DYNAMICS AND UNCERTAINTIES OF GLOBAL WARMING PATTERNS: SEA SURFACE TEMPERATURE, PRECIPITATION, AND ATMOSPHERIC CIRCULATION

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAI‘I AT MĀNOA IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

IN

METEOROLOGY

DECEMBER 2012

By

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Acknowledgments

I would like to express my gratitude to my academic advisor, Dr. S.-P. Xie, for his generous support and careful guidance throughout my Ph.D. program. I would also thank my dissertation committee members for their constructive comments and suggestions. Professors and researchers of the Department of Meteorology, Oceanography, and International Pacific Research Center, University of Hawaii are highly appreciated for their skillful teaching and helpful discussion. My fellow students and postdocs are also gratefully acknowledged. Finally, I would like to thank my parents, my wife and my son for their continuous care and warm encouragements during my Ph.D. study.

I acknowledge various modeling groups (listed in Tables 1.1 and 1.2) for producing and providing their output, the Program for Climate Model Diagnostics and Intercomparison (PCMDI) for collecting and archiving the CMIP3 and CMIP5 (CFMIP2) multi-model data, the WCRP’s Working Group on Coupled Modeling (WGCM) for organizing the analysis activity, and the Office of Science, U.S. Department of Energy for supporting these datasets in partnership with the Global Organization for Earth System Science Portals. I wish to thank M. Webb and M. Ringer for sharing CFMIP SUSI simulations, and NCAR for the CAM3.1 codes and related data. I thank GFDL for providing outputs of their ensemble integrations, and M. Watanabe for releasing the LBM codes. Also acknowledged are Y. Kosaka for providing her AMIP simulations with GFDL AM2.1, and H. Tokinaga for releasing
the WASWind dataset. The Ferret program was used for analysis and graphics. This work is supported by NSF, NOAA, NASA, and JAMSTEC.
Abstract

Precipitation and atmospheric circulation changes in response to global warming have profound impacts on the environment for life but are highly uncertain. This study investigates fundamental mechanisms controlling these changes and relates them to the effects of sea surface temperature (SST) change, using Coupled Model Intercomparison Project simulations. The SST warming is decomposed into a spatially uniform SST increase (SUSI) and deviations from it.

The SST pattern effect is found important in explaining both the multi-model ensemble mean distribution and inter-model variability of rainfall change over tropical oceans. In ensemble mean, the annual rainfall change follows a “warmer-get-wetter” pattern, increasing where the SST warming exceeds the tropical mean, and vice versa. Two SST patterns stand out: an equatorial peak that anchors a local precipitation increase, and a meridional dipole mode with increased rainfall and weakened trade winds over the warmer hemisphere. These two modes of inter-model variability in SST account for up to one third of inter-model spread in rainfall projection.

Tropospheric warming follows the moist adiabat in the tropics, and static stability increases globally. A diagnostic framework is developed based on a linear baroclinic model (LBM) of the atmosphere. The mean advection of stratification change (MASC) by climatological vertical motion, often neglected in interannual variability, is an important thermodynamic term for global warming. MASC and SST-
pattern effects are on the same order of magnitude in LBM simulations. Once MASC effect is included, LBM shows skills in reproducing general circulation model (GCM) results by prescribing latent heating diagnosed from the GCMs.

Common to all GCMs, MASC causes both the Hadley and Walker circulation to slow down as articulated by previous studies. The weakening of the Walker circulation is robust across models as the SST pattern effect is weak. The Hadley circulation change, by contrast, is significantly affected by SST warming patterns. As a result, near and south of the equator, the Hadley circulation change is weak in the multi-model ensemble mean and subject to large inter-model variability due to the differences in SST warming patterns, explaining up to four fifth of the inter-model variability in changes of the overturning circulation.
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Chapter 1

Introduction

Human societies formed where freshwater was readily available. In many parts of the world, population increase and economic development have stretched water resources to near the breaking point, rendering societies ever more vulnerable to rainfall variability and change. The looming global warming is almost certain to change the distribution of water resources (Zhang et al. 2007; Held et al. 2005; Seager et al. 2007), posing serious socioeconomic and security challenges that have profound impacts on the environment for life on Earth. Whereas the effects of the slow changes in precipitation patterns are obvious, their causes are illusively uncertain and poorly understood, because of short observations and large natural variability.

The large-scale atmospheric circulation interacts with precipitation and is essential for moisture and energy transports, tropical cyclone (TC) development, and ocean/land-atmosphere interactions. In the tropics, where the synoptic eddy effects are weak, the tropospheric circulation is primarily generated by the uneven distribution of diabatic heating/cooling, e.g., convective latent heating in convergence zones. Climatologically, these forcing terms are nearly in balance with vertical advection (e.g., Rodwell and Hoskins 1996). In global warming, vertical advection and diabatic forcing change, and the large-scale circulation must alter accordingly to regain the thermodynamic balance.
The enormity of the problem in the hydrological cycle calls for investigations into fundamental dynamics governing such changes, especially those in response to increasing greenhouse gases (GHG). For that this study analyzes general circulation model (GCM) simulations (Tables 1.1 and 1.2) in the World Climate Research Program’s (WCRP’s) Coupled Model Intercomparison Project (CMIP) phase 3 and phase 5 (Meehl et al. 2007). In model projections for climate change during the 21st century, global-mean rainfall increases at a much slower rate (2-3% per degree surface warming) (Held and Soden 2006) than atmospheric moisture content (7% K^{-1}). Meanwhile, dry static stability increases as the tropospheric warming follows the moist adiabatic profile (Knutson and Manabe 1995). These differences imply a slow-down of tropical circulation (Vecchi and Soden 2007a), a prediction confirmed for the Walker cell (Vecchi et al. 2006), though satellite-based microwave measurements question this slower increase rate (Wentz et al. 2007).

These thermodynamic constraints do not explain why the Walker cell is preferably weakened rather than the Hadley cell. Observations (Hu and Fu 2007) and general circulation model (GCM) simulations (Lu et al. 2007; Frierson et al. 2007; Johanson and Fu 2009) show a robust poleward expansion of the Hadley circulation. The intensity change of the Hadley circulation (Ma et al. 2012), however, is not as robust across models as that of the Walker cell (Vecchi and Soden 2007a). The difference in shape and robustness of these two overturning circulation responses has not been thoroughly examined in the literature, calling for research on the source of uncertainty.

Precipitation change is highly uneven in space. Its spatial variability is greater than the global mean by a factor of four (Table 3.1). Research into patterns of precipitation change starts from a “wet-get-wetter” view. It predicts that rainfall increases in the core of existing rainy regions and decreases in current dry areas,
based on an argument of intensified moisture advection due to atmospheric warming (Neelin et al. 2003; Chou and Neelin 2004; Chou et al. 2009; Held and Soden 2006; Seager et al. 2010). An “upped-ante” mechanism is raised to explain the reductions in precipitation at the convective margins (Neelin et al. 2003; Chou and Neelin 2004; Chou et al. 2009). In this mechanism, “a warm troposphere increases the value of surface boundary layer moisture required for convection to occur. In regions of plentiful moisture supply, moisture simply rises to maintain precipitation, but this increases the moisture gradient relative to neighboring subsidence regions. Reductions in rainfall then result for those margins of convection zones that have strong inflow of air from the subsidence regions and less frequently meet the increased ‘ante’ for convection.” The destabilizing effects of increased low-level moisture were once suggested to enhance tropical convection (Lindzen 1990) but not supported by simulations with one-dimensional radiative convective models (Betts and Ridgway 1989; Betts 1998) and GCMs (Knutson and Manabe 1995; Held and Soden 2006).

In the “wet-get-wetter” view, a spatial-uniform sea surface temperature (SST) increase (SUSI) is implicitly assumed, neglecting spatial variations in surface warming and the associated wind change. SST warming, however, displays considerable variations in space (Xie et al. 2010) with robust and coherent seasonal variability (Sobel and Camargo 2011). A “warmer-get-wetter” paradigm emerges, casting the relative SST warming ($T^*$), defined as deviations from the tropical mean SST increase, as important for regional changes in TC activity (Vecchi and Soden 2007b; Knutson et al. 2008; Vecchi et al. 2008; Zhao and Held 2012), precipitation (Xie et al. 2010; Sobel and Camargo 2011), and atmospheric circulation (Ma et al. 2012; Gastineau et al. 2009). Upper tropospheric warming is nearly spatially uniform.
in the tropics due to fast wave actions, so convective instability is largely determined by spatial variations in SST warming (Johnson and Xie 2010).

This SST-pattern control on rainfall reorganization can be illustrated by a comparison of two CMIP3 model simulations (Table 1.1) forced with the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions Scenarios (SRES) A1B. Figure 1.1 shows that oceanic rainfall changes are quite different between the GFDL CM2.0 and UKMO HadCM3. Over the tropical Pacific, the CM2.0 features a pronounced increase in rainfall on the equator (especially the western side) and drastic reduction on the sides. By contrast, the HadCM3 rainfall change is characterized by an inter-hemispheric asymmetry with increased rainfall north of the equator.

Spatial correlation \((r)\) in tropical \((20^\circ S-20^\circ N)\) rainfall change between the two models is only \(-0.03\). Remarkably, the disparity in rainfall response to the A1B scenario can be explained by the inter-model difference in SST warming (Fig. 1.1c). Positive SST differences (CM minus HadCM) are found collocated with enhanced rainfall in all three tropical oceans. Indeed, the spatial correlation between SST and rainfall differences reaches 0.56, illustrating the importance of SST warming patterns.

