recent decades and the future, could have profound consequences on monsoon behavior. For the forced change of Asian monsoon during the past century, several studies have addressed the weakening monsoon circulation during the second half of the 20th century, both for Indian monsoon (Annamalai et al. 2013) and East Asian monsoon (Song et al. 2014). As for future projections, modeling studies suggest intensification of future monsoon under global warming, both in mean precipitation and extreme events (Ueda et al. 2006; Seo et al. 2013; Wang et al. 2014). Understanding how the monsoon may change in the future and why the changes occur is a challenging task for hydroclimate research.

Previous studies have advanced our understanding of the physical mechanisms causing the changes of the hydrological cycle in response to global warming on a global scale. For large-scale changes, the “wet-get-wetter” or “rich-get-richer” mechanism (Held and Soden 2006) emphasizes the thermodynamic effect due to the increase in lower tropospheric water vapor in a warming atmosphere. Based on the Clausius-Clapeyron equation, the amount of saturation vapor pressure increases at a rate of about 7% per 1 K of temperature rise. Consequently, horizontal moisture transport increases within the atmosphere, leading to enhanced precipitation minus evaporation \( (P - E) \) where mean moisture converges and reduced \( P - E \) where moisture diverges. However precipitation increases are also controlled by the radiative (energy) constraints (e.g. Takahashi 2009; O’Gorman et al. 2012), making
the rate of precipitation increase less than the 7% per 1 °C. As a consequence, the atmospheric overturning circulation slows down as climate warms, especially for the Walker circulation in the tropics (Vecchi and Soden 2007). Using moisture budget analysis, Seager et al. (2010) show that the global $P - E$ change follows the “wet-get-wetter” pattern with a large part of this change coming from the thermodynamic component due to the rise in specific humidity. Seager et al. (2010) further show that in the tropics, circulation changes (dynamic contribution) offset some of the thermodynamic change because of the slowdown of the tropical divergent circulation. Based on an evaluation on tropical regional precipitation change, Chou et al. (2009) find that the thermodynamic component is a good approximation for large-scale averages, but dynamic feedback can substantially increase or decrease precipitation anomalies within the convergence zones. For spatial distribution of the tropical rainfall response, another fundamental view is the “warmer-get-wetter” mechanism (Xie et al. 2010) over the tropical ocean, suggesting that precipitation increases where the sea surface temperature (SST) warming exceeds the tropical mean and vice versa.

For the mechanisms of Asian monsoon change, rainfall is expected to increase under GHGs-induced warming mainly due to more abundant tropospheric water vapor, following the “wet-get-wetter” mechanism, accompanied by the weakening monsoon circulation (Ueda et al. 2006). Studies using the Coupled Model Intercomparison Project - Phase 3 (CMIP3) and Phase 5 (CMIP5) model projections suggest that the monsoon rainfall is likely to strengthen globally in the future due to enhanced atmospheric moisture and surface evaporation, leading to an increase in moisture convergence (Kitoh et al. 2013; Lee and Wang 2014). In the Asian monsoon region, the dynamical weakening of monsoon circulation is shown to be much lower than that of other monsoons, resulting in a larger rainfall increase (Endo and Kitoh 2014). Greenhouse warming could also induce horizontal thermal contrasts, favoring the Asian monsoon precipitation to increase: the east-west asymmetry in the sea level pressure (SLP) field generated by the “warmer land - cool ocean”; and the hemispheric SLP difference generated by the “warm northern hemisphere - cool southern hemisphere” (Wang et al. 2014). However, SST warming associated with increasing GHGs over the tropical western Pacific could affect atmospheric circulation and cause a drying trend over South Asia (Annamalai et al. 2013).
Aerosol effect is another major driver of anthropogenic climate change, particularly over Asia, considering its severe air pollution problem in recent years. The possible impacts of aerosols on rainfall have drawn much attention in recent Asian monsoon research (e.g. Bollasina et al. 2011; Ganguly et al. 2012a,b; Wu et al. 2013). Increased aerosol concentration in the atmosphere could reduce the surface solar radiation (“dimming” effect), which weakens the SST gradient in the Indian Ocean, reduces local Hadley cell circulation thereby weakening the Indian monsoon (Ramanathan et al. 2005). On the other hand, the “elevated heat pump” hypothesis (Lau et al. 2006) suggests that aerosols could enhance the meridional temperature gradient in the mid-to-upper troposphere, causing an advancement and intensification of the Indian summer monsoon rainfall.

Although there is a general consensus on the large-scale thermodynamic mechanism of hydroclimate change, the dynamical response is not well understood. Furthermore, the effects of climate change on the regional scale, especially monsoon strength and variability, are complex and uncertain (Christensen et al. 2013). The question of how Asian monsoon rainfall and circulation may respond to future anthropogenic forcing, including both aerosols and GHGs, is far from conclusive.

Observed linear trend of Asian summer monsoon rainfall (June-August mean from 1901 to 2005), however, shows large uncertainty but with a slight wetting trend for the area average (Figs. 3.1a, b). We used two available rainfall datasets for the estimate of linear trend in Fig. 3.1, the GPCC (a) and UEA CRU (b) (see section 3.2 for detailed information on these datasets), and the trends are shown only for those grid boxes where at least one rain gauge existed in any month of the June-August (JJA) season for at least 80 years. The CMIP5 multi-model simulated linear trend over the same time period presents a dominant drying signal over most of the Asian monsoon region, particularly over China and northeastern India (Fig. 3.1c). The modeled historical trend (Fig. 3.1c) shows some similarity with the observations (Figs. 3.1a, b) over India, while the discrepancy is large over eastern China where the data coverage is relatively poor. On the other hand, future projection under the representative concentration pathway 8.5 (rcp8.5) emission scenario presents a predominantly wetting trend across the entire monsoon region (Fig. 3.1d). The discrepancies between observations and model simulations as well as the contrast between
past and future changes motivate us to examine further the causal mechanisms and to explore the relative effects of aerosols and GHGs for the historical period. Given the complex nature of the radiative forcing for the 20th century with both anthropogenic aerosols and GHGs as well as other natural radiative forcing, and the large monsoon variability on interannual and decadal timescales, linear trend may not be able to accurately represent anthropogenic changes. Thus it is essential to obtain a better estimate of the forced signal
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and separate the radiatively forced component for Asian monsoon rainfall from the natural varying component. We used the signal-to-noise (S/N) maximizing empirical orthogonal function (EOF) (Ting et al. 2009) technique in this study to obtain a model-based best estimate of the radiatively forced signal in monsoon rainfall.

We examined Asian summer monsoon patterns in response to anthropogenic forcing for JJA seasonal mean using observations and the CMIP5 multi-model ensemble historical simulations under the all-forcing, aerosol-forcing and GHG-forcing scenarios and future projections under the rcp8.5 emission scenario. The main focus of this study is to investigate the possible causes of these changes for both the 20th and 21st centuries, including 1) the thermodynamic and dynamic mechanisms and 2) the relative impacts of aerosols and GHGs.

The chapter is organized as follows. Section 3.2 describes observed datasets and model simulations used in the study. Section 3.3 presents methodology of the analysis. Results regarding forced changes of Asian summer monsoon rainfall in the 20th and 21st centuries are shown in section 3.4, followed by a detailed analysis of the dynamic and thermodynamic mechanisms governing these changes in section 3.5. The relative roles of aerosols and GHGs in contributing to monsoon rainfall changes during the historical period are discussed in section 3.6. The main conclusions are summarized in section 3.7.

3.2 Data

3.2.1 Observational and reanalysis data

We used monthly data from two gridded observational datasets for precipitation: the Global Precipitation Climatology Centre (GPCC) Full Data Product version 6 from the World Climate Research Programme (WCRP) Global Climate Observing System (GCOS) (Schneider et al. 2011), and the Climate Research Unit (CRU) at the University of East Anglia (UEA) version 3.2 (Harris et al. 2014). The spatial resolution is $0.5^\circ \times 0.5^\circ$ for both datasets, interpolated to $1^\circ \times 1^\circ$ to allow for a higher rain gauge count within each grid box. Only the grid boxes where at least one rain gauge existed in any month of the JJA season for at least 80 years during the 1901-2005 period were used in the analysis in order to improve the data reliability. We used monthly data from the National Oceanic and Atmospheric
CHAPTER 3. MECHANISMS OF ASIAN SUMMER MONSOON CHANGES IN RESPONSE TO ANTHROPOGENIC FORCING IN CMIP5 MODELS

Administration (NOAA) National Climate Data Center (NCDC) Extended Reconstructed Sea Surface Temperatures (ERSST), version 3b (Smith et al. 2008) for SST and the 20th Century Reanalysis (20CR) Project version 2 (Compo et al. 2011) for specific humidity and winds, both with a $2^\circ \times 2^\circ$ spatial resolution.

3.2.2 CMIP5 model simulations

Model simulations in this study include a multi-model ensemble from the WCRP CMIP5 models (Taylor et al. 2012) data output. Based on monthly data availability including precipitation, specific humidity and wind fields, we used all realizations of 35 models for historical simulations and future projections under the high-end rcp8.5 emission scenario. Altogether 109 realizations were analyzed for the historical period and 72 for rcp8.5. For single-forcing simulations, we used all realizations of a common set of 9 models for aerosol-forcing (27 realizations) and GHG-forcing (29 realizations) experiments. In order to avoid model bias, the results were compared with historical all-forcing simulations using the same 9 models (45 realizations). Details of CMIP5 model simulations are provided in Table 3.1. All model output were interpolated to a $1^\circ \times 1^\circ$ spatial resolution for precipitation and $2^\circ \times 2^\circ$ for SST, specific humidity and winds.

3.3 Methods

3.3.1 S/N maximizing EOF analysis

We applied S/N maximizing EOF analysis (Allen and Smith 1997; Venzke et al. 1999; Chang et al. 2000) to JJA seasonal average rainfall of the CMIP5 ensemble to extract the externally forced signal, as in Ting et al. (2009). In order to focus on long-term low frequency changes, an 11-year running average was applied to the JJA seasonal mean data prior to the analysis. Given the multi-model, multi-realization ensemble, the total covariance matrix of the ensemble mean can be assumed as a sum of two linearly independent matrices, one for forced signal and one for internal variability (“climate noise”). We first determined the spatial structure of the internal modes of variability through an EOF analysis on the ?noise? matrix, formed using the second century of rainfall anomalies in the preindustrial control
Table 3.1: List of CMIP5 models and the number of realizations used in Chapter 3. Asterisks denote the common set of 9 models examined for single-forcing simulations.

<table>
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<th>Model</th>
<th>historical</th>
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<td>1</td>
<td>1</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>1</td>
<td>-</td>
<td>-</td>
</tr>
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<td>1</td>
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<td>-</td>
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run for each of the corresponding CMIP5 model in the analysis. These noise EOFs (retaining 80% of the total variance) were used to form a spatial pre-whitening transformation matrix to filter the internal variability that was not removed by multi-model ensemble averaging, so that the spatial covariance in the ensemble average is largely due to forced change. The leading EOF mode of the ensemble mean gives the dominant forced signal. Using the leading principle component (S/N PC1) as the forced index, we construct the forced patterns of rainfall, moisture, circulation, as well as moisture budget fields using regression analysis. In obtaining the multi-model mean (MMM) in this study, we first computed the ensemble mean for each model, and then performed multi-model ensemble average using each model’s ensemble mean. While this method may be affected by the models with fewer realizations, especially models with only single realization thus contribute towards larger amplitudes of internal fluctuations, it has a certain advantage in terms of reducing the risk of greatly biasing towards models with more ensemble members. We used all available realizations instead of using just one realization for each model because it provides a much larger ensemble set to eliminate internal variability.

### 3.3.2 Moisture budget analysis

The atmospheric moisture budget equation (Trenberth and Guillemot 1995; Seager and Henderson 2013; Seager et al. 2010, 2014) gives the balance between \( P - E \) and the convergence of the vertically integrated atmospheric moisture flux for monthly or longer time averages (Trenberth and Guillemot 1995). In pressure coordinates, the balance can be expressed as follows:

\[
P - E = -\frac{1}{g \rho_w} \nabla \cdot \int_0^{p_s} u q \; dp,
\]

where \( P \) is precipitation, \( E \) is evaporation, \( g \) is gravitational acceleration, \( \rho_w \) is the density of water, \( p \) is pressure and \( p_s \) surface pressure, \( u \) the horizontal wind vector \((u = u_i + v_j)\), \( q \) specific humidity. Overbars represent monthly mean values. The vertical integral in Eq. (1) is calculated as the sum over pressure levels, so we rewrite Eq. (1) as:

\[
P - E \approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_{k} q_{k} \Delta p_{k},
\]
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where \( k \) is the vertical level with a total of \( K \), \( \Delta p \) is the pressure thickness. Here the calculation is performed on pressure levels from 1000hPa to 200hPa. The total \( K \) is 10 levels in CMIP5 models. We neglected submonthly variations of surface pressure since this introduces no significant error (Seager and Henderson 2013).

We then denote departures from monthly means with primes, therefore

\[
\mathbf{u} = \mathbf{u} + \mathbf{u}', \quad q = q + q'.
\] (3.3)

Thus the monthly mean moisture flux can be expressed as:

\[
\mathbf{u}q = \mathbf{u}_c q + \mathbf{u}_a q'.
\] (4.4)

Separating the moisture divergence term into contributions of mean flow term and sub-monthly transient eddies, Eq. (2) can be written as:

\[
\overline{P} - \overline{E} \approx -\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_k \overline{q}_k \Delta p_k - \frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_k' \overline{q}_k' \Delta p_k.
\] (3.5)

For the Asian monsoon region, the mean moisture convergence term \(-\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_k \overline{q}_k \Delta p_k\) tends to balance well with \(\overline{P} - \overline{E}\), both in terms of climatology and changes. Therefore we focus on this term for further analysis. In order to gain more understanding of the mechanisms governing the changes in the moisture budget, we further separate the total changes in mean moisture convergence into those due to changes in specific humidity (the thermodynamic component), and those due to changes in circulation (the dynamic component), as in Seager et al. (2010, 2014):

\[
\overline{u} = \overline{u}_c + \overline{u}_a, \quad \overline{q} = \overline{q}_c + \overline{q}_a,
\] (3.6)

here we denote monthly climatological values with subscript \( c \), calculated using the time period 1900 to 1949 given the weak anthropogenic forcing during this period, and anomalies with subscript \( a \). Then the mean moisture convergence term can be derived as follows:

\[
-\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_k \overline{q}_k \Delta p_k = -\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \left( \mathbf{u}_{k,c} + \mathbf{u}_{k,a} \right) \left( \overline{q}_{k,c} + \overline{q}_{k,a} \right) \Delta p_k
\]

\[
= -\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,c} \overline{q}_{k,c} \Delta p_k - \frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,a} \overline{q}_{k,a} \Delta p_k
\] (3.7)

\[
-\frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,a} \overline{q}_{k,c} \Delta p_k - \frac{1}{g \rho w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,a} \overline{q}_{k,a} \Delta p_k.
\]
Taking the difference on both sides of Eq. (7) and ignoring the quadratic term by assuming small amplitude in perturbation quantity, the total change of the mean moisture convergence (hereafter $\delta MC$) can be approximated as follows:

$$\delta \left[ -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} u_k q_k \Delta p_k \right] \approx \delta \left[ -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} u_{k,c} q_{k,a} \Delta p_k \right] + \delta \left[ -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} u_{k,a} q_{k,c} \Delta p_k \right],$$  (3.8)

where the first term on the right-hand side represents the thermodynamic contribution (hereafter $\delta TH$), involving only changes in specific humidity $q$; and the second term represents the dynamic contribution (hereafter $\delta DY$), involving only changes in circulation $u$. In this study, the change "$\delta$" was calculated as the regression coefficient of the quantity onto the forced index (standardized S/N PC1).

### 3.4 Forced changes of Asian summer monsoon rainfall in CMIP5 models and observations

#### 3.4.1 S/N maximizing EOF1 and PC1

The S/N maximizing EOF analysis was performed on the 11-year running mean JJA seasonal mean land precipitation for the Asian monsoon region ($5^\circ$N-55$^\circ$N, 60$^\circ$E-150$^\circ$E) in the CMIP5 model ensemble. We used the time period 1901-2000 for historical and 2011-2094 for rcp8.5. The spatial structures and principle components of the leading S/N EOF modes are given in Fig. 3.2. For the historical period, the dominant signal is a drying trend across South and East Asian monsoon domains (Fig. 3.2a), explaining 47% of the total variance. The S/N PC1 (Fig. 3.2c) shows there is a sharp increasing trend of the drying signal from 1940 to the late 1960s followed by a relatively flat period afterwards. For rcp8.5, there is a clear wetting signal over the entire region (Fig. 3.2b) with an almost linear trend throughout the 21st century (Fig. 3.2d). This first mode explains a high percentage (83%) of the total variance, indicating that the externally forced variability of Asian summer monsoon rainfall in the future can be well represented by this uniformly wetting trend.
Figure 3.2: First EOF mode of S/N maximizing EOF analysis of 11-year running mean filtered JJA precipitation for CMIP5 historical 1901-2000 (left) and rcp8.5 2011-2094 (right) simulations. (a, b) Spatial structures and (c, d) standardized leading principle components (S/N PC1).

3.4.2 Forced 20th and 21st century rainfall changes and associated SST patterns

Figure 3.3 shows the observed (a) and model simulated (b) Asian summer (JJA) monsoon rainfall regressed onto standardized S/N PC1 using UEA CRU and CMIP5 historical ensemble, respectively, and CMIP5 rcp8.5 simulations (c). We chose the CRU data due to its relatively high station coverage with continuous rain gauge observations as shown in Fig. 3.1b. Note that the color scales are different in Fig. 3.3 for observations, CMIP5 MMM
Figure 3.3: Regressions of JJA precipitation (a, c, e, units are $\text{mm day}^{-1}$) and global SST (b, d, f, units are $^\circ\text{C}$) onto standardized S/N PC1 for observations 1901-2000 (a: CRU; b: ERSST), CMIP5 35-model MMM historical 1901-2000 (c, d), and rcp8.5 2011-2094 (e, f) simulations. Stippling denotes 5% significance based on 2-sided Student’s t test in (a, b) and 26/35 model-agreement in (c-f).
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historical and rcp8.5 future simulations in order to bring out more detailed structures in each case. While the observed regression associated with S/N PC1 (Fig. 3.3a) shows some similarities to the linear trend in Fig. 3.1b, it does indicate more drying trend overall, particularly for the region between 20°N and 35°N, than that shown in the linear trend. The CMIP5 historical MMM (Fig. 3.3c), on the other hand, indicates a predominantly drying trend expanding from eastern China to northern India. The discrepancy between model simulations and observations lies in the regions south of 20°N and north of 35°N along the east coasts of Asia continent, where observations indicate a wetting trend and models showing mostly drying. It is not clear whether this discrepancy between model and observations is due to uncertainty in observations or model deficiencies. As for future model projected change under the rcp8.5 scenario (Fig. 3.3e), monsoon rainfall enhances across the entire Asian domain with relatively high model agreement.