The ensemble-mean SST warming in CMIP3 simulations features an equatorial peak (Liu et al. 2005), which was attributed to several processes. Liu et al. (2005) suggests that equatorial wind reduction prohibits evaporation, favoring the enhanced warming there. Xie et al. (2010) proposes that weak Newtonian cooling rate in latent heating due to low wind speed and high relative humidity on the equator plays an important role, which is consistent with the model study of Seager and Murtugudde (1997). Anomalous oceanic advection from the warm pool region is also suggested to warm the western equatorial Pacific (Xie et al. 2010).
This equatorial peak warming is often characterized as El Nino-like. However, the zonal mean tropospheric warming patterns, changes in zonal wind shear, and Hadley cell are all quite different from El Nino (Lu et al. 2008). This appears due to the difference in the tropical mean SST warming relative to the spatial patterns between El Nino and global warming. In a 10-member ensemble simulation with the National Oceanic and Atmospheric Administration (NOAA) Geophysical Fluid Dynamics Laboratory (GFDL) climate model (CM2.1) for 1996-2050 under SRES A1B (Xie et al. 2010), the tropical (20°S-20°N) mean SST warming is 1.12 K with a spatial standard deviation of 0.21 K (19% of the tropical mean). By contrast, El Nino events in the same model feature an SST spatial standard deviation of 0.76 K, 140% of the tropical mean warming of 0.55 K (not shown). Thus, El Nino-Southern Oscillation (ENSO) is dominated by SST patterns while global warming by the tropical mean.

Figure 1.2 illustrates the relative importance of tropical mean SST warming vs. its patterns for tropospheric circulation change by comparing the annual mean results of the GFDL CM2.1 A1B simulation with its atmospheric model (AM2.1) forced by a SUSI of 2 K. Tropospheric warming patterns, defined as deviations from the tropical mean, are similar between two runs, with a spatial correlation coefficient (r) of 0.59 (Table 1.3). Both cases feature maxima in the subtropics and a minimum in the Indo-Pacific warm pool that extends to the Intertropical Convergence Zone (ITCZ) and South Pacific Convergence Zone (SPCZ). Fu et al. (2006) showed the enhanced subtropical warming from satellite observations and suggested that it pushes the tropospheric jet streams poleward, contributing to the Hadley cell expansion. The zonal mean warming patterns (Fig. 1.3) are very similar between SUSI and A1B (r = 0.91), both featuring an elevated maximum warming at 300 hPa, a result of moist adiabatic adjustment (Knutson and Manabe 1995). In thermal-wind balance with
temperature, the 300-850 hPa wind shear decreases with anomalous easterly shear in the tropical Pacific (Fig. 1.2). The shear response is similar between SUSI and A1B (r = 0.57 for zonal wind shear).

Apparent differences from the SUSI run include the development of meridional asymmetry in the A1B run over the eastern tropical Pacific (Figs. 1.2b and 1.3b), with southerly cross-equatorial wind shear. Besides, wind shear changes in the tropical Indian Ocean are opposite between the two runs. These differences are primarily induced by SST patterns.

While previous studies (e.g. Held and Soden 2006) about hydrological cycle change focus on global mean budget, the spatial patterns are mainly concerned here. It appears that the SST patterns play a key role in shaping regional precipitation response to global warming, while the SUSI is important for tropospheric circulation change. The present study investigates the effects of SST warming patterns on changes in tropical precipitation and circulation. It extends previous studies by analyzing a large number of coupled model simulations in the CMIP databases and by using atmospheric SUSI simulations to isolate SST pattern effects. The “warmer-get-wetter” mechanism is shown to account for much of the spatial variations in tropical rainfall response to GHG forcing as represented by the multi-model ensemble mean. As Figure 1.1 illustrates, rainfall projection varies greatly among models, and the causes of this uncertainty have not been fully explored. The differences in spatial patterns of SST warming are an important source of inter-model diversity in tropical rainfall projection, highlighting the need to study SST warming patterns as an ocean-atmosphere interaction problem.

As for the tropical tropospheric circulation response to global warming, a diagnostic framework is designed to identify robust dynamical balance and simulate
major features of circulation change. The effect of increased static stability is shown important for the slow-down of tropical circulation in SUSI. This mechanism is robust among models and can enable a linear baroclinic model (LBM) to simulate global warming patterns. Finally, the response of atmospheric overturning circulation to global warming is examined and again SST patterns are identified as an important cause of variability among models, especially for the Hadley circulation change.

The rest of the dissertation is arranged as follows. Chapter 2 describes the data and methods. Chapter 3 examines the relationship between change patterns of SST and precipitation in the CMIP3 ensemble mean and inter-model variations, and compares them with CMIP5 results. Chapter 4 diagnoses the mechanisms for tropospheric circulation change. Response of atmospheric overturning circulation to global warming is discussed in Chapter 5. Chapter 6 gives conclusions with discussion.
Table 1.1. The WCRP CMIP3 A1B models used in this study. Monthly output is directly adopted except for the listed variables converted from daily data, including zonal wind ($U$), meridional wind ($V$), and surface winds ($U_{sfc}$, $V_{sfc}$). All changes are scaled by tropical mean (20°S-20°N) SST changes for the specific models.

<table>
<thead>
<tr>
<th>Model name</th>
<th>Country</th>
<th>Atmospheric resolution</th>
<th>Oceanic resolution</th>
<th>Converted variables</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 BCCR BCM2.0</td>
<td>Norway</td>
<td>T63 L31</td>
<td>2.4° × 2.4° (0.8°)</td>
<td>$U_{sfc}$, $V_{sfc}$</td>
</tr>
<tr>
<td>2 CGCM3.1 T47</td>
<td>Canada</td>
<td>T47 L31</td>
<td>1.85° × 1.85° L29</td>
<td></td>
</tr>
<tr>
<td>3 CGCM3.1 T63</td>
<td>Canada</td>
<td>T63 L31</td>
<td>1.4° × 0.94° L29</td>
<td></td>
</tr>
<tr>
<td>4 CNRM CM3</td>
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<td>T63 L45</td>
<td>2° × 0.5° L31</td>
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</tr>
<tr>
<td>5 CSIRO Mk3.0</td>
<td>Australia</td>
<td>T63 L18</td>
<td>1.875° × 0.84° L31</td>
<td></td>
</tr>
<tr>
<td>6 CSIRO Mk3.5</td>
<td>Australia</td>
<td>T63 L18</td>
<td>1.875° × 0.84° L31</td>
<td></td>
</tr>
<tr>
<td>7 GFDL CM2.0</td>
<td>United States</td>
<td>2.5° × 2° L24</td>
<td>1° × 1° (1/3°) L50</td>
<td></td>
</tr>
<tr>
<td>8 GFDL CM2.1</td>
<td>United States</td>
<td>2.5° × 2° L24</td>
<td>1° × 1° (1/3°) L50</td>
<td></td>
</tr>
<tr>
<td>9 GISS AOM</td>
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<td>4° × 3° L12</td>
<td>4° × 3° L16</td>
<td></td>
</tr>
<tr>
<td>10 GISS EH</td>
<td>United States</td>
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</tr>
<tr>
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</tr>
<tr>
<td>12 IAP FGOALS</td>
<td>China</td>
<td>T42 L26</td>
<td>1° × 1° L33</td>
<td></td>
</tr>
<tr>
<td>13 INGV SXG</td>
<td>Italy</td>
<td>T106 L19</td>
<td>2° × 2° (1°) L31</td>
<td>$U_{sfc}$, $V_{sfc}$</td>
</tr>
<tr>
<td>14 INM CM3.0</td>
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<td>5° × 4° L21</td>
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</tr>
<tr>
<td>15 IPSL CM4</td>
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<td>16 MIROC3.1 Hi</td>
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<td>17 MIROC3.1 Med</td>
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</tr>
<tr>
<td>18 MIUB ECHO-G</td>
<td>Germany/Korea</td>
<td>T30 L19</td>
<td>2.8° × 2.8° L20</td>
<td>$T_a$, $U$, $V$, $q$</td>
</tr>
<tr>
<td>19 MPI ECHAM5</td>
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<td>1.5° × 1.5° L40</td>
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</tr>
<tr>
<td>20 MRI CGCM2.3</td>
<td>Japan</td>
<td>T42 L30</td>
<td>2.5° × 0.5° L23</td>
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</tr>
<tr>
<td>21 UKMO HadCM3</td>
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<td>3.75° × 2.5° L19</td>
<td>1.25° × 1.25° L30</td>
<td></td>
</tr>
<tr>
<td>22 UKMO HadGem1</td>
<td>United Kingdom</td>
<td>1.875° × 1.25° L38</td>
<td>1° × 1° (1/3°) L40</td>
<td></td>
</tr>
</tbody>
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Table 1.2. The CMIP5 models and scenarios adopted in this study. All changes are scaled by tropical mean (20°S-20°N) SST changes for the specific models.