The associated SST patterns (Figs. 3.3b, d, f), calculated as regressions of global SST onto the forced rainfall indices (standardized S/N PC1) display clear global warming trends in models and observations, aside from a strong cooling in northern North Atlantic in observations. There are, however, subtle differences in spatial patterns of SST in Fig. 3.3. As an example, the weak cooling off the east coast of China and Japan in Fig. 3.3d is not present in Figs. 3.3b and 3.3f. How much of the regional SST features may have contributed to the differences in rainfall patterns in Fig. 3.3 will be explored in future studies. However, the overall agreement in SST patterns between models and observations in Fig. 3.3 suggests that the forced SST pattern may not be the dominant driver of the differences between modeled and observed historical precipitation trends or between modeled past and future precipitation trends.

3.5 Thermodynamic and dynamic mechanisms of 20th and 21st century Asian summer monsoon changes

3.5.1 Changes in moisture content and monsoon circulation

To explore the physical processes leading to the monsoon rainfall change in CMIP5 models, we examined moisture and circulation changes using 850hPa specific humidity and
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horizontal wind fields, as shown in Fig. 3.4. In all cases, the lower troposphere moistens up as the temperature increases (Figs. 3.4a, c, e). The monsoon circulation intensifies in 20CR (Fig. 3.4b) while CMIP5 models indicate reduced summer monsoon circulation for both Indian and East Asian monsoon in historical simulations (Fig. 3.4d). In the 21st century rcp8.5 projections (Fig. 3.4f), Indian monsoon circulation tends to be shifted northward with a strong decrease in the south while East Asian monsoon circulation enhances.

The results in Fig. 3.4 indicate that the forced monsoon rainfall change is associated with a general increase in atmospheric moisture content for historical observations and model simulations, as well as for the future climate, whereas the monsoon circulation change differs in each of the three cases. This suggests that the differences in forced monsoon rainfall changes in Fig. 3.3 may have been caused by the discrepancies in the circulation responses. In observations, the increase of moisture and intensification of circulation both contribute towards the enhancement of monsoon rainfall. In historical model simulations, the weakened circulation may have dominated over the increased moisture content, resulting in an overall drying. However in the 21st century, increasing water vapor due to greenhouse warming is much stronger than in the historical period (note that the color scale in Fig. 3.4e is an order of magnitude larger than in Fig. 3.4c), leading to greatly enhanced summer monsoon rainfall. These mechanisms will be further explored in the next subsection using moisture budget analysis within the CMIP5 modeling framework.

3.5.2 Changes in moisture budget

Figure 3.5 shows the regressions of $P - E$ (a, b), the mean moisture convergence (c, d), the thermodynamic (e, f), and the dynamic (g, h) components of the mean moisture convergence, onto the standardized S/N PC1 for CMIP5 historical (left column) and rcp8.5 (right column) MMM. For both the historical and rcp8.5 model simulations, there is a good agreement between changes in $P - E$ and the mean moisture convergence (top two panels in Fig. 3.5), indicating a relatively minor contribution due to transient moisture convergence in the monsoon region. During the historical period in CMIP5, $\delta P - \delta E$ (Fig. 3.5a) and the total $\delta MC$ (Fig. 3.5c) display a general drying pattern over the monsoon domain, consistent with the corresponding precipitation regression (Fig. 3.3c). There are
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Figure 3.4: Regressions of JJA specific humidity (a, c, e, units are g kg\(^{-1}\)) and wind (b, d, f, units are m s\(^{-1}\)) at 850hPa onto standardized rainfall S/N PC1 for 20CR 1901-2000 (a, b), CMIP5 35-model MMM historical 1901-2000 (c, d), and rcp8.5 2011-2094 (e, f) simulations. In b, d, and f, arrows are vectors of regression coefficients of \(u\); colors show the regression coefficients of wind velocity. Stippling denotes 5% significance based on 2-sided Student’s t test in (a, b) and 26/35 model-agreement in (c-f).
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Figure 3.5: Regressions of JJA $\overline{P} - \overline{E}$ (a, b), the total mean moisture convergence (c, d), the thermodynamic component (e, f) and the dynamic component (g, h) onto standardized rainfall S/N PC1 for CMIP5 35-model MMM historical 1901-2000 (left column) and rcp8.5 2011-2094 (right column). Units of the moisture convergence terms are $\text{mm day}^{-1}$. Stippling denotes 26/35 model-agreement.

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general decreases in evaporation (not shown), resulting in a less negative, and in some area, a slight positive $\overline{P} - \overline{E}$ change in Fig. 3.5a. The thermodynamic contribution to the mean moisture convergence change, $\delta TH$ (Fig. 3.5e), shows increasing moisture convergence while the dynamic contribution, $\delta DY$ (Fig. 3.5g), shows a strong decrease in moisture convergence. The overall decrease in mean moisture convergence in Fig. 3.5c indicates that the dynamic contribution (Fig. 3.5g) dominates over the thermodynamic contribution (Fig. 3.5e) during the historical period. Thus the drying trend in the 20th century is mainly due to the weakening monsoon circulation. For the future using rcp8.5 scenario, on the other hand, the increase in mean moisture convergence (Fig. 3.5d) is dominated by the increasing moisture convergence in the thermodynamic contribution (Fig. 3.5f), which shows a general increase in the Asian monsoon region, consistent with the “rich-get-richer” mechanism (Held and Soden 2006). The dynamic contribution (Fig. 3.5h) shows a weakened moisture convergence over the Asian monsoon region, indicating anomalous mass divergence in the region. The contributions of the dynamic component (Figs. 3.5g, h) are supported by the forced patterns of the 500hPa vertical pressure velocity ($\omega$) field (not shown), with anomalous decending motion in the monsoon region for both the historical period and rcp8.5. In summary, the discrepancies between the drying trend in the historical simulations and the wetting trend in the rcp8.5 future scenario can be explained by the relative importance of dynamical and thermodynamical contributions to the total mean moisture convergence. While thermodynamic mechanism dominates in the future due to enhanced warming, the historical monsoon rainfall changes are dominated by the changes in monsoon circulation.

To further quantify the relative contributions of $\delta TH$ and $\delta DY$, we examined the temporal evolution of the thermodynamic ($-\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,c} \mathbf{q}_{k,a} \Delta p_k$) and dynamic ($-\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,a} \mathbf{q}_{k,c} \Delta p_k$) terms in Eq. (7), averaged over land area of the major monsoon domain, 5°N-35°N, 70°E-122°E. The anomalies were calculated with respect to 1900-1949 climatology. The MMM results for the 35 CMIP5 models are shown in Fig. 3.6 for historical 1900-2005 (a) and rcp 8.5 2006-2099 (b). An 11-year running mean was applied to smooth the time series in order to emphasize the long-term trend. During the historical period, the dynamic component (blue line) exhibits a downward trend while the thermodynamic
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Figure 3.6: Time series of area averaged anomalies of JJA $P - E$ (green), the total mean moisture convergence (black), the thermodynamic component (red), and the dynamic component (blue) over land area between 5°N-35°N, 70°E-122°E for CMIP5 35-model MMM historical 1900-2005 (a) and rcp8.5 2006-2099 (b). The anomalies are calculated with respect to 1900-1949 climatology. The time series are smoothed with an 11-year running average. Units are mm day$^{-1}$. 

The evolution of the total mean moisture convergence anomaly (black line) is dominated by the dynamical weakening until the late 60s, followed by a reversed trend due to the thermodynamic contribution from 1970 onwards. For rcp8.5, the total mean moisture convergence follows the strong upward trend of the thermodynamic contribution with little change in the dynamic component of the moisture convergence. In both cases, the total mean moisture convergence (black line) balances well with the $P - E$ (green line), with minor departures due to the transient eddy contribution. It is clear that in the future, the total mean moisture convergence, thus $P - E$ is predominantly controlled by the thermodynamic “wet-get-wetter” mechanism (Held and Soden 2006): the significant increase of water vapor content in the atmosphere associated with greenhouse warming leads to an enhanced hydrological cycle and the increase of monsoon rainfall. Previous studies have addressed the weakening tropical circulation as a robust response to a warmer climate (Vecchi and Soden 2007; Chadwick et al. 2013). Our results here indicate that the reduction of the dynamic component is small in the 21st century, which is consistent with the findings of Endo and Kitoh (2014), suggesting that the dynamical weakening of
the Asian monsoon is less than that of other monsoons in future projections using both CMIP3 and CMIP5 models possibly due to its distinctive geographical characteristics. It is interesting to note the reversal of the historical moisture convergence trend starting in the early 70s in Fig. 3.6a, which coincides well with the flattening of the drying trend in the S/N PC1 in Fig. 3.2c. The strong drying trend between 1940 and 1970 in Fig. 3.2c can be largely attributed to the relatively weak increase in thermodynamic component but a large decrease in the dynamic component in Fig. 3.6a, which will be discussed in more detail in section 3.6.

### 3.6 Attributing historical changes to the relative roles of aerosols and GHGs

#### 3.6.1 20th century rainfall changes forced by aerosols and GHGs

To examine the relative roles of aerosol and GHGs on monsoon change during the historical period, we further applied S/N maximizing EOF analysis to JJA rainfall on CMIP5 aerosol-only and GHG-only forcing runs using a common set of 9 models (see Table 3.1). The dominant mode in the all-forcing case using the 9 models explains 49% of the total variance as compared to 47% in the case of using all 35 models (Fig. 3.2c). For the aerosol-only and GHG-only forcing cases, the first S/N EOF mode explains 54% and 34% of the total variance, respectively.

Regressions of rainfall onto the standardized S/N PC1 under all-forcing (a), aerosol-forcing (c) and GHG-forcing (e) are shown in Fig. 3.7, as well as the standardized S/N PC1 indices (b, d, f). As illustrated by the regression plots, rainfall reduces under aerosol forcing (Fig. 3.7c) and increases under GHG forcing (Fig. 3.7e). The principle components display clear upward trends throughout the 20th century. The total change (Fig. 3.7a) depends on the relative strengths of these two competing effects. Figure 3.7a suggests that aerosol forcing dominates over the greenhouse effect for the historical period and leads to the general drying trend in CMIP5’s historical all-forcing simulations.
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Figure 3.7: Regressions (a, c, e) of JJA precipitation (units are mm day$^{-1}$) onto standardized rainfall S/N PC1 (b, d, f) from 1901 to 2000 for CMIP5 9-model MMM historical all-forcing (a, b), aerosol-forcing (c, d), and GHG-forcing (e, f) simulations. Stippling denotes 7/9 model-agreement.
3.6.2 Changes in moisture content and monsoon circulation

We also examined 850hPa specific humidity and wind fields to evaluate the thermodynamic and dynamic changes, as shown in Fig. 3.8. For moisture, lower tropospheric water vapor reduces under aerosol effect (Fig. 3.8c) while enhancing greatly under greenhouse warming (Fig. 3.8e) as expected. The overall change in lower tropospheric water vapor in the all-forcing scenario (Fig. 3.8a) is dominated by greenhouse warming. For the wind field, aerosols weaken the low-level monsoon circulation (Fig. 3.8d) while greenhouse effect shifts Indian monsoon circulation northward and enhances East Asian monsoon circulation (Fig. 3.8f), similar to that for the rcp8.5 scenario in Fig. 3.4f. In this case, aerosol effect clearly dominates the monsoon wind change, resulting in weakened total monsoon circulation (Fig. 3.8b).

The roles of aerosol forcing and GHG forcing were further examined by taking the differences between all-forcing and GHG-forcing and all-forcing and aerosol-forcing simulations for the late 20th century (1971-2005), respectively. The resulting monsoon rainfall and circulation differences (not shown) are very similar to that in Figs. 3.7 and 3.8, confirming the robustness of the S/N EOF method for estimating the forced signals.

3.6.3 Changes in moisture budget

Figure 3.9 shows the spatial distribution of changes in moisture budget terms under all-forcing, aerosol-forcing and GHG-forcing scenarios. The all-forcing (left column in Fig. 3.9) results are in gross agreement with that in Figs. 3.5c, 3.5e, and 3.5g using a large set (35) of CMIP5 historical simulations, indicating that the results are relatively robust. Under aerosol forcing (middle column in Fig. 3.9), $\delta\text{TH}$ (Fig. 3.9e) and $\delta\text{DY}$ (Fig. 3.9h) both show decreased moisture convergence over the monsoon domain, leading to the much reduced mean moisture convergence (Fig. 3.9b). The dynamical weakening is particularly strong over India, Myanmar, as well as eastern China, while $\delta\text{TH}$ is particularly strong over central China. Under GHG forcing (right column in Fig. 3.9), $\delta\text{TH}$ (Fig. 3.9f) shows clearly enhanced moisture convergence across the monsoon domain, while $\delta\text{DY}$ (Fig. 3.9i) indicates dynamical weakening over most parts of southern China, India and Indo-China regions. The total $\delta\text{MC}$ is predominantly driven by the thermodynamic enhancement,
CHAPTER 3. MECHANISMS OF ASIAN SUMMER MONSOON CHANGES IN RESPONSE TO ANTHROPOGENIC FORCING IN CMIP5 MODELS

Figure 3.8: As in Fig. 3.4, but for CMIP5 9-model MMM historical all-forcing (a, b), aerosol-forcing (c, d), and GHG-forcing (e, f) simulations, from 1901 to 2000. Stippling denotes 7/9 model-agreement.
resulting in an overall strong convergence pattern (Fig. 3.9c). Consistent with Fig. 3.8, the thermodynamic change of mean moisture convergence in the all-forcing case (Fig. 3.9d) is dominated by the GHG forcing, while the dynamic change in mean moisture convergence in the all-forcing case (Fig. 3.9g) is dominated by the aerosol forcing during the historical period.

One interesting feature in Fig. 3.9 is that the responses of the mean moisture convergence to aerosol forcing (Fig. 3.9b) and GHG forcing (Fig. 3.9c) over the western tropical Pacific
CHAPTER 3. MECHANISMS OF ASIAN SUMMER MONSOON CHANGES IN RESPONSE TO ANTHROPOGENIC FORCING IN CMIP5 MODELS

and the Indian Ocean are very similar in spatial distribution but with opposite sign. Under aerosol (GHG) forcing, there is increased convergence (divergence) to the south of the equator over the Indian Ocean, and increased divergence (convergence) at the Arabian Sea, Bay of Bengal, as well as most regions of the western Pacific. This is consistent with the findings of Xie et al. (2013), suggesting that the climate responses to aerosols and GHGs for both precipitation and SST are spatially similar and opposite in sign over the ocean, possibly due to the similar ocean-atmospheric feedbacks. However, our results indicate that the detailed mechanisms contributing to these opposite patterns in Figs. 3.9b and 3.9c may be quite different, with dynamic contribution dominates in the aerosol-forcing case, and thermodynamic contribution dominates in the GHG-forcing case.

The relative importance of aerosols and GHGs was further examined by showing the temporal evolution of the area-averaged (land domain between 5°N and 35°N, 70°E and 122°E) thermodynamic \((-\frac{1}{g\rho_{w}}\nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,c} \mathbf{q}_{k,a} \Delta p_{k})\) and dynamic \((-\frac{1}{g\rho_{w}}\nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,a} \mathbf{q}_{k,c} \Delta p_{k})\) terms under all-forcing, aerosol-forcing and GHG-forcing scenarios. The all-forcing (Fig. 3.10a) temporal evolution of the three moisture convergence terms are similar to that in Fig. 3.6a for 35 models, but showing here for the 9 models for direct comparison to the aerosol-only and GHG-only results. The aerosol-only case (Fig. 3.10b) shows a decreasing moisture convergence in both dynamic (blue) and thermodynamic (red) terms, resulting a sharp decrease in the total mean moisture convergence (black). On the other hand, the thermodynamic component features a robust rising trend of moisture convergence under GHG forcing (red line in Fig. 3.10c), and a gradual decrease in the dynamic component (blue), resulting in a large increase in the total mean moisture convergence (black). It is suggested that the tropical meridional overturning circulation may slow down to compensate for the energy imbalance between the northern and southern hemispheres induced by the hemispherically asymmetric anthropogenic aerosol emissions (Bollasina et al. 2011). The exact reason why there are large changes in atmospheric circulation (dynamic changes) due to anthropogenic aerosols as compared to the GHG forcing (Figs. 3.10b, c) will be explored in future studies using atmospheric general circulation model (AGCM) experiments.

The results in Figs. 3.9 and 3.10 confirm that aerosol forcing dominates the drying trend in the CMIP5 historical simulations through both dynamic and thermodynamic contribu-
Figure 3.10: Time series of area averaged anomalies of JJA total mean moisture convergence (black), the thermodynamic component (red), and the dynamic component (blue) over land area between 5°N-35°N, 70°E-122°E for CMIP5 9-model MMM historical all-forcing (a), aerosol-forcing (b) and GHG-forcing (c) simulations, from 1900 to 2005. The anomalies are calculated with respect to 1900-1949 climatology, smoothed with an 11-year running average. Units are mm day\(^{-1}\).
CHAPTER 3. MECHANISMS OF ASIAN SUMMER MONSOON CHANGES IN RESPONSE TO ANTHROPOGENIC FORCING IN CMIP5 MODELS

tions. Given the discrepancies between observed and modeled historical monsoon rainfall trends, this raises the question of whether aerosol effect is appropriately represented in the CMIP5 models given the uncertainty in aerosol forcing as well as the aerosol indirect effect. It also questions whether the rcp8.5 scenario, which gradually phases out the aerosol forcing, is representative of the future aerosol emission in the Asian monsoon region and how this uncertainty may affect the monsoon rainfall future projections. Nevertheless, it is clear that reducing air pollution in the Asian monsoon region not only has the benefit of clean air, but also can potentially levitate the drought threats over the monsoon region in the near future.

3.7 Summary

In this chapter, using S/N maximizing EOF analysis on the CMIP5 multi-model, multi-realization simulations, we extracted a model-based best estimate of the externally forced signal for Asian summer monsoon rainfall. The 20th and 21st century changes of Asian summer monsoon in response to anthropogenic forcing were examined using both observational data and CMIP5 model simulations. Results show that in the 20th century, CMIP5 models indicate a predominantly drying pattern expanding from eastern China to northern India. However there are significant discrepancies when compared to the observed pattern, which shows large spatial inhomogeneity. For the 21st century under the rcp8.5 scenario, monsoon rainfall enhances across the entire Asian domain.

We examined the thermodynamic and dynamic mechanisms causing these changes for both the historical and rcp8.5 using low-level specific humidity, wind, as well as the moisture budget analysis. Results reveal that the discrepancies between the drying trend in the CMIP5 historical simulations and the wetting trend in the rcp8.5 projections can be explained by the relative importance of dynamical and thermodynamical contributions to the total mean moisture convergence. While thermodynamic mechanism dominates in the future, the historical monsoon rainfall changes are dominated by the changes in monsoon circulation.

We further assessed the relative contributions of aerosols and GHGs on the 20th century
monsoon change. Rainfall reduces under aerosol forcing and increases under GHG forcing, thus the total change depends strongly on the relative strengths of these two competing effects. During the historical period, aerosol forcing dominates over the greenhouse effect, leading to the general drying trend in CMIP5’s historical all-forcing simulations. The thermodynamic change of mean moisture convergence in the all-forcing case is dominated by the GHG forcing, while the dynamic change in mean moisture convergence in the all-forcing case is dominated by the aerosol forcing during the historical period. Aerosol forcing dominates the drying trend in the CMIP5 historical simulations through both dynamic and thermodynamic contributions. It suggests that in the near future, air pollution control policies will not only have the benefit of clean air, but also potentially levitate the drought threats over the monsoon region.

Aerosol effect is one of the major uncertainties for climate models in simulating monsoon characteristics (Turner and Annamalai 2012). Our results here indicate that the modeled Asian monsoon change is largely dominated by the anthropogenic aerosols in the historical period. Thus discrepancies in aerosol forcing between model and observations as well as among individual models may have contributed largely to differences in forced monsoon trends. In addition, the Asian monsoon region is poorly covered by station observations during the early part of the 20th century, which leads to uncertainty in the observed pattern, especially over eastern China. Further modeling and understanding of the range of uncertainties associated with aerosol forcing and its impact on Asian monsoon changes is imperative for better hydroclimate projection in Asia.
Chapter 4

Understanding the Asian summer monsoon response to greenhouse warming: the relative roles of direct radiative forcing and sea surface temperature change


4.1 Introduction

It is now widely accepted that the global hydrological cycle will become more intensified in a warmer climate, as a consequence of the increase in tropospheric water vapor following the Clausius-Clapeyron relationship, leading to the so-called “wet-get-wetter, dry-get-dryer” pattern of change (Held and Soden 2006). Because of energetic constraints (Takahashi 2009; O’Gorman et al. 2012), the rate of precipitation increase is less than the rate of water
vapor, and tropical atmospheric circulation weakens as climate warms (Held and Soden 2006; Vecchi and Soden 2007). However on the regional scale, hydroclimate projections from state-of-the-art climate models show large uncertainty and model spread, particularly in the tropics and over the monsoon regions (Turner and Annamalai 2012; Christensen et al. 2013). The “warmer-get-wetter” mechanism has been proposed to explain the spatial distribution of rainfall change in the tropics, relating to sea surface temperature (SST) pattern (Xie et al. 2010; Chadwick et al. 2013; Ma and Xie 2013). Kent et al. (2015) find that the uncertainty of regional precipitation change in the tropics is predominantly related to spatial shifts in convection and convergence, associated with SST pattern and land-sea thermal contrast changes. Using a set of atmospheric general circulation models (AGCMs), He et al. (2014) show that SST pattern is not the dominated factor of atmospheric circulation and precipitation change, with most of its effects confined to equatorial oceans.

The Asian summer monsoon precipitation is projected to enhance under greenhouse warming, dominated by the “wet-get-wetter” thermodynamic mechanism (Kamae et al. 2014b; Wang et al. 2014; Li et al. 2015). On the other hand, dynamical changes related to atmospheric circulation are relatively weak with a low model-agreement (Endo and Kitoh 2014; Li et al. 2015). As highlighted by Xie et al. (2015), atmospheric circulation is the major source of uncertainty in regional rainfall projection. Thus the weak and diverging monsoon circulation response among the models may contribute largely to the uncertainty in monsoon rainfall projections, which is further complicated by other factors such as natural variability (Li and Ting 2015) and aerosol effects (Lau et al. 2006; Bollasina et al. 2011; Li et al. 2015).

The response of the climate system to rising greenhouse gases (GHGs) can be through both direct radiative effect and indirect effect via SST change. The direct radiative effect of CO$_2$ represents a fast adjustment of the atmosphere before surface warming occurs, and the indirect effect refers to the slow component induced by subsequent SST warming. From the perspective of the perturbation energy budget of the troposphere (Mitchell et al. 1987; Allen and Ingram 2002): without substantial changes in surface temperature, the fast response associated with increasing CO$_2$ causes a net decrease in radiative cooling and thus reduces the intensity of the hydrological cycle; the slow response due to tropospheric
warming, however, causes enhanced radiative cooling that scales approximately with surface temperature change. Thus in climate model simulations, the relative importance of the fast and slow responses may result in discrepancies in how hydroclimate responds to GHG warming.

Recent studies have further shown that the direct radiative forcing of CO$_2$ and SST warming may cause different responses in tropical circulation (Ma et al. 2012; He and Soden 2015a), summertime Pacific anticyclone and the Asian monsoon cyclone (Shaw and Voigt 2015), and midlatitude jets (Grise and Polvani 2014). For tropical rainfall and circulation, Bony et al. (2013) find that a large fraction of the long-term regional precipitation change can be explained by the direct atmospheric radiative response that occurs shortly after an abrupt CO$_2$ increase, independent of surface warming. On the contrary, Chadwick et al. (2014) argue that the fast dynamical precipitation response as shown in Bony et al. (2013) is dominated by surface warming patterns rather than the direct radiative effect. Several studies have shown that the direct radiative forcing and SST change exert different effects onto the land-sea thermal contrast changes, which then influence the atmospheric thermodynamic structures and circulation patterns (e.g. Joshi et al. 2008; Kamae et al. 2014a). For example, Shaw and Voigt (2015) find significant compensating effects of the two on the summertime Asian monsoon cyclone, Pacific anticyclone and Pacific jet stream, associated with the opposite responses in land-sea equivalent potential temperature contrasts.

Regional climate information under global warming is urgently needed for climate adaptation and socio-economic planning in various sectors, such as water resources, agriculture, and public health, particularly for the densely populated Asian monsoon regions (e.g. Kumar et al. 2004). However reliable predictive information remains a challenge due to the large uncertainty and discrepancy in climate model projections. On one hand, the current generation of coupled general circulation models (CGCMs) still have difficulty in simulating the present-day monsoon climatology (Turner and Annamalai 2012). On the other hand, the often-compensating effects of multiple processes may cause a weak total response and discrepancy among models. It is important to untangle the different physical pathways by which anthropogenic forcing may impact regional hydroclimate and determine their contribution to model uncertainties. Idealized AGCM experiments with prescribed SSTs have
been proved to be a useful tool in decomposing the different components (e.g. Mitchell 1983; Mitchell et al. 1987; Hansen et al. 1997; Deser and Phillips 2009; Bala et al. 2010).

In this chapter, we examine the hydroclimate response to rising GHGs with a focus on the Asian summer (June-August seasonal mean, JJA) monsoon. We utilize a set of climate model simulations including coupled model projections for the 21st century and idealized atmosphere-only climate change experiments with prescribed atmospheric conditions and SSTs. We aim to distinguish the relative roles of atmospheric radiative forcing and ocean-atmosphere interactions, and address the arising uncertainties in model projections. While the recent work by Shaw and Voigt (2015) demonstrates the opposing dynamical responses (defined using the dynamic component of the moisture flux convergence) of the two over Asia and the Pacific, we provide a detailed quantification of the changes for the South and East Asian monsoon regions. We also emphasize the opposing effect of the thermodynamic and dynamic components of the moisture flux convergence response to SST warming, which Shaw and Voigt (2015) presented in one of their supplementary figures without any discussion. We further present the possible contributions of the uncertainty and model spread within the context of coupled GCMs (CMIP5 models). The model simulations and methodology are described in section 2. Section 3 presents the precipitation change forced by rising CO$_2$, separating into the relative effects of direct radiative forcing and SST warming. Section 4 provides a detailed analysis on the thermodynamic and dynamic mechanisms. In section 5, we discuss the uncertainties in future monsoon projections. The main conclusions are summarized in section 6.

4.2 Data and methods

4.2.1 Coupled model simulations and idealized experiments

To examine the projected hydroclimate change in response to greenhouse warming, we used of a set of model simulations including CGCMs and idealized AGCM experiments. For coupled model simulations, we used monthly output from 35 Coupled Model Intercomparison Project - Phase 5 (CMIP5) models (Taylor et al. 2012) under the high-end representative concentration pathway 8.5 (rcp8.5) emission scenario. All available realizations with suffi-
cient variables were analyzed, with a total of 71. To separate the total response into fast and slow components related to direct atmospheric radiative effect and SST warming, respectively, we used outputs from the Atmospheric Model Intercomparison Project (AMIP) experiments. As part of the CMIP5 archive, the AMIP simulations are idealized AGCM experiments with prescribed SST and sea ice concentration. The following experiments were analyzed: 1) the control simulation (CTRL, called “amip” in the CMIP5 archive), run with observed SST and sea ice concentration from 1979 to 2008; 2) quadrupling CO₂ radiative forcing experiment (4 × CO₂, “amip4 × CO₂” in the CMIP5 archive), same SST and sea ice as CTRL, but with quadrupled atmospheric CO₂ concentration; 3) uniform 4K warming experiment (+4K, “amip4K” in the CMIP5 archive), same CO₂ concentration as CTRL, but adding a uniform +4K SST anomaly globally. The fast (slow) response is quantified as the difference of the 30-year climatology between 4 × CO₂ (+4K) and CTRL.

The AMIP simulations also provide another set of experiments using the SST pattern derived from CMIP3 models under the A1B scenario at the end of the 21st century (“amip-Future”), as compared to a uniform warming in the +4K experiment. Results show that although there are regional differences between the two over the Indian Ocean, southwestern India and the Indonesian Seas, the large-scale responses do not exhibit significant differences (not shown). The choice to use “amip4K” is to emphasize the larger contribution from uniform warming as compared to the SST spatial structure, while using “amipFuture” the two components cannot be clearly distinguished.

The use of atmosphere-only model experiments has its caveats due to the highly idealized design settings such as the obvious lack of coupling with the ocean. The direct radiative effect of increasing GHGs is closely coupled to the SST warming, thus only accounting for the radiative effect of the GHG increase through land warming while no ocean warming is allowed is artificial, as addressed by other similar studies (e.g. Deser and Phillips 2009). Nevertheless, the time taken for the ocean surface to warm up is longer than that for the land surface, and the direct effect here can be treated as a fast response to a switched-on CO₂ quadrupling. This type of analysis has been proven to be useful in determining the underlying physical mechanisms in several previous studies (e.g. Grise and Polvani 2014; He and Soden 2015a; Shaw and Voigt 2015). He and Soden (2015b) further show that
anthropogenic climate change in CGCMs can be well reproduced in AGCMs, thus lend further support of using the AGCM approach.

The AMIP experiments are available for 11 out of the 35 models used for rcp8.5 simulations in this study, with monthly data for 30 years in length. We used the first realization of 10 AGCMs (FGOALS-g2 was not used due to its unrealistic simulation of climatological monsoon rainfall, see Chapter 1). Details of the model simulations are provided in Table 4.1. All model outputs were interpolated to a $1^\circ \times 1^\circ$ spatial resolution in the land-only calculations to allow a better representation of the coastlines, and a $2^\circ \times 2^\circ$ spatial resolution to illustrate the spatial patterns on a global scale (a comparison shows that the change of resolution has little effect on the results).

### 4.2.2 Moisture budget analysis

We use the atmospheric moisture budget equation (Trenberth and Guillemot 1995) to analyze the changes in the hydrological cycle, following Chapter 3 (Li et al. 2015). In steady state, precipitation minus evaporation ($P - E$) balances the convergence of the vertically integrated atmospheric moisture flux, which can be expressed in pressure coordinates as follows:

$$P - E = -\frac{1}{g \rho_w} \nabla \cdot \int_{p_s}^{1000} u q \, dp,$$

where $P$ is precipitation, $E$ is evaporation, $g$ is gravitational acceleration, $\rho_w$ is the density of water, $p$ is pressure and $p_s$ surface pressure, $u$ is the horizontal wind vector ($u = u i + v j$), $q$ is specific humidity. Overbars represent monthly mean values. The vertical integral in Eq. (1) is calculated as the sum over pressure levels, so we rewrite Eq. (1) as:

$$P - E \approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_k q_k \Delta p_k,$$

where $k$ is the vertical level with a total of $K$, $\Delta p$ is the pressure thickness. Here the calculation is performed on 10 pressure levels from 1000hPa to 200hPa. We neglect sub-monthly variations of surface pressure since this introduces no significant error (Seager and Henderson 2013).

We then denote departures from monthly means with primes:

$$u = \bar{u} + u', \quad q = \bar{q} + q',$$
CHAPTER 4. UNDERSTANDING THE ASIAN SUMMER MONSOON RESPONSE TO GREENHOUSE WARMING: THE RELATIVE ROLES OF DIRECT RADIATIVE FORCING AND SEA SURFACE TEMPERATURE CHANGE

Table 4.1: List of CMIP5 and AMIP models, scenarios, and the number of realizations used in Chapter 4.

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and the monthly mean moisture flux can be expressed as:

$$ \overline{uq} = \overline{u}\overline{q} + \overline{u'}\overline{q'}. \quad (4.4) $$

Separating the moisture flux convergence term into contributions of mean moisture convergence (MC) and the sub-monthly transient eddies (TE), Eq. (2) can be written as:

$$ P - E \approx -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_kq_k} \Delta p_k - \frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u'_kq'_k} \Delta p_k. \quad (4.5) $$

Due to limitations in daily data availability for most of the AMIP models, the transient eddy component is approximated using the difference between $P - E$ and the mean moisture convergence, thus includes a residual term. Seager and Henderson (2013) have shown that the error arising from closing the moisture budget equation, i.e., differences between $P - E$ and the vertically integrated moisture convergence is small and thus can be neglected. Furthermore, since the response to the given forcing is quantified as the climatological difference between the forced and control simulations, the residual term is largely eliminated.

To quantify the effect of the given forcing (4 $\times$ CO$_2$ or +4K), we introduce a second overbar to denote the 30-year climatological mean, and a hat above an overbar to denote the departure of the monthly mean from the climatological value:

$$ \overline{\mathbf{u}} = \overline{\mathbf{u}} + \hat{\mathbf{u}}, \quad \overline{q} = \overline{q} + \hat{q}. \quad (4.6) $$

We also define

$$ \delta(.) = (.)_F - (.)_C \quad (4.7) $$

to represent the climatological difference between the forced (subscript $F$) and control (subscript $C$) experiments. Then the response of the mean moisture convergence to the
forcing can be derived as follows:

\[
\delta MC = \left( \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \bar{u}_{k,F} \bar{q}_{k,F} \Delta p_{k,F} \right) - \left( \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \bar{u}_{k,C} \bar{\bar{q}}_{k,C} \Delta p_{k,C} \right)
\]

\[
\approx \left( \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \bar{u}_{k,C} \bar{\bar{q}}_{k,C} \Delta p_{k,F} \right) - \left( \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \delta \bar{u}_{k,C} \bar{\bar{q}}_{k,C} \Delta p_{k,C} \right)
\]

\[
= \delta TH + \delta DY.
\]

\(\delta TH\) represents the thermodynamic component, with circulation fixed at its climatological value in the CTRL experiment, thus involving only changes in specific humidity; and \(\delta DY\) represents the dynamic component, with specific humidity fixed at its climatological value in the CTRL experiment, thus involving only changes in circulation. In the derivation of Eq. (8), the approximation originates from ignoring 1) the quadratic term involving covariances of departures from climatological values (small compared to the other terms); and 2) surface pressure variations that cause differences between \(\Delta p_{k,F}\) and \(\Delta p_{k,C}\) (which has been shown in Seager and Henderson (2013) to introduce little additional error).

### 4.3 Precipitation response to future GHG forcing

#### 4.3.1 Projected 21st century monsoon rainfall change in CMIP5 coupled models

We start with examining the projected summer monsoon rainfall change for the 21st century under the rcp8.5 scenario in all the realizations of the 35 coupled CMIP5 models. Figure 4.1 shows the linear trend of area averaged land precipitation from 2006 to 2099 for India \((5^\circ N - 30^\circ N, 70^\circ E - 90^\circ E, \text{Fig. 4.1a})\) and eastern China \((20^\circ N - 40^\circ N, 105^\circ E - 125^\circ E, \text{Fig. 4.1b})\). The shaded bars show the 95% confidence intervals based on 2-sided Student’s \(t\) test applied onto detrended data, thus indicating the interannual range of the trend due
to natural variability. Generally speaking, the majority of the models predict a stronger monsoon for both India and eastern China, while the magnitude differs among individual models. A few models, however, show a weak signal or drying trend. The group of models with weak or opposing signs also differs for India and eastern China.

The multi-model mean (MMM), plotted at the bottom in black, gives a wetting trend for both regions. In calculating the MMM, we first computed each model’s ensemble average, and then averaged across the 35 models. This wetting trend is consistent with previous studies suggesting an intensified monsoon (Endo and Kitoh 2014; Li et al. 2015), with stronger wetting over India than eastern China. This may be related to the larger discrepancy among individual models over eastern China, with more models showing insignificant weak responses. The spatial pattern of the change displays an overall uniform wetting, as shown in Fig. 3.1d.
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Figure 4.1 indicates that while most models show an intensified Asian monsoon rainfall, there is substantial model spread. It is not clear whether the intensification as well as the spread is caused mainly by the fast or slow responses, or, caused more by thermodynamic versus dynamic mechanisms, which will be explored in the following sections.

4.3.2 Effects of \( \text{CO}_2 \) radiative forcing and SST warming in AMIP experiments

In order to decompose the different physical processes contributing to the overall wetting and model spread as shown in Fig. 4.1, we use the AMIP experiments to analyze the relative roles of the fast response due to direct radiative forcing and the slow response due to SST change. Since rcp8.5 is the “business as usual” scenario with the highest GHG emissions and gradually phasing out anthropogenic aerosol emissions, the total response is dominated by the \( \text{CO}_2 \) forcing. A detailed estimate of the radiative forcing can be found in Shindell et al. (2013).