<table>
<thead>
<tr>
<th>Model name</th>
<th>Modeling center</th>
<th>Country</th>
<th>Scenarios</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 ACCESS1.0</td>
<td>CSIRO-BOM</td>
<td>Australia</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>2 BCC-CSM1.1</td>
<td>BCC</td>
<td>China</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>3 CanESM2/AM4*</td>
<td>CCCMA</td>
<td>Canada</td>
<td>RCP4.5, 1pctCO2, AMIP, AMIP4K, AMIP4xCO2</td>
</tr>
<tr>
<td>4 CCSM4</td>
<td>NCAR</td>
<td>United States</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>5 CNRM-CM5*</td>
<td>CNRM-CERFACS</td>
<td>France</td>
<td>RCP4.5, 1pctCO2, AMIP, AMIP4K, AMIP4xCO2</td>
</tr>
<tr>
<td>6 CSIRO-Mk3.6.0#</td>
<td>CSIRO-QCCCE</td>
<td>Australia</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>7 FGOALS-g2+</td>
<td>LASG-CESS</td>
<td>China</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>8 GFDL-CM3</td>
<td>NOAA GFDL</td>
<td>United States</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>9 GFDL-ESM2G</td>
<td>NOAA GFDL</td>
<td>United States</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>10 GFDL-ESM2M#</td>
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<td>United States</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>11 GISS-E2-R</td>
<td>NASA GISS</td>
<td>United States</td>
<td>RCP4.5</td>
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<td>12 HadGEM2-CC</td>
<td>MOHC</td>
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<td>RCP4.5</td>
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<td>13 HadGEM2-ES/-A*</td>
<td>MOHC</td>
<td>United Kingdom</td>
<td>RCP4.5, 1pctCO2, AMIP, AMIP4K, AMIP4xCO2</td>
</tr>
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<td>14 INM-CM4</td>
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<td>Russia</td>
<td>RCP4.5</td>
</tr>
<tr>
<td>15 IPSL-CM5A-LR*</td>
<td>IPSL</td>
<td>France</td>
<td>RCP4.5, 1pctCO2, AMIP, AMIP4K, AMIP4xCO2</td>
</tr>
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<td>RCP4.5</td>
</tr>
</tbody>
</table>

*Models available for CMIP5-CFMIP2 analysis on atmospheric overturning circulation.

#Outliers for TPZI.

+Outlier for TEPI.
Table 1.3. Annual mean spatial correlation coefficient \( (r, 40^\circ S-40^\circ N) \) of various variables in GFDL CM2.1 simulation under SUSI and A1B scenarios.

<table>
<thead>
<tr>
<th>( T_{va}^* ) SUSI</th>
<th>( T_{va}^* ) A1B</th>
<th>( U_{sh}^* ) SUSI</th>
<th>( U_{sh}^* ) A1B</th>
<th>( T_{zm}^* ) SUSI</th>
<th>( T_{zm}^* ) A1B</th>
<th>( \omega_{va}, T_{va}^* ) SUSI</th>
</tr>
</thead>
<tbody>
<tr>
<td>( r )</td>
<td>0.59</td>
<td>0.57</td>
<td>0.91</td>
<td>0.38</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\( T^* \) is atmospheric warming patterns, \( U' \) is the change of zonal wind, and \( \omega \) is climatological pressure velocity. Subscripts \( va \) denotes vertical (300-850 hPa) average, \( zm \) for zonal mean, and \( sh \) for 300-850 hPa wind shear.
Fig. 1.1. Comparison of annual mean rainfall changes (color shading, in mm month$^{-1}$) between (a) GFDL CM2.0 and (b) HadCM3 in the CMIP3 ensemble under the SRES A1B emission scenario. (c) Their difference along SST difference in contours [contour interval (CI): 0.2 K; the zero contour omitted].
Fig. 1.2. Annual mean 300-850 hPa averaged climatological mean pressure velocity [contour interval (CI) 0.02 Pa s$^{-1}$; zero omitted], air temperature warming deviations (color shading, K) from the tropical (40°S-40°N) mean, and 300-850 hPa wind shear change (vectors, m s$^{-1}$) simulated with GFDL AM/CM2.1 under (a) SUSI and (b) SRES A1B scenarios, respectively.
Fig. 1.3. Latitude-height section of annual and zonal mean tropospheric air temperature change (color shading, K), and climatological meridional stream function (black contours, CI 2×10^{10} kg s^{-1}; zero omitted) simulated with GFDL AM/CM2.1 under (a) SUSI and (b) SRES A1B scenarios, respectively.
Chapter 2

Data and methods

2.1 CMIP3 models

This study analyzes CMIP3 model simulations (Table 1.1) forced with the IPCC SRES A1B scenario representing the emission of a few climatically important trace gases (e.g., carbon dioxide and ozone). Based on certain socioeconomic development paths for the twenty-first century, this scenario projects a rough doubling of atmospheric CO₂ for the century as well as a recovery of the Southern Hemisphere “ozone hole” by approximately 2050. Details of the models can be found at www.pcmdi.llnl.gov/ipcc/model_documentation/ipcc_model_documentation.php, and the output at https://esg.llnl.gov:8443/index.jsp. A total of 22 models are included with one realization for each model. Monthly output is used. When monthly-means are unavailable, the data are either computed from daily output or converted from other variables. See details in Table 1.1.

To extract robust anthropogenic global warming signals, changes are computed for the twenty-first century between two 10-year periods: 2001-10 and 2091-2100. Then, they are normalized by the tropical (20°S-20°N) mean SST warming in each model before calculating the ensemble averages and the deviations from the ensemble mean.
The SUSI experiments advocated by the Cloud Feedback Model Intercomparison Project (CFMIP) (Ringer et al. 2006) are used for zonal mean comparisons with the A1B simulations (only GFDL AM2.1, MPI ECHAM5, and NCAR CAM3.1 available).

2.2 CMIP5 data

The CMIP5 output under the representative concentration pathway 4.5 (RCP4.5) is available for 22 models (Table 1.2). RCP4.5 is a scenario stabilizing radiative forcing at 4.5 W m\(^{-2}\) in 2100 without ever overshooting by employing technologies and strategies for reducing GHG emissions (Thomson et al. 2011). It includes long-term, global emissions of GHG, aerosols, and land-use-land-cover. Anthropogenic aerosol forcing peaks at the beginning of the 21\(^{st}\) century at -1.6 W m\(^{-2}\) and reduces to -0.5 W m\(^{-2}\) by 2100 for the sum of direct and first indirect effects, concentrating in the Northern Hemisphere (Bellouin et al. 2011). The changes here are calculated between 2006-15 and 2089-98. The preliminary analyses of RCP4.5 data generally support the CMIP3 A1B results.

The CFMIP2 suite of the CMIP5 simulations is used to isolate the mechanisms for changes of the overturning circulations:

- Coupled models: CO\(_2\) concentration increases at 1 percent per year until quadrupling (~140 years);
- RAD (CFMIP2): Quadrupling CO\(_2\) concentration while holding SST at the current climatology;
- SUSI (CFMIP2): SST is spatial-uniformly warmed by 4 K (Cess et al. 1990);
- SST: Effect of SST warming patterns is calculated as residual [Coupled models – (RAD+SUSI)].
Here all results are normalized by their tropical (20°S-20°N) mean SST increases. Currently, five models are available (Table 1.2).

2.3 GFDL CM2.1 diagnostics

To simulate features of tropospheric circulation change, an LBM is adopted with forcing terms diagnosed from global warming simulations by NOAA GFDL models under SUSI and A1B scenarios. The CM2.1 uses the Flexible Modeling System to couple the GFDL AM2.1 with the Modular Ocean Model version 4. The AM2.1 builds on a finite volume atmospheric dynamical core and includes atmospheric physical packages and a land surface model. Its resolution is 2° latitude x 2.5° longitude with 24 vertical levels, nine of which are located in the lowest 1.5 km to represent the planetary boundary layer. The ocean model uses a finite difference approach to solve the primitive equations. The resolution is 1° longitude by 1° latitude, with meridional grid spacing decreasing to 1/3° toward the equator. The model has 50 vertical levels, 22 of which are in the upper 220 m. A detailed description of CM2.1 can be found in GFDL Global Atmospheric Model Development Team (2004) and Delworth et al. (2006). Long integrations (~2000 years) have been performed under current climate forcing without flux correction, reaching statistically steady states similar to observations, including the annual-mean state, seasonal cycle and major modes of interannual variability (Wittenberg et al. 2006).

The SUSI experiment was performed with AM2.1 for the period of 1983-1991, by adding a uniform SST increase of 2 K. Another set of doubling CO₂ experiments by AM2.1 during 1981-2000 is also used to isolate the atmospheric response to radiative forcing (noted as RAD). Both of the SUSI and RAD experiments employ interannual-variable monthly mean observed SST.
For the SRES A1B, a 10-member ensemble simulation has been completed at GFDL with CM2.1 from 1996 to 2050, during which CO\textsubscript{2} concentration increases from 369 to 532 ppm. This study analyzes ensemble-mean, 50-year difference fields: 2046-50 minus 1996-2000. The use of ensemble means helps reduce natural variability and isolate the response to the greenhouse gas (GHG) increase. The annual-mean SST rise averaged in the tropics (20°S-20°N) is 1.12 K in CM2.1.

Changes for SUSI and A1B are normalized by the tropical mean SST warming (20°S-20°N). The RAD run is scaled by \( \frac{\delta \ln \text{CO}_2, \text{P100}}{\delta \ln \text{CO}_2, \text{P50}} = 1.91 \), since CO\textsubscript{2} radiative forcing is proportional to the logarithm of its concentration, and then by the tropical mean SST warming of CM A1B (1.12 K).

### 2.4 Atmosphere GCM (AGCM) simulations

In order to test the atmospheric response to multiple components of the SST warming, a sensitivity study is performed with the National Center for Atmospheric Research (NCAR) Community Atmosphere Model (CAM), version 3.1. CAM is a global AGCM developed by the climate research communities in collaboration with NCAR (Collins et al. 2006). Integrated with a land model and a thermodynamic sea ice model, it is suitable for examining the response of the atmospheric circulation and rainfall to changes in SST.