Figures 4.2a and 4.2b show the precipitation response (\( \delta P \)) for \( 4 \times \text{CO}_2 \) and +4\( K \), respectively. Here we show the global pattern to better identify the large-scale changes. The direct radiative forcing and SST warming exert significantly different responses both over the land and ocean. Over the ocean, the fast response (Fig. 4.2a) displays an overall drying globally, while the slow response (Fig. 4.2b) shows particularly enhanced wetting over the intertropical convergence zone (ITCZ), the western Pacific warm pool region, and the higher latitudes extending towards the polar regions. The land rainfall response displays significant regional variations in both cases. For regions such as the Mediterranean and western United States, both the fast and slow components contribute towards a drying trend. For the Sahel, on the other hand, the two oppose each other, with the direct radiative effect wetting the Sahel and the SST effect drying the region. For the Asian monsoon, the fast response enhances monsoon rainfall over most regions in China, northern and central India, but reduces rainfall in southern India and Indochina. On the other hand, the slow response dries eastern China and northern India, but has a strong wetting over Indochina, central and southern India, possibly due to oceanic influence extending from the western Pacific warm pool.
Figure 4.2: Precipitation response $\delta P$ for (a) $4 \times CO_2$, (b) $+4K$, (c) $4 \times CO_2$ plus $+4K$, and (d) rcp8.5. Stippling denotes 7 out of 10 models agree on the sign of change. Units are $mm\ day^{-1}$.

The combination of the two responses (adding Figs. 4.2a and 4.2b) is shown in Fig. 4.2c. As a comparison, Fig. 4.2d shows the rcp8.5 response using the coupled version for the same 10 models. The rcp8.5 response is calculated using the climatological difference between the 2075 to 2099 period and 2006 to 2030 period. Despite the highly idealized settings and differences in forcings, Fig. 4.2c is able to capture the general large-scale pattern in the AGCMs (Fig. 4.2d) with a weighted spatial pattern correlation of 0.55. There are regional differences including an overly wetting in the western tropical Pacific and dry bias in the Indian Ocean in the AGCMs, possibly related to the effects of the SST warming pattern (Xie et al. 2010; He et al. 2014)

Figures 4.2c and 4.2d show that the future total rainfall response to CO$_2$ forcing is a combination of direct radiative effect and SST warming. The SST effect (Fig. 4.2b) dominates the total response over the ocean. Over land, for the Mediterranean and western
United States where the two effects have the same sign, there is a strong drying trend in the combined as well as the coupled response, contributing to a more robust future precipitation projection in these regions (Seager et al. 2007, 2014; Kelley et al. 2012; Christensen et al. 2013). For most part of the Asian monsoon region, the direct radiative effect (Fig. 4.2a) dominates, resulting in enhanced monsoon rainfall as climate warms, consistent with Bony et al. (2013). However note that the southern part of India as well as Indochina display opposite responses compared to northern India and eastern China, possible influenced by the surrounding oceans.

4.4 Thermodynamic and dynamic mechanisms of CO₂-induced rainfall changes: direct radiative forcing versus SST warming

4.4.1 Moisture and circulation

Why does monsoon precipitation respond differently to direct radiative forcing and SST warming? In this section, we analyze the thermodynamic and dynamic mechanisms contributing to the difference, and compare with the results of Shaw and Voigt (2015). Figures 4.3a-d show the change of 850hPa specific humidity and winds for $4 \times CO_2$ and +4K. The moisture response is dominated by the strong increase due to SST warming (Fig. 4.3b), with little change related to direct radiative forcing (Fig. 4.3a). Atmospheric circulation, on the other hand, shows distinct compensating effects over most regions in the northern hemisphere. In particular, direct radiative forcing (Fig. 4.3c) shifts the Asian monsoon circulation northward and shows a generally intensified monsoon circulation, which is similar to the response in the coupled models, both rcp8.5 and GHG-only single-forcing simulations (Fig. 3.4d, Fig. 3.8d); SST warming, on the other hand, weakens the monsoon circulation over Asia (Fig. 4.3d). Consistent with the 925hPa stationary eddy streamfunction and 500hPa vertical motion responses as shown in Shaw and Voigt (2015), direct radiative forcing enhances the Pacific subtropical anticyclone and SST warming weakens it. In the southern hemisphere, the fast and slow components show similar circulation responses, with
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Figure 4.3: Response of 850hPa (a, b) specific humidity $\delta q$ (units $g \text{kg}^{-1}$) and (c, d) winds $\delta u$ (units $m \text{s}^{-1}$) for (a, c) $4 \times CO_2$ and (b, d) +4K. In (c, d), arrows are vectors of the response and shading shows the change in wind speed. Stippling as in Fig. 4.2.

a poleward shifted jet stream, consistent with Grise and Polvani (2014).

One possible explanation for the opposite changes in circulation for $4 \times CO_2$ and +4K is the difference in land-sea thermal contrast changes (Kamae et al. 2014a; Shaw and Voigt 2015). Figures 4.4a and 4.4b show the air temperature at 925hPa (shown in thermodynamic energy unit, $c_p T$, where $c_p$ is the specific heat of air at constant pressure, $T$ is air temperature) for the $4 \times CO_2$ and +4K experiments. In the $4 \times CO_2$ experiment, land warms more than ocean as the heat capacity of the land is much lower than water. The exception is a cooling trend over southern India in $4 \times CO_2$, possibly induced by increased evaporation (Fig. 4.5a), although the exact reason remains unclear. In the +4K case, there is strong ocean warming, which leads to a greatly moistened lower troposphere (Fig. 4.3b). Note, however, that land still warms slightly more than ocean in +4K (Fig. 4.4b), consistent with Joshi et al. (2008) who propose local feedbacks and the hydrological cycle over land as the mechanism for enhanced land-warming (see also Byrne and O’Gorman (2013)). Thus one cannot explain the opposing monsoon circulation response purely based on changes in temperature.
Figure 4.4: Response of 925 hPa (a, b) air temperature in energy unit $\delta c_p T$, (c, d) specific humidity in energy unit $\delta L_v q$, and (e, f) moist static energy $\delta MSE$ for (a, c, e) $4 \times CO_2$ and (b, d, f) $+4K$. Stippling as in Fig. 4.2. Units are $kJ \ kg^{-1}$. 
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Figures 4.4c and 4.4d show the change in atmospheric moisture (shown in energy unit, $L_v q$, where $L_v$ is the latent heat of vaporization, $q$ is the specific humidity) at 925hPa for $4 \times CO_2$ and $+4K$. It is clear that the $+4K$ experiment is associated with a much stronger moistening over ocean than over land (Fig. 4.4d). There are also significant changes in the moisture component over land for the $4 \times CO_2$ case (Fig. 4.4c), particularly over the monsoon region as a response to enhanced moisture transport due to stronger monsoon circulation (Fig. 4.3c). The total moist static energy ($MSE$) changes are shown in Figs. 4.4e and 4.4f, which combines the effect of temperature and moisture changes (defined as $MSE = c_p T + L_v q + gZ$, $g$ is gravity, and $Z$ is the geopotential height; the potential energy change associated with $gZ$ is negligible in this case). Figure 4.4 suggests that the land-sea moisture contrast is the dominant contributor to the total $MSE$ contrast for the $+4K$ experiment, and to a large extent, to the $4 \times CO_2$ experiment as well. Our results here indicate that $MSE$ increases more (less) over land than over ocean in $4 \times CO_2$ ($+4K$) experiment, thus enhancing (weakening) the monsoon circulation (Fig. 4.3c, d). This conclusion is consistent with Shaw and Voigt (2015) using the sub-cloud equivalent potential temperature. The dominance of land-sea moisture contrasts in the $MSE$ contrasts in $+4K$ is insensitive to whether SST warming is uniform or not, as it is confirmed using “amipFuture” experiments with patterned SST warming (not shown), suggesting that gradients in SST is not essential in driving the land-sea thermodynamic contrasts and monsoon circulation change.

4.4.2 The moisture budget

In this subsection, we further examine the thermodynamic and dynamic mechanisms causing the rainfall changes using moisture budget analysis. The pattern of $P - E$ change, the net change in surface water balance, largely follows the rainfall changes (Figs. 4.2a, b) for both the direct radiative forcing (Fig. 4.5c) and SST warming (Fig. 4.5d), particularly over land. For oceanic regions, evaporation increases greatly due to SST warming (Fig. 4.5b) but decreases in the $4 \times CO_2$ response (Fig. 4.5a). The reduction in evaporation in the $4 \times CO_2$ case can be understood through the atmospheric heat budget due to quadrupling CO$_2$ while keeping the SST fixed. On average, the atmosphere cools through longwave radiation, which is largely balanced by latent heat flux into the atmosphere at the surface.
and thus precipitation (sensible heat flux also contributes, but much less than latent heat flux). The quadrupling of CO₂ leads to a decrease in atmospheric radiative cooling due to the anomalous absorption of longwave radiation by the atmosphere, which has to be balanced by a reduction in surface latent heat flux into the atmosphere through reduced surface evaporation (Yang et al. 2003; Bala et al. 2008; Andrews et al. 2009; Bala et al. 2010). Over the Asian monsoon region, there are small increases in evaporation over land due to increased precipitation. We will explore further the causes for the change in \( P - E \) through the atmospheric moisture budget.

Changes in the mean moisture convergence (\( \overline{\delta MC} \)) and transient eddies (\( \overline{\delta TE} \)) are shown in Fig. 4.6. In the tropics, \( \overline{\delta MC} \) (Figs. 4.6a, b) dominates over \( \overline{\delta TE} \) (Figs. 4.6c, d), and agrees well with the \( \delta(P - E) \) pattern. For the higher latitudes, \( \overline{\delta TE} \) plays an important role in the +4K case. In particular, the high latitude wetting trend, suggested by Zhang et al. (2013) to be primarily driven by the increasing poleward atmospheric moisture transport,
Figure 4.6: As in Fig. 4.5, but for (a, b) the mean moisture convergence $\delta MC$ and (c, d) transient eddies $\delta TE$.

is dominated by the transient contributions.

To illustrate the change in the surface hydrological cycle and the model spread the Asian monsoon region, Fig. 4.7 shows the area averaged changes in precipitation, evaporation, $P - E$, and the mean and transient moisture convergence terms over Indian and eastern China for 4 $\times$ $CO_2$ (red), +4K (blue) and rcp8.5 (green), with black dots denoting the MMMs. For both India (Fig. 4.7a) and eastern China (Fig. 4.7b), $\delta MC$ dominates $\delta (P - E)$ and $\delta P$, with $\delta TE$ and $\delta E$ terms being relatively small across all three types of experiments. In both regions, direct radiative forcing (red) enhances the mean moisture convergence, while SST warming (blue) shows a large spread and a slightly increased moisture convergence in MMM for India and a slightly decreased moisture convergence for eastern China. The rcp8.5 response (green), although not directly comparable in magnitude, largely follows the sum of the two effects (not shown). Figure 4.7 also indicates that the enhanced Asian monsoon rainfall in the rcp8.5 scenario is dominated by direct radiative forcing (fast response) rather
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Figure 4.7: Area averaged changes of moisture budget terms ($\delta P$, $\delta E$, $\delta (P - E)$, $\delta MC$, $\delta TE$) over (a) India and (b) East China for (red) $4 \times CO_2$, (blue) $+4K$, and (green) rcp8.5. The black dots show the 10-model MMMs. Units are mm day$^{-1}$. 

The black dots show the 10-model MMMs. Units are mm day$^{-1}$. 

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than the equilibrium response due to SST warming. Also note that the model spread in the +4K experiment is larger than that in $4 \times CO_2$, suggesting higher uncertainty in the slow response related to SST change. One possible factor contributing to the larger spread is the dipole structure with a strong wetting over central and southern India and drying to the northeast (Figs. 4.2b, 4.5d, 4.6b).

Figure 4.8 shows the thermodynamic ($\delta TH$) and dynamic ($\delta DY$) components of the mean moisture convergence change, as well as the quadratic term ($\delta quad$) which is small compared to $\delta TH$ and $\delta DY$ in all cases. Similar to the specific humidity response (Figs. 4.3a, b), SST warming (Fig. 4.8b) dominates over the direct radiative forcing (Fig. 4.8a) in terms of thermodynamical changes. Over the monsoon regions, this leads to strongly enhanced wetting for both India (Fig. 4.9a) and eastern China (Fig. 4.9b). Dynamical changes, on the other hand, exhibit competing effects in response to direct radiative forcing (Fig. 4.8c) and SST warming (Fig. 4.8d) in the northern hemisphere including the North Pacific, North Atlantic, the Asian and African monsoons, and North America. Both the fast and slow responses are robust across the models, as indicated by the stippling (70% model agreement). The area averaged $\delta DY$ for India (Fig. 4.9a) and eastern China (Fig. 4.9b) show distinct opposing responses: in all the models, direct radiative effect (red) enhances moisture convergence while SST warming (blue) weakens it. Due to this cancellation, the rcp8.5 response (green) is very weak for the MMM, with a large model spread.

Figures 4.7 and 4.9 indicate that the different responses to direct radiative forcing and SST warming of the total mean moisture convergence, and hence precipitation, is mainly caused by the opposing effects of the dynamic component, related to changes in atmospheric circulation. These results are consistent with Shaw and Voigt (2015), but focusing specifically on the South and East Asian monsoons. Furthermore, the thermodynamic and dynamic effects of SST warming, which evolve on the longer timescale, largely cancel out (also see Shaw and Voigt (2015), Fig. S1g, h). Thus the total mean moisture convergence is dominated by the fast response that occurs shortly after the CO$_2$ increase due to dynamical changes, in agreement with Bony et al. (2013). Note that similar to the precipitation responses, southern India shows different changes in the moisture budget terms as compared to that over northern India and eastern China.
Figure 4.8: As in Fig. 4.5, but for (a, b) the thermodynamic component $\delta TH$, (c, d) the dynamic component $\delta DY$, and (e, f) the quadratic term $\delta quad$. 
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Figure 4.9: As in Fig. 4.7, but for $\delta MC$, $\delta TH$, $\delta DY$, and $\delta quad$. 
4.5 Understanding uncertainties in future monsoon projections

In section 4, we have shown that the dynamical responses to direct radiative forcing of rising CO$_2$ and subsequent SST warming largely oppose each other over the Asian monsoon region. What does this imply for future monsoon projections? As discussed in Shaw and Voigt (2015), this competing effect may lead to a weak circulation response and lack of consensus among the models. This is confirmed in Fig. 4.9, in which the 10-model averaged rcp8.5 response of $\delta D$ is almost zero for both regions, with almost equal number of models showing positive and negative sign changes. On the other hand, the thermodynamical response, although different in magnitude, is largely consistent in sign across the models. Here we further examine the model discrepancy using the larger ensemble of the 35 CMIP5 coupled models.

Figures 4.10 and 4.11 show the linear trend of the thermodynamic and dynamic components of the mean moisture convergence [see Chap. 3 for detailed definition of the terms] over 2006-2099 under the rcp8.5 scenario for India and eastern China, respectively. The shaded bars are defined as in Fig. 4.1. While almost all the models display significant wetting trends due to the thermodynamic component (Figs. 4.10a, 4.11a), the dynamic component (Figs. 4.10b, 4.11b) shows a clear model spread for both regions. The projected rainfall change (Fig. 4.1) is determined by the relative sign and magnitude of the two components. For example, the group of models with drying trends over India (CMCC-CESM, CMCC-CMS, FIO-ESM, MPI-ESM-LR, MPI-ESM-MR) all show strong dynamical weakening. The MMM trend (black) of the dynamic component is very weak for both regions, consistent with Chapter 3 (Li et al. 2015) using the larger monsoon region. There are two reasons for this weak dynamic response: first, the cancellation effect due to inter-model disagreement, as there are models showing significantly increased or decreased moisture convergence; second, the weak responses in some of the models, possibly related to the cancellation effect between direct radiative forcing and SST warming. Thus the uncertainty in future circulation change is related to both model discrepancy and the multiple physical processes involved. While previous studies have suggested that atmospheric circulation dominates
Figure 4.10: Linear trend of area averaged (a) thermodynamic and (b) dynamic components of the mean moisture convergence over land in 35 CMIP5 models under the rcp8.5 scenario from 2006 to 2099 for India. The black and shaded bars are defined as in Fig. 4.1. Units are $\text{mm day}^{-1}\text{10yr}^{-1}$.

The inter-model variations in the tropics focusing on the standard deviation (Kent et al. 2015; Xie et al. 2015), we emphasize here that over the monsoon regions there is substantial uncertainty in both the magnitude and the sign of the dynamical changes.

Our results indicate that the Asian monsoon response to uniform SST warming exhibits a larger model spread as compared to direct radiative forcing in terms of the mean moisture convergence change (Fig. 4.7), which may come from the cancellation between the thermodynamical and dynamical processes (Fig. 4.9). Furthermore, SST spatial pattern may not be the dominant cause of the model spread, although it certainly could make a contribution. The large uncertainty in the dynamical contribution to mean moisture convergence as shown in Figs. 4.10b and 4.11b is largely due to the cancellation of the circulation response to direct radiative forcing and SST warming. The degree of each model’s relative importance of the two responses may have led to the large spread in the dynamical component. The weak MMM dynamical response for the rcp8.5 scenario is largely caused by this large
model spread rather than a uniformly weak circulation response in individual models. Thus a better projection of the future monsoon rainfall changes depends strongly on reducing the uncertainty in monsoon circulation responses to greenhouse warming.

Previous studies have proposed other possible mechanisms contributing to the uncertainty in regional hydroclimate changes. For example, Bony et al. (2013) suggest the fact that climate models show a large range of climate sensitivity (Andrews et al. 2012) may alter the importance of the thermodynamical changes relative to the dynamical changes and hence the precipitation response. On the contrary, Kent et al. (2015) find that the intermodel uncertainty of tropical rainfall and circulation is not strongly influenced by global mean temperature changes. We have tested the relationship between the thermodynamic and dynamic components and the equilibrium climate sensitivity (ECS) for India and eastern China in the CMIP5 archive (not shown). Results show that while the thermodynamical enhancement of a model is positively correlated with its ECS, the dynamical change only shows a very weak negative correlation with ECS. Other contributing factors to the model uncertainty include the ability of global climate models to simulate the present-day regional climate including monsoon climatology (see Chapter 1), natural variability such as

![Figure 4.11: As in Fig. 4.10, but for East China.](image)
the teleconnection between Asian monsoon and the El Niño-Southern Oscillation (Li and Ting 2015), as well as the possible influence of model resolution. Using a GCM downscaled to a ∼ 35 km high-resolution over the South Asian monsoon region, Krishnan et al. (2015) show that the model predicts a persistent drying in the 21st century, opposing the global model response robust across CMIP5 models. In addition, aerosol forcing represents a major uncertainty for climate prediction (Turner and Annamalai 2012), with opposing effect to GHG forcing on monsoon rainfall (Li et al. 2015). Further work is needed to advance the understanding and reduce the uncertainty in future monsoon projections.