The model runs for 20 years with triangular truncation at T42 (equivalent grid spacing of 2.88°) and 26 vertical levels. The CAM experiments are forced with the observed monthly mean SST climatology plus changes (except the control run) derived from the CMIP3 ensemble and annual mean SST warming, which is decomposed into SUSI and patterns. Specifically, they include the following cases.
• CAM_A1B: SST increases as the CMIP3 ensemble mean;
• CAM_SUSI: SST is spatial-uniformly warmed by 2 K;
• CAM_T*: Only spatial patterns of SST change (T*) are applied, defined as the deviations of the CMIP3 warming from the tropical (20°S-20°N) mean, equivalent to CAM_A1B minus CAM_SUSI;
• CAM_NEPI: The equatorial peak of T* (Fig. 3.1) is eliminated by applying a Gaussian weight in the meridional direction;
• CAM_EP: Calculated as CAM_T* minus CAM_NEPI.

Again, results are normalized by their own tropical mean SST warming accordingly before post-calculations.

2.5 LBM

This study adopts an LBM to study mechanisms for tropospheric circulation change. It is the dry version of a global, time-dependent, primitive equation atmospheric model based on a set of linearized equations for vorticity, divergence, temperature, and the logarithm of surface pressure (Watanabe and Kimoto 2000, 2001; Watanabe and Jin 2004). The model variables are expressed horizontally in the spherical harmonics at T42 while finite difference is used for the vertical discretization with 20 σ-levels. The model includes biharmonic horizontal diffusion with an e-folding time of 3 hours for the highest wave number. It also employs Rayleigh friction and Newtonian cooling, whose e-folding time scales are set to be 20 days in most of the free troposphere, but 0.5 and 1 day for the three lowest (σ > 0.9) and two upper-most (σ < 0.03) levels, respectively.
LBM is widely used to study atmospheric variability, but its utility for global warming research has not been investigated. Here the LBM is adapted for the latter purpose by a reformulation that accounts for the effect of global increase in static stability (Fig. 1.3).

2.6 Moisture budget analysis

A moisture budget analysis is performed to decompose the atmospheric dynamic and thermodynamic contributions to rainfall change over ocean (Seager et al. 2010). Once the atmospheric moisture equation is vertically integrated, one obtains

\[ P - E = -\left( \nabla \cdot \left( \bar{\nabla} \bar{q} \right) \right) + Eddy, \quad (2.1) \]

where \( P \) is precipitation, \( E \) is evaporation, \(< >\) represents column mass integration throughout the troposphere (approximated as 200-1000 hPa), and the over-bar denotes the monthly average. \( \mathbf{V} \) denotes three-dimensional atmospheric velocity, but here two-dimensional fields are used to include more models, by assuming that pressure velocity can be neglected at the tropopause and ocean surface. The eddy term is due to sub-monthly variability and calculated as residual.

In global warming, the perturbation of \( P - E \) can be linearly decomposed as

\[ \delta (P - E) = -\left( \nabla \cdot \left( \delta \bar{\nabla} \bar{q} \right) \right) - \left( \nabla \cdot \left( \bar{\nabla} \delta q \right) \right) + \delta Eddy, \quad (2.2) \]

where the first term on the right-hand-side represents the contribution of circulation change (dynamic effect), and the second term moisture content change (thermodynamic effect).

2.7 Statistical methods
Empirical orthogonal function (EOF) and singular value decomposition (SVD) analyses are applied to the CMIP3 and CMIP5 ensembles to investigate the inter-model variations in SST change patterns and its contributions to changes in other variables.
Chapter 3

Regional patterns of SST change and uncertainty in future rainfall projection

This chapter investigates the effects of SST warming patterns on changes in tropical precipitation. The “warmer-get-wetter” mechanism is examined for the spatial variations in tropical rainfall response to GHG forcing as represented by the CMIP3 multi-model ensemble mean. Then, the SST pattern effect is shown important for the inter-model variations in tropical precipitation change, highlighting the need to study SST warming patterns as an ocean-atmosphere interaction problem. Finally a comparison is made between the CMIP3 A1B and CMIP5 RCP4.5 results.

3.1 CMIP3 ensemble mean change patterns

This section examines tropical rainfall change under global warming and relates it to SST warming patterns. It starts with an analysis of the CMIP3 ensemble mean, followed with a water vapor budget and AGCM experiments.

3.1.1 SST, rainfall, and surface winds
To highlight the effect of spatial variations in SST warming, the CMIP3 models projections under the SRES A1B emission scenario is compared with simulations with their atmospheric components in response to a SUSI of 2 K, the latter available through the CFMIP (Ringer et al. 2006). Figure 3.1 presents the zonal mean rainfall changes over ocean in these model ensembles, with climatological precipitation and SST change for reference. Rainfall change in SUSI runs (Fig. 3.1a) resembles the climatology (Fig. 3.1b). They share an equatorial minimum sandwiched by double peaks on either side, with \( r = 0.67 \) in 20°S-20°N. A maximum of inter-model variations anchoring the Northern Hemispheric peak appears consistently in both fields. This relationship in SUSI is consistent with the “wet-get-wetter” mechanism (Xie et al. 2010).

The SST change develops patterns in space, here measured by \( T^* \), the deviations of SST warming from its tropical (20°S-20°N) mean increase. In zonal mean (Fig. 3.1b), major features of these patterns include an equatorial peak (Liu et al. 2005) and south-to-north gradients (Xie et al. 2010). The mean rainfall change of the A1B ensemble (Fig. 3.1a) shows little correlation with SUSI \( (r = 0.18) \). Instead of an equatorial minimum in SUSI, A1B precipitation features a broad equatorial increase with large inter-model spread, apparently forced by the equatorial maximum in \( T^* \), which also shows considerable spread. The subtropical reduction in A1B precipitation seems to fit the “dry-get-drier” pattern but is actually associated with reduced SST warming \( (T^* < 0) \) especially in the Southern Hemisphere. In A1B simulations, the ensemble mean precipitation change and relative SST warming are highly correlated at \( r = 0.80 \). This illustrates the dominance of the “warmer-get-wetter” mechanism in the coupled models.
Figures 3.2a and b compare percentage precipitation change with relative SST warming in the 22 CMIP3 models under A1B scenario (Table 1.1). A clear correlation ($r = 0.68$) in space emerges in the ensemble mean, with increasing $\delta P/P$ generally collocated with positive $T^*$, and vice versa. In particular, the equatorial maximum in $T^*$ anchors a large precipitation increase in the equatorial Pacific, while precipitation generally decreases in the subtropical Southern Hemisphere where SST warming is subdued ($T^* < 0$). The reduced SST warming is associated with the intensified southeasterly trade winds (Fig. 3.2c), suggestive of wind-evaporation-SST (WES) feedback (Xie and Philander 1994). Reduced warming and suppressed rainfall are also found over the subtropical North Atlantic, a result of enhanced evaporative damping rate (Leloup and Clement 2009) and ocean circulation change (Xie et al. 2010).

SST change patterns are robust for the equatorial peak and Southern Hemispheric minima (Fig. 3.2a). The robustness of rainfall change there (Fig. 3.2b) is an SST effect. Moderate uncertainty in rainfall change for the central equatorial Pacific may be due to differences in model physics/coupling scheme (e.g. the intensity of the climatological equatorial cold tongue).

Strong spatial correlation between $\delta P/P$ and $T^*$ in the A1B ensemble mean suggests an empirical relation (Fig. 3.3)

$$\delta P/P = \alpha T^* + \beta \bar{T}, \quad (3.1)$$

where $\alpha = 44\% \ K^{-1}$, $\beta = 2\% \ K^{-1}$, and $\bar{T} = 1 \ K^{-1}$ (the tropical mean warming normalized by itself). In SUSI, $T^* = 0$ and $\delta P$ is proportional to $P$, representing the “wet-get-wetter” mechanism. $\beta$ measures the percentage increase in the tropical average rainfall due to SUSI and direct radiative effects. In A1B, $T^*$ is only a fraction of the tropical mean SST warming (Table 3.1), but its effect on rainfall change [the first term on the right hand side of Eq. (3.1)] is an order of magnitude greater than the
second term (Fig. 3.3). In Table 3.1, a common rule stands out for both ocean and land. The standard deviation of 2-m air temperature warming is only a fraction of its global mean, whereas the spatial variability in rainfall change is four times larger than the mean. The mean land warming is 1.5 times of the ocean warming, but the spatial variability is similar in magnitude. For precipitation, the mean and variability are both smaller over land than over ocean.

The Clausius-Clapeyron equation predicts that the atmospheric moisture content increases at a rate of \( \alpha_0 = 7\% \text{ K}^{-1} \) (Held and Soden 2006). The fact that \( \alpha >> \alpha_0 \) indicates that circulation change is important for regional precipitation change. Figure 3.2c shows that the SST patterns dominate the sea surface wind change and moisture convergence. Indeed, convergence is generally found where precipitation increases and \( T^* > 0 \), indicative of the strong positive feedback between circulation and convection that is commonly seen in the tropics. Because of this interaction, \( \alpha \) is much larger than \( \beta \), which is determined by global mean water vapor content increase at \( \alpha_0 \) deducted by reduction of convective mass flux.

For individual models (Fig. 3.4), \( \alpha \) varies in the range of 10-70\% K\(^{-1}\) with a right-skewed distribution, and \( \beta \) in the range of -1-5\% K\(^{-1}\). Not surprisingly, models with large \( \alpha \) feature a high correlation \( (r) \) between \( \delta P/P \) and \( T^* \) (Fig. 3.4d). The inter-model correlation is 0.62 between \( r \) and \( \alpha \).

### 3.1.2 Moisture budget analysis

A moisture budget analysis (Eq. 2.2) helps identify whether the SST pattern control on regional precipitation is through spatial variations in water vapor increase or associated with atmospheric circulation change by quantifying the relative importance of the atmospheric dynamic and thermodynamic contributions to
δ\(\overline{P-E}\) (Seager et al. 2010; Chou et al. 2012). Figure 3.5 illustrates the CMIP3 ensemble-mean results over ocean. The \(\overline{P-E}\) change (Fig. 3.5a) is well correlated in space with the contribution by circulation change (Fig. 3.5b), with \(r = 0.73 \pm 0.10\) in the multi-model ensemble. Especially, the rainfall enhancement in the equatorial Pacific and reduction in the southeastern Pacific are due to circulation change induced by SST patterns. While the moisture increase (Fig. 3.5c) produces the “wet-get-wetter” pattern, its correlation with \(P - E\) is quite low \((r = 0.30 \pm 0.17)\). This confirms that over ocean, although \(T^*\) is only a fraction of the tropical mean SST warming (Table 3.1), near surface atmospheric circulation changes induced by SST patterns dominates regional precipitation response to global warming. The eddy contribution (Fig. 3.5d) shows a clear poleward expansion of the Hadley cell in the Pacific and Atlantic Oceans.

### 3.1.3 AGCM experiments

AGCM experiments are performed to test how different components of SST warming, including the SUSI and spatial patterns, influence the atmospheric circulation and regional rainfall. Figure 3.6 evaluates the ability of CAM3.1 to simulate the ensemble mean change in the CMIP3 models. With \(r = 0.62\) in 20°S-20°N, the CAM_A1B experiment can reproduce the regional precipitation change in Figure 3.2b quite well, including the strong equatorial enhancement and the reduction in the southeastern Pacific, subtropical Atlantic and Indian Oceans. Surface wind change is also well simulated, with the enhanced southeasterly trades in the southeastern Pacific and weakening of the Walker circulation.
The CAM_SUSI experiment (Fig. 3.6b) shows that the tropical mean SST warming contributes to the rainfall reduction in the northeastern Pacific and the Mediterranean Sea. Besides, the SUSI causes cyclonic circulation in major subtropical ocean basins, which corresponds to the slow-down of surface winds. Specifically, this is consistent with the weakening of the Walker circulation.

Figures 3.6c and d compare the effect of $T^*$ evaluated with different methods as the difference between the CAM_A1B and CAM_SUSI runs, and the atmospheric response to $T^*$. Basically, the rainfall and surface wind change patterns are very similar between the two methods. In fact, the major features in Figure 3.6a are largely reproduced by both methods, illustrating the importance of SST patterns in reorganizing regional precipitation in a changing climate.

Without the equatorial peak in $T^*$ (CAM_NEP), it becomes clear that the south-to-north gradient of SST warming (Fig. 3.6e) is associated with a basin-scale WES feedback in the Pacific and Atlantic (Xie and Philander 1994), with enhanced/reduced trades collocating with weaker/stronger SST warming in the southeastern/northeastern Pacific. Note that (Figs. 3.6d and e) the high-pressure center is displaced southwest of the SST minimum in the southeastern Pacific, a feature that needs further investigation. The similarity between Figures 3.6c and d suggests that the AGCM experiments are linearly additive. Thus, the CAM_NEP experiment (Fig. 3.6e) is taken to separate the SST patterns into two modes: The equatorial peak and inter-hemispheric asymmetry. Calculated as CAM$_T$ minus CAM_NEP, the equatorial peak effect (Fig. 3.6f) is accompanied by meridional surface wind convergence, associated with rainfall increase on the equator and reduction on the sides. It also contributes to the reduction of the Walker circulation.
The above analysis shows that the CMIP3 ensemble-mean SST warming patterns are composed of two leading modes: the equatorial peak and south-to-north gradient. SST patterns interact with the atmospheric circulation and dominate rainfall reorganization.

3.2 Inter-model variations in CMIP3

precipitation change

SST and precipitation changes vary considerably among models, and this section shows that their inter-model variations are correlated over ocean. An inter-model EOF analysis is performed on the SST changes among the CMIP3 models in the tropics (20°S-20°N). Fig. 3.7 shows the leading modes with regressions for multiple variables. The first mode represents inter-model variability in cross-equatorial SST gradient, with large SST anomalies in the subtropics. $T^*$ (Fig. 3.7a), air temperature (Fig. 3.7c), and $\delta P/P$ (consistent with the low-level moisture convergence) (Fig. 3.7e) are all asymmetric between the hemispheres, with a warmer and wetter Northern Hemisphere. The surface wind (Fig. 3.7a) and vertical wind shear (Fig. 3.7c) show consistent baroclinic patterns, suggestive of a basin-scale WES feedback, with enhanced/reduced trades in the Southern/Northern Hemisphere.

The second modes are more symmetric about the equator (Fig. 3.7b, d, f), with enhanced rainfall collocated with positive SST anomalies in the equatorial Pacific. The equatorial peak warming is associated with the slow-down of the Pacific Walker circulation, which can be seen for both surface wind (Fig. 3.7b) and vertical wind shear (Fig. 3.7d). It is noteworthy that the equatorial mode of inter-model variability peaks in the western Pacific. In Xie et al. (2010), anomalous warm oceanic advection
due to the weakened south equatorial current is important in this region. Indeed, the inter-model variability in SST warming over the central equatorial Pacific is associated with the net surface heat flux that damps the SST signal (not shown), indicative of an ocean dynamic origin. Thus, the coherence between surface wind and SST patterns in the western equatorial Pacific may indicate the model dependency of an air-sea interaction process there.

Fig. 3.8 shows the principle components (PCs) of the leading EOF modes for each tropical ocean basin. The phase of PCs represents the intensity of each mode in individual models. The inter-model variations in the Pacific and Atlantic are generally coherent in phase, with correlation coefficients of 0.52 and 0.32 for PC 1 and PC 2, respectively. The Indian Ocean, somehow, shows opposing phase to the Pacific/Atlantic in a number of models (especially for PC 2), which may be related to the Indian Ocean dipole (IOD) feature and needs further investigation.

Reduced warming in the Southern Hemisphere subtropics and the equatorial-enhanced warming are dominant patterns of SST response to global warming (Figs. 3.2a and b). The EOF analysis above shows that models display considerable differences in representing the magnitude of these patterns. Remarkably, the leading two EOF modes for SST explain about one third of the inter-model spread in precipitation projection (Table 3.2). The EOF analysis has been repeated for zonal mean SST, yielding the inter-hemispheric and equatorial patterns as the leading modes (Fig. 3.9). The SST modes explain 36% of the inter-model variability in zonal-mean precipitation (Table 3.2). The strong SST regulation of variability in rainfall change among models indicates that SST patterns are an important source of uncertainty for regional rainfall projection.
3.3 Comparison with CMIP5

This section compares the regional patterns of SST and rainfall projections between the CMIP3 A1B and CMIP5 RCP4.5 datasets. Several global warming feature indices are devised to characterize major patterns of SST and $\delta P/P$ change as seen in CMIP3 results. They include:

- SST/$\delta P/P$ meridional gradient index (TMGI/PMGI) =
  \[\text{Variables averaged in 10°-20°N minus those in 10°-20°S};\]

- SST/$\delta P/P$ equatorial peak index (TEPI/PEPI) =
  \[\text{Variables averaged in 5°S-5°N minus those in 15°-25°N and 15°-25°S};\]

- SST/$\delta P/P$ Pacific zonal index (TPZI/PPZI) =
  \[\text{Variables averaged in 5°S-5°N, 80°-120°W minus those in 5°S-5°N, 140°E-180°};\]

- SST/$\delta P/P$ Indian Ocean zonal index (TIZI/PIZI) =
  \[\text{Variables averaged in 5°S-5°N, 45°-65°E minus those in 5°S-5°N, 80°-100°E}].\]

Figure 3.10 shows the SST indices. Apparently, three outlier models stand out from the CMIP5 ensemble, with two extremes for TPZI and one lower extreme for TEPI. These outliers significantly influence the statistical characteristics, especially for the inter-model variations. Thus, this section introduces results with and without the outliers, but only the latter is focused on with details shown in tables and figures.

For most indices, the inter-model variance is larger in CMIP5 than in CMIP3 (Fig. 3.10). Consistent with Figs. 3.1b and 3.2a, the ensemble-mean TMGI is positive in CMIP3. In CMIP5, it is twice as large with enhanced robustness, indicating stronger south-to-north warming gradient. With outliers removed, the TEPI has similar ensemble means between two datasets with higher spread in CMIP5. If the outliers are not removed, the CMIP5 TEPI has a lower ensemble mean and much larger spread,
which can make the equatorial peak mode dominate in the inter-model EOFs (discussed later). The TPZI characterizes the El Nino-like feature among the models. Without the outliers, CMIP5 shows much more significant and robust El Nino patterns than CMIP3, though inter-model variations are much larger in CMIP5. With the outliers, the inter-model spread is even more in CMIP5, making the zonal gradient one of the leading EOF modes. The IOD-like feature (TIZI), by contrast, is lower in CMIP5 than in CMIP3.