4.6 Conclusions

In this chapter, we have examined the regional hydroclimate response to rising GHGs using coupled and idealized atmosphere-only models. The future total rainfall response to increasing CO$_2$ is a combination of the fast component due to direct radiative forcing and the slow component due to SST changes. While it is expected that the fast (slow) component would weaken (enhance) the global hydrological cycle due to associated changes in atmospheric radiative cooling, the Asian monsoon rainfall shows the opposite tendency. The CO$_2$ direct radiative effect leads to a much intensified monsoon rainfall, and SST warming either weakens (eastern China) or slightly intensifies (India) monsoon rainfall. Furthermore, the intensified monsoon rainfall associated with direct radiative forcing is mainly due to circulation changes, rather than through thermodynamic changes in the moisture budget sense. The thermodynamic mechanisms, on the other hand, do suggest a much intensified mean moisture convergence due to SST warming, but the enhancement is largely offset by the opposite change in circulation that weakens the monsoon. Overall for most part of the Asian monsoon region (southern India shows different characteristics), the monsoon rainfall changes are dominated by the direct CO$_2$ radiative effect through enhanced monsoon circulation.

The relative effects of direct radiative forcing and SST warming on monsoon circulation has important implications for the uncertainty in future projections. While the thermodynamical response is robust across the models and well understood, there are substantial
uncertainties in both the magnitude and the sign of the dynamical changes. For the Asian monsoon, the model spread for the uniform +4K SST warming case is larger than the corresponding one for 4 × CO₂ radiative forcing. The main cause of the model spread in the uniform +4K SST warming case may come from the cancellation between the dynamical and thermodynamical processes. While the MMM rcp8.5 scenario indicates a weak contribution due to dynamical mechanisms of future monsoon changes (Li et al. 2015), the results here indicate that this is largely caused by the large model spread rather than a uniformly weak circulation response in individual models. The lack of consensus among the models and weak MMM responses for the circulation changes in CMIP5 models may be related to the multiple physical processes evolving on different time scales. Understanding the physical mechanisms underlying the circulation changes and the model spread due to uniform SST warming is essential towards constraining the uncertainties in regional climate projection.
Chapter 5

Fast adjustments of the Asian summer monsoon to anthropogenic aerosols


5.1 Introduction

Over the past few years, “Asian smog” has become one of the most pressing environmental threats around the world, featuring unprecedentedly high pollution levels and severe impacts onto the world’s most densely populated region. According to a study conducted by the Global Burden of Disease from Major Air Pollution Sources (GBD MAPS) Working Group, ambient fine particular matter ($PM_{2.5}$) is a major contributor to mortality and disease burden in China, with an estimated contribution of 916,000 deaths in 2013, among which 40% is caused by coal combustion (GBD MAPS Working Group 2016). Apart from the direct health impacts, air pollution has been shown to affect meteorological conditions, as well as long-term climate variations (Gong et al. 2007; Boucher et al. 2013). Aerosols can affect global and regional climate by altering the radiation budget and interacting with clouds through microphysical processes, thereby causing subsequent changes in surface tem-
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Temperature as well as the hydrological cycle (Allen and Ingram 2002; Ming et al. 2010; Levy et al. 2013; Li et al. 2016).

The most populated regions of the Asian continent are characterized by a distinct monsoonal climate, spanning from the Indian subcontinent to the extratropics of eastern Asia. The summer monsoon brings over 80% of the annual rainfall in these regions, essential for local economy, agriculture, ecosystems and human health (Piao et al. 2010; Hong and Kim 2011). In addition, as a major part of the summertime overturning circulation in the northern hemisphere tropics, the Asian monsoon has profound remote influences on the global-scale climate (Rodwell and Hoskins 1996; Lin and Wu 2012). Previous studies have shown that the Asian summer monsoon has weakened during the 20th century, with anthropogenic aerosols being a likely cause (Ramanathan et al. 2005; Bollasina et al. 2011; Ganguly et al. 2012a; Li et al. 2015). However, the exact trends and reasons remain unclear with large uncertainties in climate model simulations (Turner and Annamalai 2012).

Despite prominent influence of natural variability (Kumar et al. 2006; Li and Ting 2015), the Asian summer monsoon is predominantly affected by greenhouse gases (GHGs) and aerosols in the 20th century (Turner and Annamalai 2012; Singh 2016). It has been shown that GHGs and aerosols have significant competing effects on monsoon rainfall change, with aerosols dominating the total drying trend during the 20th century (Li et al. 2015). A number of studies have addressed the possible physical mechanisms of aerosol-induced tropical rainfall and monsoon changes (Lau et al. 2006; Ganguly et al. 2012a; Guo et al. 2013, 2015; Hwang et al. 2013). Unlike GHGs that induce a strong atmospheric moistening through an increase in sea surface temperature (SST), aerosols affect monsoon rainfall largely through changes in atmospheric circulation (Li et al. 2015). Increased aerosols in the atmosphere could reduce the surface solar radiation (“dimming” effect) which reduces the local SST gradient in the Indian Ocean (Ramanathan et al. 2005; Chung and Ramanathan 2006) and introduces a hemispheric energy imbalance caused by the spatial inhomogeneity of aerosol distributions (Bollasina et al. 2011), as well as increase atmospheric stability through direct and indirect effects (Lau and Kim 2017), contributing to weakened monsoon circulation. Some other studies find that aerosols may enhance monsoon rainfall and circulation over the South China Sea and western Pacific (Jiang et al. 2013) or cause an earlier onset and
enhanced June rainfall over India (Bollasina et al. 2013) due to absorbing aerosols such as black carbon (Lau et al. 2006), indicating high complexity and uncertainty associated with aerosol-monsoon interactions.

From an energetics perspective, the response of the climate system to an external forcing involves two components on different time scales: the fast response without the mediation of SST and the slow response due to SST feedbacks (Allen and Ingram 2002; Andrews et al. 2009). The fast and slow components may lead to differing responses in the hydrological cycle, often studied using idealized atmospheric general circulation model (AGCM) experiments with prescribed SSTs (Hsieh et al. 2013; Shaw and Voigt 2015; Richardson et al. 2016). A number of studies have shown that GHGs may cause compensating effects due to direct radiative forcing (fast response) and SST change (slow response) in atmospheric circulation, particularly over the Asian monsoon region, leading to large uncertainties and model spreads in coupled model simulations (Shaw and Voigt 2015; Li and Ting 2017). Also for GHGs, some studies have emphasized the importance of the fast adjustments in explaining the total rainfall responses (Bony et al. 2013; Li and Ting 2017). However for anthropogenic aerosols, it has been shown that the slow response due to SST change may dominate the total monsoon rainfall and circulation changes over India (Ganguly et al. 2012b) and East Asia (Kim et al. 2016) while the fast adjustments due to sulfate aerosols contribute to a slightly intensified East Asian monsoon (Kim et al. 2016), both studies using a single AGCM. How much of the total monsoon response can be explained by the aerosol fast response without the mediation of SSTs? What is the role of SST feedbacks? These questions have not been fully explored for aerosol forcing.

Here we examine the Asian summer (June-August seasonal mean, JJA) monsoon response to anthropogenic aerosols on different time scales, with a focus on the fast adjustments independent of SST changes. We analyze the physical mechanisms of the total and fast responses, and discuss the possible role of SST feedbacks. We address the question by using a suite of coupled general circulation models (CGCMs) in the Coupled Model Intercomparison Project - Phase 5 (CMIP5), and AGCMs with prescribed aerosol concentration and SSTs. Model simulations and analysis methodology are provided in section 2. We present the aerosol-induced total and fast monsoon rainfall responses in section 3, and
analyze the thermodynamic and dynamic mechanisms in section 4. The main conclusions are summarized in section 5.

5.2 Methods

5.2.1 Model simulations

To examine the effect of aerosol forcing, we used multiple model simulations including both CGCMs and AGCMs. For CGCMs, we used output from CMIP5 models (Taylor et al. 2012) under the historical aerosol-only scenario, with monthly data from 1861 to 2005. The total response is quantified using the climatological difference between 1981-2005 and 1861-1885. To examine the fast response, we used idealized AGCM simulations with prescribed SST and sea ice concentration that are part of the CMIP5 archive. The control simulation (called “sstclim” in the CMIP5 archive) uses fixed climatological SST and sea ice concentration from the pre-industrial control simulation, and pre-industrial anthropogenic aerosols. The forced experiment (“sstclimAerosol” in the CMIP5 archive) uses the same SST and sea ice as the control simulation, but with year 2000 aerosols from the CMIP5 historical simulations. The 30-year climatological difference between the forced and control simulations quantifies the fast response, thus independent of SST changes.

The CGCM and AGCM monthly outputs are available for 13 and 11 models (multiple realizations available for some of the CGCMs) in the CMIP5 archive, respectively, however with only 5 models in common (Table 5.1). The results using all available models and the subset using the 5 common models are largely robust (Fig. 5.2). Here we show the results using the 5 common models using only the first realization for consistency purposes. All model outputs were interpolated to a $2^\circ \times 2^\circ$ spatial resolution.

We further performed idealized AGCM experiments using the Community Earth System Model (CESM) version 1.2.0 with F-compset to examine the slow response due to aerosol-induced SST changes. F-compset consists of interactive atmosphere model (CAM5) and land surface model (Community Land Model 4.0, CLM4) with prescribed SSTs and sea ice concentration. A detailed description of the model can be found in Neale et al. (2012). We use $1.9^\circ$ latitude $\times$ $2.5^\circ$ longitude horizontal resolution, 26 vertical levels, and the
### Table 5.1: List of CMIP5 CGCMs, AGCMs, and available realizations used in Chapter 5.

Asterisks denote the common 5 models used in the study (using the first realization). For aerosols, “F” means “fully-interactive”, “S” means “semi-interactive”, and “N” indicates “non-interactive” (See Table 9.1 in Flato et al. (2013)).

<table>
<thead>
<tr>
<th>Model</th>
<th>CGCM</th>
<th>AGCM</th>
<th>Aerosol</th>
</tr>
</thead>
<tbody>
<tr>
<td>bcc-csm1-1</td>
<td>-</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>CanESM2 *</td>
<td>5</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
<td>CCSM4 (p10)</td>
<td>3</td>
<td>-</td>
<td>S</td>
</tr>
<tr>
<td>CESM1-CAM5</td>
<td>3</td>
<td>-</td>
<td>F</td>
</tr>
<tr>
<td>CSIRO-Mk3-6-0 *</td>
<td>5</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
<td>FGOALS-g2</td>
<td>1</td>
<td>-</td>
<td>S</td>
</tr>
<tr>
<td>FGOALS-s2</td>
<td>-</td>
<td>1</td>
<td>S</td>
</tr>
<tr>
<td>GFDL-CM3 *</td>
<td>3</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
<td>GFDL-ESM2M</td>
<td>1</td>
<td>-</td>
<td>S</td>
</tr>
<tr>
<td>GISS-E2-H (p107)</td>
<td>5</td>
<td>-</td>
<td>F</td>
</tr>
<tr>
<td>GISS-E2-H (p301)</td>
<td>5</td>
<td>-</td>
<td>F</td>
</tr>
<tr>
<td>GISS-E2-R (p107)</td>
<td>5</td>
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<td>HadGEM2-A</td>
<td>-</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
<td>IPSL-CM5A-LR *</td>
<td>1</td>
<td>1</td>
<td>S</td>
</tr>
<tr>
<td>MIROC5</td>
<td>-</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
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<td>-</td>
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<td>N</td>
</tr>
<tr>
<td>MRI-CGCM3</td>
<td>-</td>
<td>1</td>
<td>F</td>
</tr>
<tr>
<td>NorESM1-M *</td>
<td>1</td>
<td>1</td>
<td>F</td>
</tr>
</tbody>
</table>
cam4 physics package with prescribed gases (except for water vapor) and bulk aerosols. The control simulation (CTRL) is run with year 1850 aerosol concentration and 1951-2000 climatological SST and sea ice concentration from the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset (Rayner et al. 2003). The forced experiment (AEROSST) uses the same aerosol and sea ice concentration as the control experiment, but adding an SST anomaly derived from the CMIP5 historical aerosol-forcing only simulations to the observed climatological SST. The SST anomaly is the climatological difference between 1981-2005 and 1861-1885 using the multi-model mean (MMM) of the 5 common CMIP5 models listed in Table 5.1. All other gaseous species are fixed at preindustrial levels. Both experiments are run for 60 years (after an initial 1-year spin-up), the climatological difference between AEROSST and CTRL quantifies the slow response.

5.2.2 Moisture budget analysis

We analyze the atmospheric moisture budget to quantify the changes in the hydrological cycle. A detailed derivation and discussion of the possible errors can be found in Chapter 4 (Li and Ting 2017) and Seager and Henderson (2013). In steady state, the following balance can be expressed in pressure coordinates as:

\[
P - E = -\frac{1}{g \rho_w} \nabla \cdot \int_0^{p_s} u q \, dp \approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_k q_k \Delta p_k,
\]

where \( P \) is precipitation, \( E \) is evaporation, \( g \) is gravitational acceleration, \( \rho_w \) is the density of water, \( p \) is pressure and \( p_s \) surface pressure, \( u \) is the horizontal wind vector, \( q \) is specific humidity, \( k \) is the vertical level with a total of \( K \) (here \( K = 10 \), from 1000hPa to 200hPa), \( \Delta p \) is the pressure thickness. Overbars represent monthly mean values.

We then separate the moisture flux convergence term into the mean moisture convergence (MC) and the sub-monthly transient eddies (TE):

\[
P - E \approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_k q_k \Delta p_k - \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_k' q_k' \Delta p_k,
\]

where primes denote departures from monthly means. Here we approximate the transient eddy component using the difference between \( P - E \) and the mean moisture convergence due to limited availability of daily data.
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To quantify the forced response, we define

\[ \delta(\cdot) \equiv (\cdot)_F - (\cdot)_C \]  

(5.3)

to represent the difference between the forced (subscript \( F \)) and control (subscript \( C \)) experiments, where the second overbar denotes the 30-year climatological mean. Then the change in the mean moisture convergence can be derived as follows:

\[
\delta MC = \left( -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,F} \bar{q}_{k,F} \Delta \bar{p}_{k,F} \right) - \left( -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,C} \bar{q}_{k,C} \Delta \bar{p}_{k,C} \right) \\
\approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \mathbf{u}_{k,C} \delta \bar{q}_k \Delta \bar{p}_k - \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \delta \mathbf{u}_k \bar{q}_{k,C} \Delta \bar{p}_k \\
= \delta TH + \delta DY.
\]  

(5.4)

Here for the thermodynamic component (\( \delta TH \)), circulation (\( \mathbf{u} \)) is fixed at the CTRL climatological value, thus representing changes due to specific humidity (\( q \)); and for the dynamic component (\( \delta DY \)), specific humidity (\( q \)) is fixed at the CTRL climatological value, thus representing changes due to circulation (\( \mathbf{u} \)). The quadratic term involving covariances of departures from climatological values of \( \mathbf{u} \) and \( q \) is small compared to \( \delta TH \) and \( \delta DY \) (not shown).

5.3 Total versus fast monsoon precipitation response to aerosol forcing

What is the monsoon rainfall response to anthropogenic aerosols and how much can it be explained by the fast adjustment? Figure 5.1a, b shows the summertime precipitation change to aerosol forcing for (a) total and (b) fast responses, using MMMs of CGCMs and AGCMs, respectively. The total response (Fig. 5.1a) displays a strong drying pattern, consistent with Chapter 3 (Li et al. 2015) using a larger ensemble set of 35 models. This drying pattern in the coupled models is largely reproduced by the AGCMs without the SST feedbacks (Fig. 5.1b), particularly over eastern China and northern India. On the other hand, between \( 0^\circ - 20^\circ N \) over the oceanic regions as well as southern India, the fast response (Fig. 5.1b) shows a wetting trend that opposes the total response (Fig. 5.1a).
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Figure 5.1: (a, c) Total response in CGCMs and (b,d) fast responses in AGCMs of (a, b) precipitation and (c, d) 550nm aerosol optical depth (AOD) to historical aerosol forcing using the 5 common models. In (a, b), stippling denotes 4 out of 5 models agree on the sign of change, units are $\text{mm day}^{-1}$.

The similarity between Figs. 5.1a and 5.1b to the north of $20^\circ$N indicates that the fast adjustments dominate the monsoon rainfall response over the majority of the land regions, with an even stronger drying than the total response over eastern China.

The overall drying in the total response and the meridional land-ocean dipole structure in the fast response are largely robust using a larger set of available models (Fig. 5.2a, b). Separating the models with and without a fully-interactive aerosol scheme further shows that the aerosol indirect effects (aerosol-cloud interactions) dominate both the total and fast responses, particularly in the fast response which indicates almost no signal over land in models without fully-interactive aerosols (Fig. 5.2c-f). Among the five common models, the one model with semi-interactive aerosols (IPSL-CM5A-LR) shows much weaker responses compared to the other models with fully-interactive aerosols (not shown).

It should be noted that the total and fast responses are not precisely comparable due to the differences of both model setting and aerosol forcing. The fast response is quantified
Figure 5.2: (a, c, e) Total and (b, d, f) fast precipitation responses for (a, b) all available model realizations listed in Table S1, (c, d) models with “fully-interactive” aerosols, and (e, f) models with “semi-interactive” or “non-interactive” aerosols. Stippling indicates approximately 75% of the models agree on the sign of change. Units are mm day$^{-1}$. 
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using idealized AGCMs without the coupling to the ocean, which differs from the fully coupled model used in the total response. Furthermore, the aerosol forcing corresponding to these two responses are slightly different since the CGCMs are transient runs in which the aerosol forcing changes with time, whereas the AGCMs have fixed aerosol forcing. As shown in Fig. 5.1c, d using 3 models with available data, the response in aerosol optical depth (AOD) at 550nm corresponding to the total and fast rainfall responses are almost identical in both magnitude and spatial distribution over Asia, lending support to the comparison between the aerosol-induced rainfall changes. However, it should be noted that the differences between the AOD patterns, particularly over Europe, may cause remote influences on the Asian monsoon (Bollasina et al. 2014).

5.4 Identifying mechanisms of aerosol-induced monsoon changes on different time scales

Monsoon rainfall is controlled by both moisture supply (thermodynamics) and atmospheric circulation (dynamics). What are their relative roles in shaping the responses in Fig. 5.1a, b? We use the atmospheric moisture budget analysis to quantify the thermodynamic and dynamic mechanisms contributing to the rainfall changes, as illustrated in Fig. 5.3. Changes in the column-integrated mean moisture convergence ($\delta\overline{MC}$, Fig. 5.3a, b) balances well with the net surface water budget, precipitation minus evaporation ($\delta(P-E)$, not shown), in both the CGCMs and AGCMs, confirming that the transient eddies play a minor role over the monsoon region. Consistent with Li et al. (2015) using a slightly larger ensemble set of 9 models, the total response of $\delta\overline{MC}$ (Fig. 5.3a) is a combination of thermodynamic ($\delta\overline{TH}$, Fig. 5.3c) and dynamic ($\delta\overline{DY}$, Fig. 5.3e) effects, resulting in a strong decrease of moisture convergence. However for the fast response, the thermodynamic component ($\delta\overline{TH}$, Fig. 5.3d) contributes very little due to the limited moisture supply from the fixed-SST setting; and the dynamic component ($\delta\overline{DY}$, Fig. 5.3f) dominates the strong reduction of the mean moisture convergence, particularly over land to the north of 20°N.

The dynamical changes for the total and fast responses are confirmed using the 500hPa vertical pressure velocity ($\omega$) in Fig. 5.4a, b. The total response (Fig. 5.4a) shows a moder-
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Figure 5.3: (a, c, e) Total response in CGCMs and (b, d, f) fast response in AGCMs of (a, b) the mean moisture convergence ($\delta MC$), (c, d) the thermodynamic component ($\delta TH$), and the dynamic component ($\delta DY$). Stippling denotes 4 out of 5 models agree on the sign of change. Units are mm day$^{-1}$.
Figure 5.4: (a, c) Total and (b, d) fast responses of (a, b) 500hPa vertical pressure velocity ($\omega$) and (c, d) surface air temperature. In (a, b), contours show climatological values of 1861-1885 for (a) and the control simulation climatology for (b). Solid (dashed) contours show positive (negative) values, indicating sinking (rising) motion. Contour intervals are 0.04 Pa s$^{-1}$. The thick solid contour denotes 0. Stippling denotes 4 out of 5 models agree on the sign of change. (e, f) Area average of (green bar) land-only and (blue bar) ocean-only surface air temperature response for global, northern hemisphere, and southern hemisphere; the overall northern and southern hemisphere averages are shown in orange and magenta, respectively. Units are 10$^{-3}$ Pa s$^{-1}$ for (a, b) and K for (c-f).
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ate anomalous sinking motion over most of India and southern China. Between $0^\circ - 20^\circ N$, the response is relatively weak and inhomogeneous with low model agreement, but there is strong anomalous rising motion over the Indian Ocean south of the equator. For the fast response (Fig. 5.4b), unlike the weak response in Ganguly et al. (2012b) and the slight enhancement in Kim et al. (2016), the vertical motion change shows a distinct meridional dipole structure with high model agreement: strong sinking anomalies over the land monsoon regions from northern India to eastern China, and strong rising anomalies to the south over the adjacent oceans as well as southern India. The difference with previous studies as well as within the model ensemble (not shown) suggests that the results may be model dependent. This meridional pattern is well reproduced in $\delta MY$ (Fig. 5.3f) and $\delta MC$ (Fig. 5.3b).

The difference between Fig. 5.4a and Fig. 5.4b, particularly the absence of the strong rising anomaly over the oceans at $0^\circ - 20^\circ N$ in the total response, suggests that the slow response due to SST change may play an important role in shaping the monsoon circulation response over the oceanic regions. Previous studies have addressed the roles of land sea thermal contrasts and SST gradients in determining the aerosol-induced circulation changes (Chung and Ramanathan 2006; Bollasina et al. 2011; Richardson et al. 2016). Here we show the surface air temperature changes in the total and fast responses in Figs. 5.4c and 5.4d, respectively, and area-averaged values in Figs. 5.4e and 5.4f. While the land (green bars in Fig. 5.4e, f) cools off more than the ocean (blue bars in Fig. 5.4e, f) in both cases, aerosols also induce a strong meridional temperature gradient with a much cooler northern hemisphere in the total response (Fig. 5.4c, e), consistent with the stronger aerosol masking effect on global warming shown in Lau and Kim (2017). In the fast response, on the other hand, the temperature change is confined to the continental regions with little change over ocean (Fig. 5.4d, f). The meridional temperature gradient in the total response and the land-ocean temperature contrasts in both the total and fast responses are much weaker in the models without fully-interactive aerosols (Fig. 5.5), explaining the weaker rainfall responses (Fig. 5.2). The moist static energy (MSE, defined as $MSE = c_p T + L_v q + g Z$, where $c_p$ is the specific heat of air at constant pressure, $T$ is air temperature, $L_v$ is the latent heat of vaporization, $q$ is specific humidity, $g$ is gravity, and $Z$ is the geopotential height) response at
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Figure 5.5: As in Fig. 5.2, but for surface air temperature. Units are $K$. 

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925hPa, which incorporates the effects of both temperature and moisture, shows the large-scale meridional gradient in the total response is largely dominated by the temperature contribution while moisture also contributes over the Indian Ocean (not shown).

The meridional temperature gradient associated with SST feedbacks, coupled with the stronger land cooling, induce significant changes in the local atmospheric overturning circulation. Figure 5.6 shows the 60°E to 140°E zonally averaged vertical motion change for the total (Fig. 5.6a) and fast (Fig. 5.6b) responses. The climatological rising motion (dashed contours) spans a wide latitude band from 10°S to 40°N, representing an expanded cross-equatorial overturning circulation during the summer monsoon months and the seasonal migration of the intertropical convergence zone (ITCZ) (Schneider et al. 2014). The climatological sinking branch (solid contours) is located to the south of 10°S. The AGCMs reproduce well the regional overturning circulation (contours in Fig. 5.6b). The aerosol-induced total response (shading in Fig. 5.6a) features an overall reduction of the climatological circulation with anomalous sinking motion over most of the climatological convection region and anomalous rising motion south of the equator, consistent with previous studies (Bollasina et al. 2011; Li et al. 2015). However, the anomalous sinking motion is confined to the north of 20°N in the fast response (shading in Fig. 5.6b), where the major land mass is located (illustrated by the number of land grid points in Fig. 5.6e). There is strong anomalous rising motion between 0°−20°N over the oceanic regions. The difference between Figs. 5.6a and b (Fig. 5.6c, assuming linearity, approximates the “slow response”) shows strong anomalous sinking motion over 0°−20°N and anomalous rising motion south of the equator. The slow response due to aerosol-induced SST anomaly in CAM5 (Fig. 5.6d) shows very similar patterns to that in Fig. 5.6c. The consistency between Figs. 5.6c and 5.6d confirms that the anomalous overturning circulation near the equator is predominantly caused by the slow response due to SST feedbacks.

How does anthropogenic aerosol forcing affect the Asian monsoon circulation? We summarize the different physical pathways in Fig. 5.7. On a shorter time scale without the mediation of SSTs, the addition of anthropogenic aerosols in the atmosphere causes a rapid cooling of the local land surface, which induces anomalous sinking motion over land and anomalous rising motion over the adjacent oceans (fast response). On a longer time scale,
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Figure 5.6: 60°E to 140°E zonally averaged vertical pressure velocity (ω) of (a) total response in the 5 CMIP5 CGCMs; (b) fast response in the 5 AGCMs; (c) the difference between (a) and (b), approximating the slow response; (d) slow response in CAM5. Contours show climatological values of 1861-1885 for (a) and the control simulation climatology for (b-d). Solid (dashed) contours denote positive (negative) values, indicating sinking (rising) motion. Contour intervals are 0.009 Pa s⁻¹. The thick solid contour denotes 0. (e) Total number of land grid points for each latitude (using 1° × 1° resolution) from 60°E to 140°E. Stippling indicates 4 out of 5 models agree on the sign of change for (a-c) and statistically significant at 5% significance level using 2-tailed z-test for (d). Units are 10⁻³ Pa s⁻¹ for (a-d).
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Figure 5.7: Total, fast and slow monsoon circulation responses to anthropogenic aerosols. On a shorter time scale, aerosols cause a rapid cooling of the local land surface, which induces anomalous sinking motion over land and anomalous rising motion over the adjacent oceans (fast response). On a longer time scale, aerosols cause a decrease in SSTs, as well as a meridional temperature gradient with a stronger cooling in the northern hemisphere, resulting in anomalous sinking (rising) motion north (south) of the equator (slow response). The combination of the fast adjustments and SST feedbacks leads to an overall reduction of the local overturning circulation (total response).
aerosol forcing further causes a decrease in SSTs, while also inducing a meridional temperature gradient with a stronger cooling in the northern hemisphere due to asymmetry in aerosol emissions between the two hemispheres. Based on the weak temperature gradient approximation, horizontal temperature gradients are small in the tropical atmosphere due to the weak Coriolis effect near the equator (Sobel et al. 2001). As a consequence, an anomalous cross-equatorial upper-tropospheric flow occurs to offset the hemispheric temperature and energy imbalance, resulting in anomalous sinking (rising) motion north (south) of the equator (slow response). The combination of the fast adjustments and SST feedbacks, involving competing effects at $0^\circ - 20^\circ N$, causes an overall reduction of the local overturning circulation (total response).

5.5 Summary

In this chapter, using multiple CGCMs and idealized AGCM experiments, we have identified the physical mechanisms of the total and fast monsoon responses to aerosol forcing. The single-forcing model simulations and the AGCM approach, although highly idealized, provide an effective way to separate the physical processes driving the aerosol-induced monsoon responses on different time scales, essential for furthering the mechanistic understanding and reducing the uncertainties related to aerosol forced changes. Our results show the fast adjustment that occurs before surface temperature adjusts to the forcing largely explains the total reduced rainfall over the major land regions, dominated by changes in atmospheric circulation. However, this rainfall suppression over land is largely absent in models where indirect aerosol interaction with clouds is not included. This is similar to the findings for the total aerosol-induced response in Guo et al. (2015) based on 24 CMIP5 CGCMs, associated with a large increase in the cloud droplet number in models with indirect aerosol effects, thus reducing the precipitation efficiency and increasing atmospheric stability in the lower troposphere (Allen and Sherwood 2010; Lau and Kim 2017).

We have proposed possible physical pathways by which aerosols impact the local overturning circulation on different time scales. Unlike GHGs with competing effects in the fast and slow responses which lead to an overall weak circulation response (Shaw and Voigt
2015; Li and Ting 2017), both the fast and slow responses to aerosol forcing cause an anomalous overturning circulation, however centered around different latitudes. Both the land-ocean asymmetry and meridional temperature gradient are key factors in determining the aerosol-induced circulation responses. On the other hand, the possible competing effects of the fast and slow responses at the land-ocean boundary where multiple physical processes are entangled may contribute to a higher level of uncertainty. It is also unclear whether the SST feedbacks are dominated by overall aerosol-induced cooling, the global meridional temperature gradient, or the regional SST pattern, which will be addressed in a subsequent study through further idealized AGCM experiments.
Chapter 6

Asian summer monsoon response to forced changes of sea surface temperature


6.1 Introduction

As the two major forcing agents of human-induced climate change, greenhouse gases (GHGs) and anthropogenic aerosols lead to distinct different responses in the climate system. Aerosols partly offset the strong global warming signal caused by GHGs while inducing a hemispheric temperature contrast due to the spatial inhomogeneity of aerosol emissions (Bollasina et al. 2011; Lau and Kim 2017; Li et al. 2018). The opposing radiative forcing as well as the contrasting effects on the surface temperature cause different responses in atmospheric circulation, the global hydrological cycle, and regional hydroclimate (e.g. Xie et al. 2013; Singh 2016; Wang et al. 2016a). For the Asian monsoon region, in contrast to the temperature response dominated by GHGs, climate models suggest that the historical rainfall changes are dominated by the aerosol-induced drying trend during the 20th century, as shown in Chapter 3 (Li et al. 2015) using Coupled Model Intercomparison Project
- Phase 5 (CMIP5) models. However, the regional response to external forcing is still not well understood and uncertainties are poorly constrained, particularly for processes related to atmospheric circulation, ocean-atmosphere interactions, and aerosol effects (Turner and Annamalai 2012; Xie et al. 2015).

The climate response to GHGs or aerosols can be viewed as a combination of two components on different time scales: 1) the fast adjustment due to radiative effects as well as the interactions with clouds and the land surface; and 2) the slow response due to changes in sea surface temperature (SST) (Allen and Ingram 2002; Andrews et al. 2009). The fast response of precipitation and atmospheric circulation has been addressed in a number of studies for both GHG and aerosol forcing (Bony et al. 2013; Richardson et al. 2016; Li et al. 2018), as well as the possible compensating effects between the fast and slow responses (Shaw and Voigt 2015; Li and Ting 2017). The slow response involving ocean-atmosphere interactions has been shown to be important for the spatial pattern distribution of tropical rainfall (Xie et al. 2010; Chadwick et al. 2014) as well as the monsoon rainfall and circulation response to aerosol forcing over India and East Asia (Ganguly et al. 2012b; Kim et al. 2016). Furthermore, changes in SSTs contribute significantly to the uncertainties in regional precipitation and circulation projections (Ma and Xie 2013), as well as affecting possible future variations of monsoon variability (Li and Ting 2015).

For the slow response, an external forcing (GHGs/aerosols) can alter precipitation and atmospheric circulation through changes on both the mean SST and the spatial pattern. The overall warming of global SSTs induces an increase in atmospheric water vapor of 7% \(K^{-1}\) following the Clausius-Clapeyron relationship, while the global mean precipitation or evaporation only increases at the rate of 2-3% \(K^{-1}\) due to radiative constraints (Takahashi 2009; O’Gorman et al. 2012). As a result, the tropical atmospheric circulation weakens, leading to the “wetter-get-wetter, dry-get-drier” (Held and Soden 2006; Chou et al. 2009) pattern of change which intensifies the Asian summer monsoon rainfall. On the other hand, the spatial pattern of SSTs causes precipitation to increase where the SST warming exceeds the tropical mean and vice versa, following the “warmer-get-wetter” mechanism (Xie et al. 2010; Chadwick et al. 2013).

Compared to the relatively homogeneous spatial distribution of well-mixed GHGs, the
aerosol radiative forcing displays large meridional asymmetry concentrated over the northern hemisphere continents (Shindell et al. 2013). The asymmetric thermal forcing can lead to changes in the Hadley circulation and a southward shift of the intertropical convergence zone (ITCZ) (Kang et al. 2008; Ming and Ramaswamy 2009, 2011; Seo et al. 2014). Previous studies have shown that the hemispheric difference and meridional temperature gradient cause significant changes in monsoon circulation (Bollasina et al. 2011; Li et al. 2018), while for GHGs the mean SST change is more dominant than the spatial pattern in forcing tropical circulation and precipitation (He et al. 2014). Hill et al. (2015) show that the Hadley circulation weakens due to both mean and patterned SST changes driven by GHGs, while aerosols induce northward energy flux anomalies largely due to the spatial pattern. Furthermore, despite the difference in the spatial distribution of GHG and aerosol forcing, Xie et al. (2013) find that GHGs and aerosol induce similar responses in the spatial pattern of SSTs and oceanic rainfall with opposite signs. Studies have further shown that the SST spatial pattern dominates the uncertainty and intermodel spread of precipitation and atmospheric circulation in CMIP3 and CMIP5 models (Ma and Xie 2013; Kent et al. 2015; Chen and Zhou 2015).

Previous studies have highlighted the importance of improving the understanding of the various physical pathways by which GHGs and aerosols impact atmospheric circulation and precipitation for better constraining the uncertainties. In particular, the slow response of monsoon rainfall and circulation to anthropogenic forcing related to SST feedbacks has not been fully explored. Most of the aforementioned studies (e.g. Ma and Xie 2013; Hill et al. 2015) address global scale changes in precipitation and circulation, therefore not specifically focused on the monsoon region. For GHGs, Chapter 4 (Li and Ting 2017) has shown that the uniform SST warming induces a larger model spread in monsoon rainfall and circulation response over land compared to the fast adjustment. For aerosols, Chapter 5 (Li et al. 2018) suggests that while the fast adjustment dominates over eastern China and northern India, the slow response is important in altering the meridional circulation over the oceanic regions. However, Chapter 5 (Li et al. 2018) did not examine in detail the mechanisms driving the SST-induced changes. The motivation of this study is to gain a thorough mechanistic understanding of the responses of monsoon rainfall and circulation...
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to SST changes. The goal is to identify the relative importance of uniform and patterned SST changes in driving the monsoon response, as well as providing a direct comparison of GHG and aerosol forced changes.

We design a suite of idealized experiments using two atmospheric general circulation models (AGCMs) to delineate the effects of different SST forcing and examine the monsoon responses. We use two state-of-the-art AGCMs to help determine the model dependence and uncertainty of the responses. The chapter is organized as follows. Section 2 describes the model simulations and analysis methodology. In sections 3 and 4, we present results on the monsoon rainfall and circulation response to SST anomalies induced by GHGs and aerosols, respectively. In section 5, we compare the GHG and aerosol forced cases and discuss the uncertainties. The main conclusions are summarized in section 6.

6.2 Model simulations and methodology

6.2.1 Atmospheric general circulation models

We use two AGCMs to perform idealized experiments: the Community Atmosphere Model version 5.3 (CAM5) developed at the National Center for Atmospheric Research (NCAR), and the Geophysical Fluid Dynamics Laboratory Atmospheric Model version 3 (GFDL-AM3). CAM5 is the atmospheric component of the Community Earth System Model (CESM) version 1.2.0 with 30 vertical levels, coupled to an interactive land model (Community Land Model version 4, CLM4). GFDL-AM3 is the atmospheric component of the coupled GCM (GFDL-CM3) with 48 vertical levels, coupled to the land model LM3. Both models simulate full aerosol indirect effects with prognostic cloud schemes. The cam5 physics package is used for CAM5 which includes prognostic modal aerosols from the 3-mode version of the modal aerosol model (trop_mam3) chemistry package. GFDL-AM3 uses a prognostic cloud droplet number concentration scheme based on aerosol activation (Ming et al. 2006). Anthropogenic aerosol emissions are from the IPCC AR5 emission data set (Lamarque et al. 2010) that include emission for anthropogenic aerosols and precursor gases: sulfur dioxide ($SO_2$), primary organic matter (POM), and black carbon (BC). For both AGCMs, we use 1.9° latitude × 2.5° longitude horizontal resolution. The detailed
descriptions of the two models including the dynamical core and the physical parameterizations can be found in Neale et al. (2012) for CAM5 and Donner et al. (2011) for GFDL-AM3.