Figure 3.11 shows the ensemble-mean SST and rainfall change patterns of the 19 CMIP5 RCP4.5 GCMs without the outliers (Table 1.2). As the spatial variance raises in both SST and rainfall patterns in comparison to CMIP3 (Fig. 3.2), the major patterns remain similar: meridional gradient and equatorial peak. As pointed out by the TMGI, the south-to-north gradient of SST warming is enhanced in CMIP5 because of a more warmed Northern Hemisphere than CMIP3. The equatorial peak of CMIP5 becomes more El Nino-like, with SST warming peak in the eastern equatorial Pacific and rainfall peak in the mid-equatorial Pacific. This feature is robust as shown by the robustness and TPZI. The “warmer-get-wetter” view is quite applicable in the CMIP5 models, with tropical correlation of SST and rainfall patterns at $r = 0.69$.

As mentioned above, the outlier models significantly influence the inter-model variations. With all 22 CMIP5 RCP4.5 GCMs (Table 1.2), the first EOF mode represents inter-model variability in the equatorial peak (see the TEPI in Fig. 3.10 for a measure of variance). The second mode in the Pacific is a zonal mode, representing the variance caused by the outliers in the TPZI (Fig. 3.10). Without the outliers, RCP4.5 results (Figs. 3.12 and 3.13) are qualitatively consistent with CMIP3 (Figs. 3.7 and 3.9). Specifically, the dominant modes remain similar: meridional gradient as the first mode and the equatorial peak the second. Note that the equatorial peak mode
is more ENSO-like in CMIP5 than in CMIP3, with SST peak in the eastern equatorial Pacific and rainfall peak in the mid-equatorial Pacific (Figs. 3.12b and d).

Coherent rainfall changes are associated with SST modes of inter-model variability (Figs. 3.12 and 3.13). To represent this coherence, rainfall feature indices are calculated with the same manner as SST for each model. High inter-model correlations of the SST and rainfall indices are observed in CMIP5, on a similar order of magnitude with CMIP3 (Table 3.3). The first two modes of SST inter-model variability can explain rainfall variation by one fourth for each ocean basin (a bit lower than in CMIP3) and 39% for zonal mean (Table 3.2), indicating the importance of SST patterns to uncertainty in rainfall projection.

3.4 Summary

In this chapter, relationships among SST, precipitation, and atmospheric circulation changes in response to global warming are examined by using a large ensemble of CMIP simulations. Spatial patterns of SST warming are found to play a key role in determining regional precipitation change. In the ensemble mean, the annual mean rainfall change over tropical oceans follows a “warmer-get-wetter” pattern. The moisture budget analysis shows that this SST control is not simply a result of spatial variations in water vapor increase (the Clausius-Clapeyron relation), but through adjustments in atmospheric circulation. The “warmer-get-wetter” pattern dominates in coupled models and deviates from the “wet-get-wetter” pattern realized in atmospheric response to uniform SST increase.

Both the ensemble mean and inter-model variability feature two major patterns of SST change: the equatorial peak and cross-equatorial gradient. The equatorial peak drives low-level moisture convergence and enhances local convection/precipitation.
This pattern is more El Nino-like in CMIP5 than in CMIP3. The south-to-north gradient pattern is associated with inter-hemispheric WES feedback, with enhanced/reduced trades and drying/wetting in the Southern/Northern Hemisphere. These patterns are robust in the CMIP ensemble mean but their magnitude varies among models. The diversity in representing these two modes among models is an important source of uncertainty for rainfall projection over tropical oceans.
Table 3.1. Ensemble-means of spatial mean ($M_{x,y}$) and variability ($\sigma_{x,y}$) of changes in air temperature at 2 m and precipitation in the 22 CMIP3 models. Changes are defined as the annual mean of 2091-2100 minus that of 2001-2010, normalized by the tropical mean SST warming. The calculations are limited to nearly ice-free regions ($60^\circ$S-$60^\circ$N).

<table>
<thead>
<tr>
<th></th>
<th>Air temperature at 2 m (K)</th>
<th>Precipitation (mm month$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Global</td>
<td>Ocean</td>
</tr>
<tr>
<td>$M_{x,y}$</td>
<td>1.16</td>
<td>0.98</td>
</tr>
<tr>
<td>$\sigma_{x,y}$</td>
<td>0.45</td>
<td>0.34</td>
</tr>
</tbody>
</table>
Table 3.2. Inter-model variance explained by the two leading EOF modes of SST variability (20°S-20°N).

<table>
<thead>
<tr>
<th></th>
<th>%</th>
<th>Pacific</th>
<th>Atlantic</th>
<th>Indian Ocean</th>
<th>Zonal mean</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>CMIP3</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T^*$</td>
<td>54</td>
<td>69</td>
<td>60</td>
<td>84</td>
<td></td>
</tr>
<tr>
<td>$\delta P/P$</td>
<td>33</td>
<td>37</td>
<td>27</td>
<td>36</td>
<td></td>
</tr>
<tr>
<td><strong>CMIP5</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T^*$</td>
<td>47</td>
<td>56</td>
<td>56</td>
<td>90</td>
<td></td>
</tr>
<tr>
<td>$\delta P/P$</td>
<td>23</td>
<td>26</td>
<td>23</td>
<td>39</td>
<td></td>
</tr>
</tbody>
</table>
Table 3.3. Inter-model correlation of SST and rainfall feature indices.

<table>
<thead>
<tr>
<th></th>
<th>r</th>
<th>TMGI/PMGI</th>
<th>TEPI/PEPI</th>
<th>TPZI/PPZI</th>
<th>TIZI/PIZI</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMIP3</td>
<td>0.91</td>
<td>0.36</td>
<td>0.62</td>
<td>0.68</td>
<td></td>
</tr>
<tr>
<td>CMIP5</td>
<td>0.82</td>
<td>0.67</td>
<td>0.68</td>
<td>0.77</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 3.1. Comparison of annual and zonal mean oceanic rainfall changes between A1B and SUSI simulations, in relation to the climatological precipitation and relative SST warming. The ensemble means are shown in (a) for A1B (solid) and SUSI (dashed) rainfall changes ($\delta P$, in mm day$^{-1}$) normalized by tropical ($20^\circ$S-$20^\circ$N)-mean SST warming, and in (b) for normalized A1B SST warming patterns ($T^*$, in K, solid) and rainfall climatology ($P$, in 20 mm day$^{-1}$, dashed), with inter-model spreads (ensemble mean $\pm$ 1 standard deviation) marked by the shaded ranges. The model ensemble includes GFDL CM2.1, MPI ECHAM5, and NCAR CCSM3.
Fig. 3.2. Relationship between annual mean rainfall and SST change patterns projected by the 22 CMIP3 models under the SRES A1B emission scenario. The ensemble means (color shading) of (a) relative SST warming ($T^*$ in K) and (b) percentage rainfall change ($\delta P/P$ in %), along with robustness defined as the ratio of the ensemble mean (absolute value) to inter-model spread (values > 0.75 mapped with grid). (c) The ensemble-mean change in surface wind (vectors in m s$^{-1}$) and divergence (color shading in $10^{-7}$ s$^{-1}$).
Fig. 3.3. Scatter plot between the percentage change of tropical (20°S-20°N) rainfall and relative SST warming in the ensemble mean of CMIP3 models under A1B scenario. Also marked are the spatial correlation ($r$), standard deviation ($\sigma$) of rainfall changes, growth rate ($\alpha$) and intercept ($\beta$) of the linear fit.
Fig. 3.4. Histogram of (a) $\alpha$, (b) $\beta$, and (c) $r$ for individual models. Dashed lines mark the ensemble mean values. (d) Scatterplot between $r$ and $\alpha$. $\alpha$ and $\beta$ are defined in Eq. (3.1), and $r$ denotes correlation between $\delta P/P$ and $T^*$. 
Fig. 3.5. Annual-mean moisture budget terms (Eq. 2.2, in mm month$^{-1}$) in CMIP3 ensemble mean. The vertical integration is performed in the troposphere (200 - 1000 hPa). The eddy term is calculated as the residual.
Fig. 3.6. Percentage rainfall change (δP/P, shading, in %) and surface winds (vectors, in m s⁻¹) simulated by the AGCM experiments with the NCAR CAM3.1. SST forcing for each experiment is shown in contours (CI: 0.1 K and 0.05 K adjacent to 0; the zero contour omitted). (a) The total response forced by CMIP3 A1B ensemble mean SST change is illustrated with the component SST effects including (b) SUSI, (c) result of (a) minus (b) to be compared with (d) relative SST warming, and SST patterns (e) without the equatorial peak and (f) with the equatorial peak only.
Fig. 3.7. Leading EOF modes of inter-model SST variability [color shading in (a), (b)] in CMIP3 A1B projections, normalized by tropical mean SST warming. The SST EOF analysis is done within each ocean basin and the explained variance for each mode is marked on a neighboring continent. Regressions on these modes are conducted for [(a), (b)] surface winds (vectors); [(c), (d)] tropospheric (300-850 hPa) temperature (color shading) and vertical wind shear (vectors); [(e), (f)] $\delta P/P$ (color shading; variance explained by each SST mode marked for each basin) and 700-1000 hPa moisture divergence (contours).
Fig. 3.8. PCs of the leading EOF modes for each tropical ocean basin.
Fig. 3.9. Inter-model EOF modes of zonal-mean SST changes and regression of zonal-mean $\delta P/P$ in CMIP3 A1B ensemble.
Fig. 3.10. Global warming feature indices devised for major patterns of SST change in CMIP3 A1B and CMIP5 RCP4.5 ensembles. Purple cross marks the outlier models. Circle shows the ensemble mean and error bar means ±1 standard deviation. Statistical variables are calculated after removal of the outliers.
Fig. 3.11. Annual mean rainfall and SST change patterns projected by the 19 CMIP5 models along the RCP4.5. The ensemble means (color shading) of (a) relative SST warming ($T^*$ in K) and (b) percentage rainfall change ($\delta P/P$ in %), along with robustness defined as the ratio of the ensemble mean (absolute value) to inter-model spread (values $> 0.75$ mapped with grid).
Fig. 3.12. Leading EOF modes of inter-model SST variability [(a), (b)] in CMIP5 RCP4.5 projections, normalized by tropical mean SST warming. The SST EOF analysis is done within each ocean basin and the explained variance for each mode is marked on a neighboring continent. Regressions of $\delta P/P$ [(c), (d)] on these modes (variance explained by each SST mode marked for each basin).
Fig. 3.13. Inter-model EOF modes of zonal-mean SST changes and regression of zonal-mean $\delta P/P$ in CMIP5 RCP4.5 ensemble.
Chapter 4

Atmospheric circulation change: A linear model study

The SUSI results in Figures 1.2 and 1.3 illustrate that in the absence of gradients in SST forcing, considerable changes in atmospheric temperature and circulation patterns are excited by global warming, and they are similar to those in the A1B simulation by a coupled model. This chapter investigates mechanisms for these changes and seeks to address the following specific questions: Why does the troposphere warm less over convective than subsidence regions? To what extent do patterns of SST change affect patterns of tropospheric warming and circulation change? What determines the change in tropical overturning circulations and its uncertainty? A diagnostic framework is designed to identify robust dynamical balance and simulate major features of circulation change with an LBM.