6.2.2 Experimental design

We design prescribed SST experiments to examine the influence of different SST forcing on monsoon changes. For all the experiments, both GHGs and aerosol emissions are fixed at year 1850 level, sea ice concentration is set at the 1951-2000 climatological values of the Hadley Centre Sea Ice and Sea Surface Temperature (HadSST) dataset (Rayner et al. 2003), ensuring that the response is solely due to the change in the imposed SSTs. For the control experiment (“CTRL”), we use climatological SST of 1951-2000 from the HadSST data. We derive the total SST anomaly from the historical GHG/aerosol-only simulations of the respective fully-coupled model (CESM-CAM5 and GFDL-CM3, available as part of the CMIP5), calculated as the climatological difference between 1981-2005 and 1861-1885. For both CESM-CAM5 and GFDL-CM3, the ensemble mean of the available three ensemble members are used. This total SST anomaly (total) is added to the CTRL climatology for the “ghgsst/aerosst” experiments. We then separate the total SST anomaly into two components: 1) the uniform warming/cooling component (uni), calculated as the annual mean of the tropical mean ($30^\circ S - 30^\circ N$) of total; 2) the spatial pattern component, calculated as the difference between the total anomaly and the uniform warming/cooling component. This decomposition method follows Ma and Xie (2013) using CAM3 and Hill et al. (2015) using GFDL-AM2.1. The list of experiments and their SST forcing are provided in Table 6.1. All the experiments are run for 60 years (after an initial 1 year spin-up). The climatological difference between each of the forced experiment and CTRL quantifies the response to the prescribed SST anomaly (i.e., total, uniform, and spatial pattern).

Figure 6.1 shows the annual mean of the total SST anomaly derived from the respective fully-coupled models for the (a, b) GHG and (c, d) aerosol experiments used in (a, c) CAM5 and (b, d) GFDL-AM3. The key statistics of the SST anomalies, including the global mean, northern hemisphere, southern hemisphere, and tropical mean are provided in Table 6.2. Both CESM-CAM5 and GFDL-CM3 show distinct warming in response to GHG forcing (Fig. 6.1a, b) and cooling to aerosol forcing (Fig. 6.1c, d). GFDL-CM3 has a stronger tem-
### Table 6.1: List of AGCM experiments and SST forcing.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>SST</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>HadSST 1951-2000 climatology (<em>clim</em>)</td>
</tr>
<tr>
<td>ghgsst</td>
<td><em>clim</em> + total SST anomaly from CGCM GHG-only simulations (<em>total</em>)</td>
</tr>
<tr>
<td>uniwarm</td>
<td><em>clim</em> + uniform warming of tropical mean of <em>total</em> (<em>uni</em>)</td>
</tr>
<tr>
<td>ghgpattern</td>
<td><em>clim</em> + <em>total</em> - <em>uni</em></td>
</tr>
<tr>
<td>aerosst</td>
<td><em>clim</em> + total SST anomaly from CGCM aerosol-only simulations (<em>total</em>)</td>
</tr>
<tr>
<td>unicool</td>
<td><em>clim</em> + uniform cooling of tropical mean of <em>total</em> (<em>uni</em>)</td>
</tr>
<tr>
<td>aeropattern</td>
<td><em>clim</em> + <em>total</em> - <em>uni</em></td>
</tr>
</tbody>
</table>

![Figure 6.1: Annual mean of global sea surface temperature anomalies (difference between 1981-2005 and 1861-1885) in (a, b) GHG-only and (c, d) aerosol-only simulations of (a, c) CESM-CAM5, and (b, d) GFDL-CM3. Units are K.](image)
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Table 6.2: Annual mean sea surface temperature anomalies (in K) for the global mean, northern hemisphere (NH), southern hemisphere (SH), tropical mean (30°S – 30°N), the difference between NH and SH, and the ratio between the hemispheric difference and the tropical mean for CESM-CAM5 and GFDL-CM3.

<table>
<thead>
<tr>
<th>Model / Experiment</th>
<th>Global</th>
<th>NH</th>
<th>SH</th>
<th>Tropics</th>
<th>NH - SH</th>
<th>(NH - SH) / Tropics</th>
</tr>
</thead>
<tbody>
<tr>
<td>CESM-CAM5 GHG</td>
<td>0.94</td>
<td>1.03</td>
<td>0.86</td>
<td>0.74</td>
<td>0.17</td>
<td>0.23</td>
</tr>
<tr>
<td>GFDL-CM3 GHG</td>
<td>1.34</td>
<td>1.85</td>
<td>0.93</td>
<td>1.27</td>
<td>0.92</td>
<td>0.72</td>
</tr>
<tr>
<td>CESM-CAM5 AERO</td>
<td>-0.44</td>
<td>-0.67</td>
<td>-0.26</td>
<td>-0.42</td>
<td>-0.41</td>
<td>0.98</td>
</tr>
<tr>
<td>GFDL-CM3 AERO</td>
<td>-0.67</td>
<td>-1.12</td>
<td>-0.33</td>
<td>-0.70</td>
<td>-0.79</td>
<td>1.13</td>
</tr>
</tbody>
</table>

Temperature response during the historical period for both GHGs and aerosols, with the annual mean globally averaged SST changes of 1.34K (GHGs) and -0.67K (aerosols) compared to 0.94K (GHGs) and -0.44K (aerosols) in CESM-CAM5. In terms of the spatial pattern, both models show stronger warming (cooling) in the northern hemisphere in response to GHGs (aerosols). This hemispheric temperature contrast is stronger in GFDL-CM3, with a 0.92K difference (0.17K for CESM-CAM5) for the GHG-only scenario and -0.79K (-0.41K for CESM-CAM5) for the aerosol-only scenario.

6.2.3 Moisture budget analysis

We analyze the atmospheric moisture budget to quantify the physical processes contributing to changes in precipitation, following Chapters 4 and 5 (Li and Ting 2017; Li et al. 2018) and Seager and Henderson (2013). The moisture budget equation expresses the following balance in pressure coordinates at steady state:

\[
\overline{P} - \overline{E} = -\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} u_q \, dp \approx -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^{K} u_k q_k \Delta p_k,
\]

where \( P \) is precipitation, \( E \) is evaporation, \( g \) is gravitational acceleration, \( \rho_w \) is the density of water, \( p \) is pressure and \( p_s \) surface pressure, \( u \) is the horizontal wind vector, \( q \) is specific humidity, \( k \) is the vertical level with a total of \( K \) (here \( K = 10 \), both models interpolated
to standard pressure levels and integrated from 1000hPa to 200hPa), $\Delta p$ is the pressure thickness. Overbars represent monthly mean values.

Using primes to denote departures from monthly means, the moisture flux convergence term can be separated into the mean moisture convergence ($MC$) and the sub-monthly transient eddies ($TE$):

$$
\mathbb{P} - \mathbb{E} \approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_k q_k} \Delta p_k - \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_k' q_k'} \Delta p_k.
$$

(6.2)

The contribution of the transient eddy component is small over the monsoon region (see Chapter 5 (Li et al. 2018)), thus we quantify the forced response using the mean moisture convergence term, and define

$$
\delta(\cdot) = \overline{(\cdot)}_F - \overline{(\cdot)}_C
$$

(6.3)

to represent the difference between the forced (subscript $F$) and control (subscript $C$) experiments, where the second overbar denotes the 30-year climatological mean. Then the change in the mean moisture convergence can be separated into a thermodynamic component due to changes in moisture ($\delta TH$), in which circulation ($u$) is fixed at the CTRL climatological value, and a dynamic component due to changes in atmospheric circulation ($\delta DY$), in which specific humidity ($q$) is fixed at the CTRL climatological value, derived as follows:

$$
\delta MC = \left( -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_k,F q_k,F} \Delta p_k,F \right) - \left( -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_k,C q_k,C} \Delta p_k,C \right)
\approx -\frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{u_k,C} \delta q_k \Delta p_k - \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \overline{\delta u_k q_k,C} \Delta p_k
= \delta TH + \delta DY.
$$

(6.4)

The quadratic term involving covariances of departures from climatological values of $u$ and $q$ is small compared to $\delta TH$ and $\delta DY$ (Li and Ting 2017; Li et al. 2018; Seager and Henderson 2013).
6.3 Monsoon response to GHG-induced SST anomalies

6.3.1 Meridional overturning circulation

Figure 6.2 shows the zonal mean vertical motion response to the imposed SST forcing in the GHG experiments for (a, c, e) CAM5 and (b, d, f) GFDL-AM3, depicting the global meridional overturning circulation. The Hadley circulation weakens in response to a uniform oceanic warming (Fig. 6.2c, d), with anomalous rising at 5°S – 30°S and sinking at the climatological rising branch north of the equator (contours show the climatology), while the location of the anomalous sinking slightly differs for the two models (closer to the equator for GFDL-AM3). The weakening of the Hadley cell is consistent with the large-scale “wet-get-wetter” argument (Held and Soden 2006) in which the tropical circulation tends to weaken to balance the stronger increasing rate of atmospheric water vapor (7% K\(^{-1}\) following the Clausius Clapeyron scaling) compared to precipitation (2-3% K\(^{-1}\)).

In the SST pattern experiment (Fig. 6.2e, f), both models show anomalous overturning circulation in the tropical region in response to the meridional temperature gradient. For CAM5, there is anomalous rising at 0° – 25°N and sinking at 0° – 10°S, whereas for GFDL-AM3 the anomalous overturning is centered at 5°S and much stronger in magnitude (note the difference of color scales).

The meridional overturning circulation response to the total SST forcing is shown in Fig. 6.2a, b, which represents a combination of the uniform and patterned components. The total response in CAM5 (Fig. 6.2a) is largely driven by the uniform warming component (Fig. 6.2c), whereas for GFDL-AM3 (Fig. 6.2b) the strong anomalous overturning circulation due to the spatial pattern (Fig. 6.2f) dominates the response near the equator. Furthermore, GFDL-AM3 displays a stronger atmospheric circulation response than CAM5 in all cases.

The change in the large-scale meridional overturning circulation in response to the SST forcing has direct consequences on the Asian monsoon. In the next two subsections, we present results of the regional monsoon response and examine the contributing thermodynamic and dynamic mechanisms.
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Figure 6.2: Zonal mean of vertical pressure velocity (ω) response to SST forcing in the GHG experiments: (a, b) total SST anomaly, (c, d) uniform warming, and (e, f) SST pattern for (a, c, e) CAM5, and (b, d, f) GFDL-AM3. Contours show the CTRL climatology, solid (dashed) contours denote positive (negative) values, indicating sinking (rising) motion. Contour intervals are 0.01 \( Pa \ s^{-1} \). The thick solid contour denotes 0. Stippling indicates statistically significant at 5% significance level using 2-tailed z-test. Units are \( 10^{-3} Pa \ s^{-1} \).
6.3.2 Monsoon rainfall and circulation

The summertime precipitation responses in the GHG experiments are shown in Fig. 6.3. Both models suggest a general increase of rainfall over the monsoon region in response to the GHG-induced total SST anomaly (Fig. 6.3a, b), mainly over southern China and the Arabian Sea for CAM5 (Fig. 6.3a), but over India, the Bay of Bengal, as well as Indochina and the western Pacific for GFDL-AM3 (Fig. 6.3b). This rainfall increase is largely attributable to the uniform oceanic warming particularly over land (Fig. 6.3c, d). The patterned SST anomaly induces a meridional pattern of rainfall response, which dominates the total response over the equatorial region particularly for GFDL-AM3: rainfall increases in the north but decreases in the south (Fig. 6.3e, f). The rainfall response is generally consistent with the large-scale meridional overturning circulation for both models (Fig. 6.2), in which the rainfall increase corresponds to anomalous rising and the rainfall decrease corresponds to anomalous sinking.

Figure 6.4 shows the 850hPa wind response in the GHG experiments. CAM5 (Fig. 6.4a) and GFDL-AM3 display clear discrepancy for the low-level monsoon circulation response to the total SST anomaly. CAM5 shows significant weakening of monsoon circulation, largely driven by the uniform oceanic warming over northern India and eastern China (Fig. 6.4c), consistent with the weakening of the tropical overturning circulation (Fig. 6.2a, c). However for GFDL-AM3, monsoon circulation enhances (and slightly shifts northward) over India and the Bay of Bengal while the response is weak over eastern China (Fig. 6.4b). The enhancement of the South Asian monsoon circulation is due to a combination of the uniform (Fig. 6.4d) and patterned (Fig. 6.4f) components. The discrepancy between the uniform warming experiments of the two models’ monsoon circulation response is likely to be related to the location of the anomalous large-scale overturning circulation: CAM5 shows anomalous sinking between $10^\circ N - 20^\circ N$ (Fig. 6.2c) while GFDL-AM3 (Fig. 6.2d) shows anomalous rising. Both models show enhanced monsoon circulation over southern India driven by the SST pattern component (Fig. 6.2e, f), however the stronger magnitude in GFDL-AM3 further contributes to the overall enhancement in the total response.
Figure 6.3: Precipitation response to SST forcing in the GHG experiments: (a, b) total SST anomaly, (c, d) uniform warming, and (e, f) SST pattern for (a, c, e) CAM5, and (b, d, f) GFDL-AM3. Stippling indicates statistically significant at 5% significance level using 2-tailed z-test.
Figure 6.4: 850mb wind response to SST forcing in the GHG experiments: (a, b) total SST anomaly, (c, d) uniform warming, and (e, f) SST pattern for (a, c, e) CAM5, and (b, d, f) GFDL-AM3. Arrows show the anomalous wind vectors and colors show the anomalous wind speed. Units are $m\ s^{-1}$. 
6.3.3 The moisture budget

To directly quantify the physical processes driving the change in monsoon rainfall, we examine the atmospheric moisture budget to identify the relative contributions of the thermodynamical and dynamical mechanisms. Figure 6.5 shows the moisture budget terms zonally averaged over the monsoon region \((60^\circ E - 140^\circ E)\) in which black, red, and blue lines are the mean moisture convergence \((\delta MC)\), the thermodynamic component \((\delta TH)\), and the dynamic component \((\delta DY)\), respectively. As a comparison, precipitation \((\delta P)\) is shown in green. The spatial distribution is shown in Fig. 6.6 for \(\delta TH\) and Fig. 6.7 for \(\delta DY\).

In all the experiments, the change in precipitation (green lines in Fig. 6.5) is predominantly explained by the change in the mean moisture convergence (black lines), indicating minor contributions from evaporation and the transient eddy term. The thermodynamic component features a strong increase of moisture convergence over the monsoon region in response to the total GHG-induced SST anomaly (red lines in Fig. 6.5a, b, Fig. 6.6a, b), almost entirely driven by the uniform oceanic warming for both models (red lines in Fig. 6.5c, d, Fig. 6.6c, d). The dynamic component in response to uniform warming shows mixed signals in CAM5 including decreased moisture convergence over northern India and the equatorial region, as well as a dipole structure over eastern China (blue line in Fig. 6.5c, Fig. 6.7c). On the other hand, GFDL-AM3 shows a stronger dynamical response to uniform warming with increased moisture convergence at \(10^\circ N - 20^\circ N\) and decreased moisture convergence over the equatorial region (blue line in Fig. 6.5d, Fig. 6.7d), which dominates the change in \(\delta MC\) (black line in Fig. 6.5d). The patterned SST experiment shows little thermodynamic response for both models (Fig. 6.6e, f), and thereby \(\delta MC\) is dominated by \(\delta DY\) (Fig. 6.5e, f): increased moisture convergence over \(5^\circ N - 25^\circ N\) for CAM5 and \(5^\circ S - 20^\circ N\) for GFDL-AM3; and decreased moisture convergence south of the equator and \(5^\circ S\) for CAM5 and GFDL-AM3, respectively. The dynamic change in the total SST forcing experiment (Fig. 6.7a, b) is a combination of the uniform and patterned components for both models. However \(\delta DY\) is relatively weak for CAM5 in response to the total SST anomaly thus \(\delta MC\) is predominantly driven by \(\delta TH\) (Fig. 6.5a). On the other hand, GFDL-AM3 displays stronger dynamical responses in all cases, and \(\delta MC\) in the total SST experiment is
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Figure 6.5: Response of the moisture budget terms zonally averaged over the monsoon region \( (60^\circ E - 140^\circ E) \) to SST forcing in the GHG experiments: (green) precipitation, (black) the mean moisture convergence \( (\delta MC) \), (red) the thermodynamic component of \( (\delta TH) \), and (blue) the dynamic component \( (\delta DY) \) for (a, b) total SST anomaly, (c, d) uniform warming, and (e, f) SST pattern for (a, c, e) CAM5, and (b, d, f) GFDL-AM3. Units are \( \text{mm day}^{-1} \).
Figure 6.6: Response of the thermodynamic component of the mean moisture convergence ($\delta T_H$) to SST forcing in the GHG experiments: (a, b) total SST anomaly, (c, d) uniform warming, and (e, f) SST pattern for (a, c, e) CAM5, and (b, d, f) GFDL-AM3. Units are mm day$^{-1}$. Stippling indicates statistically significant at 5% significance level using 2-tailed z-test.
Figure 6.7: As in Fig. 6.6, but for the dynamic component ($\delta DY$).
largely driven by $\delta DY$ while the thermodynamic enhancement also contributes (Fig. 6.5a).

The moisture budget decomposition indicates that the general increase of monsoon rainfall in response to GHG-induced SST anomalies is largely due to the thermodynamic enhancement resulting from the uniform oceanic warming. However, GFDL-AM3 shows a strong dynamical response, particularly due to the SST pattern, which further contributes to the overall meridional dipole structure of rainfall change in the equatorial region. While the thermodynamic responses are somewhat similar across the two models, the dynamic responses differ greatly between the two models, which highlight the large uncertainties in monsoon circulation response to GHG forcing (Li and Ting 2017).

6.4 Monsoon response to aerosol-induced SST anomalies

6.4.1 Meridional overturning circulation

We perform similar analysis on the aerosol experiments, beginning with the large-scale meridional overturning circulation response to the imposed aerosol-induced SST anomalies, shown in Fig. 6.8. For the uniform cooling experiment (Fig. 6.8c, d), both models largely display an enhancement of the Hadley circulation, the opposite response to the uniform warming in the GHG experiments (Fig. 6.2c, d). On the other hand, the SST pattern experiment (Fig. 6.8e, f) shows a reduction of the Hadley circulation in both models in response to the meridional temperature gradient, as depicted in Chapter 5 (Li et al. 2018). The location of the anomalous sinking slightly differs: CAM5 (Fig. 6.8e) has anomalous sinking over the entire climatological convection region ($0^\circ - 25^\circ N$) while for GFDL (Fig. 6.8f) the sinking branch is confined between $5^\circ S - 10^\circ N$. Different from the GHG experiments, the meridional circulation response to the total aerosol-induced SST forcing (Fig. 6.8a, b) is clearly dominated by the SST pattern component in both models. Similar to the GHG experiments, GFDL-AM3 also displays a stronger atmospheric circulation response in the aerosol experiments.

6.4.2 Monsoon rainfall and circulation

Figure 6.9 shows the monsoon rainfall response in the aerosol experiments. Both models
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Figure 6.8: As in Fig. 6.2, but for the aerosol experiments.
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Figure 6.9: As in Fig. 6.3, but for the aerosol experiments.
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display a clear meridional dipole structure in response to the total SST forcing (Fig. 6.9a, b), predominantly driven by the SST pattern (Fig. 6.9e, f). For CAM5, rainfall strongly decreases between 10°N – 25°N over India, Bay of Bengal, and the western Pacific (Fig. 6.9a, e). This strong reduction largely corresponds to the prominent weakening of monsoon circulation (Fig. 6.10a, e) as well as the anomalous sinking in the overturning circulation (Fig. 6.8a, e). For GFDL-AM3 (Fig. 6.9b, f), there is little change over land while the band of precipitation reduction is located near the equator, consistent with the location of the anomalous sinking branch (Fig. 6.8b, f). For both models, precipitation increases south of the equator around 10°S, where the overturning circulation shows anomalous rising motion (Fig. 6.8a, b). The uniform cooling component (Fig. 6.9c, d, Fig. 6.10c, d) contributes little to the overall rainfall and circulation responses, particularly for CAM5. The monsoon circulation response to uniform oceanic cooling in GFDL-AM3 (Fig. 6.10d) shows a southward shift with reduction over northern India and eastern China and enhancement over southern India. The magnitude of monsoon circulation response over land in the aerosol case is much weaker in GFDL-AM3 compared to CAM5, explaining the weak land precipitation responses.

6.4.3 The moisture budget

We further examine the moisture budget responses in the aerosol experiments, as illustrated in Figs. 6.11-6.13. For both models, the thermodynamic component shows a slight decrease of moisture convergence over the monsoon region, mostly driven by the uniform cooling component whereas the SST pattern experiment shows very little change (red lines in Fig. 6.11, Fig. 6.12). In all cases, the dynamic component (blues lines in Fig. 6.11) dominates the change in the mean moisture convergence (black) as well as precipitation (green). In terms of the spatial distribution, despite the similar increase of moisture convergence south of the equator, the two models exhibit significant discrepancy over the monsoon region. For CAM5, similar to the rainfall and wind responses, $\delta DY$ in the total SST forcing experiment (Fig. 6.13a) shows strong reduction of moisture convergence at 10°N – 25°N but an increase over eastern China and northeastern India. This dynamical response is almost entirely dominated by the SST spatial pattern (Fig. 6.13e) while the uniform cool-
Figure 6.10: As in Fig. 6.4, but for the aerosol experiments.
Figure 6.11: As in Fig. 6.5, but for the aerosol experiments.
Figure 6.12: As in Fig. 6.6, but for the aerosol experiments.
Figure 6.13: As in Fig. 6.7, but for the aerosol experiments.
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ing (Fig. 6.13c) contributes very little. For GFDL-AM3, the total dynamical response (Fig. 6.13b) is weak over land, but shows a strong decrease in moisture convergence near the equator, driven by the spatial pattern component (Fig. 6.13f). The uniform oceanic cooling in GFDL-AM3 (Fig. 6.13d) contributes little over most of the monsoon region but induces a meridional dipole structure over Indonesia, further enhancing the dynamical weakening around 5°S. The overall meridional dipole responses of δDY and δMC as well as the discrepancy between the models are evident in the monsoon region zonally averaged pattern (Fig. 6.11a, b, e, f). Both models indicate an increase in moisture convergence and precipitation south of 5°S, but the decrease in moisture convergence is located much further north in CAM5 compared to GFDL-AM3. Therefore, the rainfall response in the major land region exhibits clear discrepancy between the two models, particularly over India.

6.5 Discussions

6.5.1 Comparison of GHG and aerosol experiments

As shown in sections 3 and 4, the tropical meridional overturning circulation and the Asian monsoon system respond significantly different to the SST forcing induced by GHGs and aerosols. In this section, we compare the GHG and aerosol forced cases and discuss the possible contributing mechanisms.

Compared to aerosols, GHGs cause a stronger global SST response but also more uniformly distributed (Fig. 6.1). The ratio of the hemispheric SST difference and the tropical mean warming, which evaluates the relative importance of SST pattern compared to the mean change, is 0.23 (CAM5) and 0.72 (GFDL-AM3) for GHGs compared to 0.98 (CAM5) and 1.13 (GFDL-AM3) for aerosols (Table 6.2). This suggests that the SST spatial pattern is more prominent in the aerosol-forced case and thus more likely to dominate changes in the subsequent rainfall and circulation responses. Although interestingly, despite the homogeneous radiative forcing of GHGs, global SSTs still indicate stronger warming in the northern hemisphere particularly for GFDL-AM3.

In response to a uniform SST change, the meridional overturning circulation weakens under GHG-induced warming and enhances under aerosol-induced cooling, consistent with
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the expected outcome following the thermodynamic scaling argument. In response to the SST pattern, the Hadley circulation weakens in the cooler hemisphere (Southern Hemisphere for GHGs and Northern Hemisphere for aerosols) and strengthens in the warmer hemisphere, indicating a spatially similar but opposite in sign response as suggest by Xie et al. (2013). As a combination of uniform and patterned changes, the total circulation response to SST forcing largely depends on the relative strength of the two components. For GHGs, the uniform warming dominates, leading to the overall weakening of atmospheric overturning circulation. For aerosols, the spatial pattern dominates, which also results in an weakening of the meridional circulation. Therefore for CAM5, in which the ratio between the hemispheric temperature contrast and the mean change is clearly higher for GHGs than aerosols, the total circulation response is largely similar for the GHG and aerosol forced cases, including the global scale meridional circulation (Figs. 6.2a and 6.8a) and the regional monsoon circulation (Figs. 6.4a and 6.10a). However for GFDL-AM3, in which GHGs also cause a strong meridional temperature gradient, the total circulation response is largely dominated by the spatial pattern component for both GHGs and aerosols, leading to the largely different and opposing circulation patterns as shown in Figs. 6.2b and 6.8b, 6.4b and 6.10b.

The regional monsoon circulation response is largely consistent with the global meridional circulation, however does show discrepancies from the theoretical outcome as well as between the two models, indicating higher uncertainty on the regional scale. On the other hand, monsoon rainfall is not only driven by the circulation change, but also largely influenced by the thermodynamic mechanism. For both GHGs and aerosols, thermodynamic changes are dominated by the increase/decrease of moisture induced by the mean SST warming/cooling. As a result, since the uniform SST component plays a more important role in the GHG case, the thermodynamic mechanism tends to be more prominent in determining the monsoon rainfall responses to GHG-induced SST anomalies. While in the aerosol case, monsoon rainfall is almost entirely driven by atmospheric circulation.
6.5.2 Uncertainties

Although the modeled results are generally consistent with the large-scale thermodynamical and dynamical changes one would expect from the imposed thermal forcing, CAM5 and GFDL-AM3 do display apparent discrepancies in regional responses. The discrepancy is likely to be a combination of uncertainties originating from the model’s own physics, the experimental setup, as well as the actual climate response. Here we provide a brief discussion on the possible factors contributing to the uncertainties in our model simulations.

In our experiments, we have imposed SST forcing derived from the corresponding fully-coupled model for CAM5 and GFDL-AM3 separately, which ensures the consistency between the model and the SST anomaly. However, as shown in Fig. 6.1, the SST anomaly in CAM5 and GFDL-AM3 display significant differences in both the magnitude and spatial pattern, particularly for the GHG case. Previous studies have demonstrated that the climate response to an external thermal forcing can be sensitive to the structure of the imposed SST forcing (Kang and Held 2012; Kang and Xie 2014). We have not fully explored the sensitivity of our results to the different SST anomalies. Such sensitivity studies will help identify how much of the difference between CAM5 and GFDL-AM3 can be attributed to the SST forcing. Nevertheless, the SST anomalies induced by aerosols are spatially similar for the two models while inducing larger discrepancy in the response compared to GHGs, suggesting that the pattern of the imposed SST forcing is unlikely to be the dominant factor.

In addition, the baseline SST climatology for the experiments has been prescribed using observational data rather than the modeled climatology to avoid model bias. The experiments have been performed using AGCMs without the coupling to the ocean. The separation of the total SST anomaly into the uniform and patterned components also do not include the possible nonlinear interactions. These factors and the potential uncertainties they might introduce should be taken into consideration when interpreting our results. Further sensitivity studies are needed to provide a better constraint on the uncertainties of the regional climate response to SST changes.
6.6 Conclusions

In this chapter, using two comprehensive AGCMs (CAM5 and GFDL-AM3) with prescribed SSTs, we have investigated the response of the Asian monsoon rainfall and circulation to SST anomalies induced by both GHGs and aerosols. GHGs induce a stronger magnitude of the mean SST change compared to aerosols, while the hemispheric asymmetry is larger for aerosol-induced changes. We have separated the total SST change into a uniform warming / cooling component, and a spatial pattern component predominantly characterized by the meridional temperature gradient. The total monsoon rainfall and circulation response to SST forcing largely depends on the relative importance of the two components.

For GHG-induced SST anomalies, the uniform warming component contributes significantly to the total monsoon rainfall increase through the thermodynamic enhancement due to the increase in atmospheric moisture. The atmospheric circulation response is determined by the relative magnitude and location of the weakened Hadley circulation due to uniform warming and enhanced Hadley circulation due to the SST pattern. Thus the total circulation response in CAM5 is dominated by the uniform warming, while the SST pattern component is more prominent for GFDL-AM3 in driving enhanced monsoon circulation.

For aerosol-induced SST anomalies, the overall monsoon rainfall response is dominated by the SST pattern component via changes on atmospheric circulation. The strong meridional temperature gradient leads to weakened monsoon circulation, which drives a north-south decrease-increase dipole structure of rainfall change. However the two models show significant discrepancy for the latitude location of the northern drying, possibly related to both the SST forcing and model uncertainty.

Our results have demonstrated that the effect of SST changes on monsoon rainfall and circulation is sensitive to the partitioning between the mean and patterned components, which may contribute to model uncertainty and spread in coupled GCMs. The idealized decomposition helps to identify the fundamental physical processes entangled in the total SST-driven response, as well as quantify the similarities and differences for the GHG and aerosol forced cases. Our study contributes to a better mechanistic understanding of the monsoon system response to external forcing, essential for the improvement of the projection
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for regional hydroclimate change.
Chapter 7

Conclusions

7.1 Summary of major findings

Using observations, large ensemble sets from Coupled Model Intercomparison Project - Phase 5 (CMIP5) models, and a combination of statistical analysis and idealized atmospheric general circulation model (AGCM) experiments, this dissertation examines the response of the Asian summer monsoon system to greenhouse gases (GHGs) and anthropogenic aerosols for both the historical period and future projections. GHGs and aerosols induce significantly different changes in monsoon rainfall through both thermodynamical and dynamical processes. These changes can be separated into the fast adjustments related to radiation and cloud processes and the slow response due to changes in sea surface temperature (SST). The major findings and conclusions addressing the questions raised in Chapter 1 are summarized as follows.

- What are the effects of external forcing on monsoon variability?

The Asian monsoon interannual variability is strongly connected to the El Niño-Southern Oscillation (ENSO). Chapter 2 examines the ENSO-monsoon relationship in the 20th and 21st century using signal-to-noise (S/N) maximizing empirical orthogonal function (EOF) analysis to separate the anthropogenically forced and naturally varying components. During the 20th century, natural variability plays a dominant role in the ENSO-monsoon relationship. However in the 21st century, external
forcing leads to enhanced monsoon rainfall associated with SST warming, which may contribute to a slightly weakened ENSO-monsoon relation in the future.

- What are the changes and mechanisms of the Asian summer monsoon in response to anthropogenic forcing for the historical period and future projections?

Chapters 3 quantifies the thermodynamic and dynamic mechanisms driving the changes of the Asian summer monsoon in observations and fully-coupled GCMs, using the atmospheric moisture budget. CMIP5 models indicate a predominantly drying Asian monsoon in the 20th century, while monsoon rainfall strongly enhances in the 21st century under the representative concentration pathway 8.5 (rcp8.5) scenario. The contrasting drying trend in the CMIP5 historical simulations and the wetting trend in the rcp8.5 projections can be predominantly explained by the strong aerosol-induced dynamical weakening during the 20th century and the thermodynamic enhancement due to GHGs in the 21st century.

- What are the relative roles of the fast adjustments and the slow response (via uniform change and SST spatial pattern) for GHGs and aerosols?

Chapter 3 indicates that the monsoon circulation response to GHGs is much weaker compared to aerosols. This is largely related to the different effects of the fast and slow responses to GHGs and aerosols, presented in Chapters 4 and 5 using multiple AGCMs with prescribed SSTs, respectively. For GHGs (Chapter 4), the fast and slow monsoon circulation changes largely oppose each other, leading to an overall weak response and large inter-model spread. For aerosols (Chapter 5), however, the strongly weakened monsoon circulation over land due to aerosols is largely driven by the fast adjustments related to aerosol-radiation and aerosol-cloud interactions.

The slow response due to the SST anomalies induced by both GHGs and aerosols is further investigated in Chapter 6 using idealized experiments with two AGCMs - the Community Atmosphere Model version 5.3 (CAM5) and the Geophysical Fluid Dynamics Laboratory Atmospheric Model version 3 (GFDL-AM3). For GHG-induced SST anomalies, the uniform warming component contributes largely to the monsoon rainfall increase through the thermodynamic enhancement due to moisture increase.
CHAPTER 7. CONCLUSIONS

For aerosol-induced SST anomalies, the overall monsoon rainfall response is dominated by the SST pattern component (dominated by the hemispheric temperature contrast) via changes on atmospheric circulation, leading to weakened monsoon circulation.

- What are the possible factors contributing to the uncertainty of monsoon changes in climate model simulations?

Chapters 3 to 6 indicate that CMIP5 models show large uncertainties and model spreads of externally-forced monsoon changes, as well as significant discrepancies with observations. Furthermore, while Chapter 2 shows that CMIP5 models simulate the ENSO-monsoon spatial structure reasonably well when using the multi-model mean, there are still large variations across the model ensemble. Given the dominating role of aerosols in historical modeled monsoon changes, this raises the question of whether aerosol effect is appropriately represented in the CMIP5 models, particularly aerosol-cloud interactions. Furthermore, compared to the general consensus on the thermodynamic response, the dynamical mechanism related to atmospheric circulation contributes to higher uncertainty and larger model spreads.

7.2 Implications

The summer monsoon brings over 80% of the annual rainfall over Asia, with critical importance for local and global climate. Analyzing and understanding the characteristics of monsoon change has important implications for climate predictions and adaptations, as well as socioeconomic development and human well-being. This thesis provides a detailed analysis of the multiple physical processes entangled in the human-induced changes of the Asian monsoon system, advancing our mechanistic understanding of the effects of external forcing on monsoon rainfall and circulation and the associated uncertainties. Results indicate that the reversal of decreased historical monsoon rainfall and the future increase can be attributed to the opposing roles of aerosols and GHGs. Aerosols dominate historical monsoon rainfall decrease mainly through the dynamic mechanism (changes in atmospheric circulation). On the other hand, GHGs dominate future monsoon rainfall increase mainly through the thermodynamic mechanism (changes in moisture) due to uniform SST warm-
CHAPTER 7. CONCLUSIONS

The various physical pathways evolving on different time scales (radiative and cloud effects versus SST change) contribute significantly to the discrepancies and models spreads. In particular, the uncertainty in aerosol representations and aerosol-induced responses including aerosol indirect effects, are poorly constrained for the current generation of global climate models. Given the future scenario in which aerosol forcing gradually phases out during the 21st century, it is urgent to address the question of how the decreasing aerosol emissions will affect the monsoon system alongside the increasing GHGs. Furthermore, extreme events pose significant climate risk and impacts, however the responses of monsoon rainfall extremes to GHGs/aerosols have not been fully explored. Further studies are needed in order to obtain a better physical understanding and improve the predictions of future monsoon change.

7.3 Future directions

7.3.1 Aerosol-cloud interactions

As illustrated throughout this thesis, GCMs show significant uncertainties in simulating the externally forced changes of monsoon rainfall and circulation, particularly the aerosol effects. This is largely related to the lack of adequate representation of aerosol and cloud processes in coarse-resolution GCMs. It is important to improve the understanding of aerosol-cloud-precipitation interactions and their representation in global climate models. High-resolution model simulations in the new generation of GCMs can be utilized to obtain a better estimate of these processes and determine the influence of model resolution. Furthermore, natural aerosols are also a major source of uncertainty in the effective radiative forcing by aerosols (Carslaw et al. 2013; Ghan et al. 2013). Evaluating and improving the representation of the processes involving natural aerosols and clouds in a GCM are also essential for constraining the uncertainty for examining the impacts of anthropogenic aerosols. These topics will be explored in future research to provide a better constraint on cloud and aerosol parameterizations in GCMs, and a more confident estimate of the effects of anthropogenic aerosols on current and future hydroclimate.
7.3.2 Extreme events

Hydroclimate extremes and the associated flood/drought can cause disastrous human and economic losses. During the recent decades, precipitation extremes in many regions have significantly increased for both frequency and intensity (Donat et al. 2013). This is expected to continue in the future (O’Gorman 2012; Kharin et al. 2013) due to both increasing greenhouse warming and declining aerosol emissions (Wang et al. 2016b) because aerosols and GHGs tend to have opposite effects on the changes of extreme events (Mascioli et al. 2016). Similar to the mean precipitation, the sensitivity of precipitation extremes to GHGs and aerosols may be largely different. Lin et al. (2016) find that the efficiency of aerosols in altering land rainfall extremes can be up to 10 times of those due to GHGs over eastern and southern Asia.

Precipitation extremes are strongly influenced by stochastic atmospheric and oceanic variability on multiple time scales, as well as external forcing due to human activities. Large uncertainties exist in future projections of precipitation extremes, along with a lack of understanding of the underlying physical processes. In particular, for the monsoon regions including south Asia and west Africa, low-pressure systems such as lows and depressions contribute up to 80% to the total summer rainfall (Hurley and Boos 2015). However, the response of monsoon depressions to external forcing has not been fully explored.

For future research, I plan to extend my current analysis in monsoon variability and change to extreme rainfall events. The goal will be to improve the mechanistic understanding of the effects of GHGs and aerosols on monsoon depressions and rainfall extremes, and quantify the uncertainty. The results will improve our scientific understanding as well as aid decision making in both public and private sectors of weather and climate extremes in a changing climate.
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