Pioneered by Matsuno (1966) and Gill (1980), LBMs forced with SST-induced diabatic heating are very useful to simulate and understand the mean state and variability of tropical circulation. Similarity between SUSI and A1B results (Figs. 1.2 and 1.3) shows something is missing in LBMs and adjustments are necessary for such models to become a useful diagnostic tool for global warming research. The LBM used in this chapter was originally developed by Watanabe and Kimoto (2000, 2001),
and is modified here for global warming studies. It is important to include the effect of increased static stability. With this modification, the LBM is skillful in reproducing salient features of circulation change in both GCM runs under the SUSI and A1B forcing. The LBM proves to be useful in quantifying SST pattern effects.

This chapter first introduces the diagnostic framework utilizing the LBM, describes the forcing factors, and presents the experimental designs. Then, it reports results from the LBM simulations and explores effects of various forcing factors.

### 4.1 Diagnostic framework with LBM

#### 4.1.1 LBM for global warming studies

LBM is a powerful tool to relate circulation change to the geographical distribution of diabatic heating. This section designs a diagnostic framework that employs the LBM with necessary modifications, to identify robust dynamical balance and simulate major features of circulation change. An equation for global warming pattern formation is derived from the approximate thermodynamic equation of the atmospheric temperature change, i.e.,

\[
\begin{align*}
\frac{\partial T'}{\partial t} &+ B \omega \frac{\partial \theta}{\partial p} + B \omega' \frac{\partial \theta'}{\partial p} = LH + SH + Q_R, \\
Q_R &= SW + LW_c + LW_f - LW_R(T')
\end{align*}
\]

where the overbar and prime denote the mean and change terms, respectively. \( T \) is air temperature, \( \theta \) is potential temperature, \( p \) is pressure, \( B = \frac{T}{\theta} = \left( \frac{p}{p_S} \right)^{R/c_p} \), where \( p_S \) is surface pressure, \( R \) is the gas constant for air, and \( c_p \) is the specific heat at constant pressure. As changes in time mean fields are interested in, \( \frac{\partial T'}{\partial t} \) may be omitted. The
diabatic forcing includes latent heating (LH), sensible heating (SH) from vertical diffusion, and radiation (Q_R). Q_R is composed of forcing in short wave (SW), long wave (LW) from cloud (LWC) and clear sky GHG (LWF), and feedback [LWR(T')]. LW_R(T') is largely due to clear sky LW change in direct response to atmospheric warming, and is parameterized as the Newtonian damping (εT') in the LBM.

The global average is denoted as 〈x〉, the spatial patterns as x* = x – 〈x〉 and x* = x' – 〈x'〉 for the mean and change fields, respectively. After the global average, Eq. (4.1) becomes

\[ B \left( \omega' \frac{\partial \tilde{\theta}}{\partial p} \right) + B \left( -\frac{\omega \partial \tilde{\theta}'}{\partial p} \right) = \langle LH \rangle + \langle SH \rangle + \langle Q_r \rangle. \]

Because of the global mass conservation, i.e., 〈ω〉 = 0 and 〈ω'〉 = 0,

\[ \left( \omega' \frac{\partial \tilde{\theta}}{\partial p} \right) = \left( \omega \frac{\partial \tilde{\theta}^*}{\partial p} \right), \text{ and } \left( -\omega \frac{\partial \tilde{\theta}'}{\partial p} \right) = \left( -\omega \frac{\partial \tilde{\theta}^*}{\partial p} \right). \]

Hence,

\[ B \left( \omega' \frac{\partial \tilde{\theta}^*}{\partial p} \right) + B \left( -\frac{\omega \partial \tilde{\theta}^*}{\partial p} \right) = \langle LH \rangle + \langle SH \rangle + \langle Q_r \rangle, \]

(4.2)

where the left hand side represent the contributions of spatial patterns to the global mean change.

By subtracting Eq. (4.2) from Eq. (4.1), one can derive the equation for pattern formation, i.e.,

\[ B \left( \omega' \frac{\partial \tilde{\theta}^*}{\partial p} \right) + B \left( -\frac{\omega \partial \tilde{\theta}^*}{\partial p} \right) + B \left( -\frac{\partial \langle \tilde{\theta}' \rangle}{\partial p} - \left( -\frac{\omega \partial \tilde{\theta}^*}{\partial p} \right) \right) = LH^* + SH^* + Q_r^*. \]

Consistent with the “weak temperature gradient” approximation (Bretherton and Sobel 2003), the horizontal variations of potential temperature are much less than the
tropical mean (Fig. 1.3): \( \frac{\partial \tilde{\Theta}^*}{\partial p} \ll \frac{\partial \tilde{\Theta}}{\partial p} \), and \( \frac{\partial \Theta^*}{\partial p} \ll \frac{\partial \langle \Theta' \rangle}{\partial p} \). Therefore, one can simplify the above equation into

\[
B \omega \frac{\partial \tilde{\Theta}^*}{\partial p} + B \omega \frac{\partial \tilde{\Theta}}{\partial p} = -B \omega \frac{\partial \langle \Theta' \rangle}{\partial p} + LH^* + SH^* + Q_{R^*}^*.
\]  

(4.3)

Eq. (4.3) forms a diagnostic framework that solves regional response to imposed heating patterns in global warming. This framework defines terms solvable with the LBM as the response and places them on the left hand side. They include the warming patterns (\( \Theta^* \)) and circulation change (vertical motion, \( \omega' \)). The imposed forcing terms are arranged on the right hand side.

The major modification of the LBM is to treat \( -B \omega \frac{\partial \langle \Theta' \rangle}{\partial p} \) as a forcing term. As an effect of the global mean warming on pattern formation, it represents the mean advection of stratification change (MASC). It is part of atmospheric response in full GCMs. The MASC effect is important for circulation change, as illustrated in Fig. 4.1 and the following sections.

Figures 4.1a-f show the relative importance of Eq. (4.3) terms in the GFDL models. For SUSI, the terms representing warming patterns (Fig. 4.1c), SH*, and \( Q_{R^*} \) (Fig. 4.1e) are negligible and the circulation change (Fig. 4.1c) is largely in balance with the MASC and LH* (Fig. 4.1a). For A1B, the \( Q_{R^*} \) effect is also significant (Fig. 4.1f). It features maxima in the western equatorial Pacific and the Tibetan Plateau, resembling the LH* patterns (\( r = 0.64 \)). The LBM experiment (not shown) indicates that the \( Q_{R^*} \) effects are similar to the LH* on changes of temperature, wind shear, Walker and Hadley circulation, albeit much weaker. Actually, the \( Q_{R^*} \) is dominated by the LW\(_C^*\) (\( r = 0.88 \), Fig. 4.1f), the cloud radiative effect associated with deep convection that produces LH*. They are combined and still denoted as LH* in Figs.
4.1h, j and hereinafter. Unavailable in the data set, the SH*$_{A1B}$ could be important in the boundary layer but should be weak in the free troposphere. Thus, only the circulation change, MASC, and LH* terms reach a magnitude of 0.1 K day$^{-1}$. Their sums are small for both runs (Figs. 4.1g-j), indicating that they are nearly balanced in the tropics.

By neglecting small terms (sensible and radiative heating), Eq. (4.3) can be further simplified as
\[
\bar{B}\omega \frac{\partial \theta^*}{\partial p} + B\omega \frac{\partial \theta}{\partial p} = -\bar{B}\omega \frac{\partial \langle \theta \rangle}{\partial p} + LH^*.
\] (4.4)

By imposing MASC and LH* diagnosed from the GCM output, LBM solves for both temperature patterns and circulation change including vertical motion.

### 4.1.2 MASC mechanism

Model simulations suggest that tropical tropospheric warming follows the moist adiabat in the vertical (Fig. 1.3), resulting in an increase in dry static stability $-\frac{\partial \langle \theta \rangle}{\partial p}$.

With this upward-increasing atmospheric warming, the MASC mechanism refers to the cold/warm advection in climatological convective/subsidence regions. It is much smaller than $\bar{B}\omega \frac{\partial \theta^*}{\partial p}$ (Eq. 4.3) for interannual variability but is of great importance for global warming. This difference is caused by the fact that $O\left(\frac{\text{SST}^*}{\langle \text{SST} \rangle}\right) >> 1$ for the former but $O\left(\frac{\text{SST}^*}{\langle \text{SST} \rangle}\right) < 1$ for the latter.

The meridional structure of tropospheric warming in Fig. 1.3 suggests the importance of the MASC effect. In the mid-troposphere (e.g. 600 hPa) in the
subtropics (20°-30°), the enhanced warming is anchored by the climatological subsidence, while the reduced warming over the equator by the mean upward motion. This feature is quite robust among CMIP3 models (see Fig. S2 of Lu et al. 2007). Similar collocations of minimum in mid-tropospheric warming with convection and maximum with subsidence centers can also be found in Fig. 1.2. Correlations between the annual mean warming patterns and climatological pressure velocity are 0.38 (Table 1.3). This correlation is moderate since the LH* also affects the tropospheric warming patterns.

4.1.3 Forcing distributions

Besides LH*, MASC is a major forcing mechanism for the circulation response to global warming in the LBM (Eq. 4.4). This section discusses the annual-mean tropical distributions of both terms in SUSI (Fig. 4.1a) and A1B (Fig. 4.1b) runs. The MASC follows the patterns of climatological vertical motion, which is similar but slightly different between the two runs, due to SST biases of the coupled model (A1B) compared to observations (SUSI), e.g., the double ITCZ bias (de Szoeke and Xie 2008). The effect of the SST biases on the MASC is discussed in Section 7.1.2. In both SUSI and A1B, $T^*$ features negative centers in the Indo-Pacific warm pool that extend to the ITCZ and SPCZ, and positive centers over subsidence regions (Fig. 1.2).

The LH* distributions resemble the precipitation patterns and are quite different between the SUSI and A1B, due to SST patterns in A1B (Xie et al. 2010). In the SUSI case, it is roughly opposite to the MASC ($r = -0.39$). This is not coincident since the LH*$_{SUSI}$ warms the convective region, representing the “wet-get-wetter” pattern of rainfall change, while the MASC cools the convective region by definition. The so-called “upped-ante” mechanism calls for reduced rainfall on the convective boundary...
between the warm pool and the equatorial cold tongue due to inflow of dry air from the subsidence regions (Chou and Neelin 2004; Chou et al. 2009), reinforcing the MASC effect. Following rainfall patterns (Xie et al. 2010), the LH*_{A1B} distribution features maximum on the equator with a peak in the western Pacific. The heating is greater in the Northern than the Southern Hemisphere. It is worth mentioning that with SST patterns, spatial variability in LH* (precipitation) is greater in magnitude than in SUSI, especially clear in the equatorial and zonal means (Figs. 4.1g-j).

The equatorial forcing (Figs. 4.1g, h) is important for the Walker circulation change. Both the MASC and LH*_{SUSI} (Fig. 4.1g) weaken the Pacific Walker cell, the latter due to the above-mentioned “upped-ante” mechanism. However, LH*_{A1B} (Fig. 4.1h) tends to accelerate the circulation, especially in the eastern Pacific, with a maximum effect at 140°W.

The zonal mean forcing is important for the Hadley circulation changes. The MASC tends to weaken the Hadley circulation (Fig. 4.1i) by definition, cooling/warming for upward/downward motions. By contrast, LH*_{SUSI} accelerates the Hadley cell, consistent with the theory of Lindzen (1990). The MASC and LH*_{SUSI} almost balance each other in the deep tropics (r = -0.77) but the MASC is slightly greater in the subtropics. The zonal mean LH*_{A1B} (Fig. 4.1j) is strongly influenced by SST patterns, especially the equatorial warming peak and the inter-hemispheric asymmetry with a greater SST increase north than south of the equator. LH*_{A1B} has a negative correlation (r = -0.55) with the MASC effect, but it is weaker than LH*_{SUSI}.

### 4.1.4 Experimental designs

In Eq. (4.4), MASC and LH* are prescribed in the LBM as follows.

\[
\text{SUSI: } \text{MASC} + \text{LH*}_{\text{SUSI}} \quad \text{and} \quad \text{(4.5)}
\]
A1B: MASC + LH*\text{A1B} = MASC + LH*\text{SUSI} + LH*\text{RAD} + LH*\text{SST}. \quad (4.6)

Annual-mean MASC and LH* are diagnosed from the SUSI and A1B runs. For the A1B run, LH* is further decomposed into components due to radiative forcing, SUSI and SST patterns, as detailed in Table 4.1. By definition, MASC is likely to be similar among different models with some variations in model climatology. The latent heating term, and its sub-components may differ substantially among models. See Xie et al. (2010) for a comparison between models developed at GFDL and NCAR.

The LBM is linearized around the annual mean climatology of AM2.1 during 1983-91 for the SUSI-related experiments, and that of CM2.1 during 1996-2005 for the A1B ones. The model is integrated for 60 days to equilibrium. Since the LBM does not represent the synoptic eddy-mean flow interaction properly, the thermodynamic forcing is restricted in 40°S-40°N.

### 4.2 General survey of tropospheric temperature and wind shear changes

This section presents the LBM response to individual and combined forcing factors in Eqs. (4.5, 4.6) and compares the results with the GCMs. Besides the tropospheric warming patterns, the vertical wind shear changes are examined in both vector and magnitude of absolute zonal wind shear (|U_{300}-U_{850}|) change.

Figure 4.2 shows the results for the SUSI run. As one expects, the MASC (Fig. 4.2a) cools the deep tropics, especially the tropical Pacific, while it warms the subtropical subsidence regions. The MASC response is generally symmetric about the equator. Consistent with the thermal-wind relation, the tropical tropospheric wind shear is reduced almost everywhere. By contrast, LH*\text{SUSI} (Fig. 4.2b) tends to warm
the tropics and cool the subtropics, especially in the South Pacific, causing wind shear to increase in the subtropical Pacific and the tropical Atlantic region of frequent cyclone development (Vecchi and Soden 2007c). The MASC and LH* SUSI responses oppose each other and the spatial correlation is -0.69 for tropospheric temperature and -0.4 for wind shear (Table 4.2). In the central and eastern equatorial Pacific, however, LH* SUSI reduces the wind shear, apparently due to the “upped-ante” cooling to the west of the climatological equatorial cold tongue (Fig. 4.1a). LBM simulations forced by observed diabatic heating have been very successful in capturing circulation change associated with natural variability such as ENSO. If one was to do the same for global warming as in the SUSI run, he/she would be surprised that the simulated circulation change is almost opposite to the GCM results. This illustrates the importance of MASC.

The combined MASC and LH* SUSI response (Fig. 4.2c) resembles the AM2.1 (SUSI) results (Fig. 4.2d), with spatial correlation for tropospheric temperature/wind shear changes at $r = 0.34/0.64$. This illustrates the skills of LBM in simulating the circulation response to global warming. The magnitude of LBM results is slightly weaker than that of AM2.1, possibly due to the arbitrarily set damping and horizontal diffusion. The tropospheric cooling in the mid-latitude South Pacific is not strong enough in the LBM, possibly related to eddy effects. The peaks of subtropical warming by LH* SUSI (Fig. 4.2b) are slightly shifted to lower latitudes than in AM2.1 (Fig. 4.2d), and the MASC (Fig. 4.2a) helps to adjust them to the right positions (Fig. 4.2c). The cooling over the Tibetan Plateau is due to the MASC (Figs. 4.2a, c, d) associated with orographic convection anchored by the Himalayas and advected by the westerly jet stream. The tropical Pacific wind shear is generally reduced (Figs.
4.2c, d), with the contributions from LH_{SUSI} in the central Pacific basin (Fig. 4b) and from MASC effect in the subtropics (Fig. 4.2a).

The zonal means better illustrate the importance of MASC. The MASC and LH_{SUSI} effects (Fig. 4.2e) are both symmetric about the equator, and oppose each other (r = -0.85). Their combined (MASC+LH_{SUSI}) effect follows the MASC response, and resembles the AM2.1 results (r = 0.53) to the north of 20°S. For the wind shear change (Fig. 4.2f), the MASC effect again opposes the LH_{SUSI} effect, with r = -0.66 (Table 4.2). Primarily following the shape of MASC, the total effect weakens the tropical wind shear, resembling the AM2.1 results (r = 0.69). It is noteworthy that the MASC effects flip sign at around 30°S and 30°N (Fig. 4.2f), where the climatological wind shear/meridional temperature gradients are strongest. This is equivalent to a poleward shift of the jet streams/expansion of the Hadley cell, consistent with the argument of Fu et al. (2006).

For the A1B run, the LH* effects include components induced by RAD and SST patterns. The RAD effect (Fig. 4.3a), with the prescribed SST, is much smaller than the SST pattern effect, with a modest contribution over the mid-latitude Eurasian and North American continents. It hardly affects the Pacific Walker circulation but reduces wind shear in the western tropical Indian Ocean. The RAD influence on the Hadley circulation is weak in zonal mean (Fig. 4.3e) but reinforces the MASC (Fig. 4.2e), with r = 0.62. The SST patterns have a significant influence on the inter-hemispheric asymmetry of tropospheric warming (Figs. 4.3b, e). The SST patterns enhance wind shear in the eastern equatorial Pacific, South Pacific, and South Atlantic, but reduce it in the eastern tropical Indian Ocean, western equatorial Pacific, tropical North Pacific, and tropical North Atlantic (Figs. 4.3c, d, e). The LBM response to SST patterns is too strong compared to the wind shear change in CM2.1 (Fig. 4.3d).
The comparison of LBM results with CM2.1 is greatly improved (Table 4.2) from MASC+LH* SUSI (Fig. 4.2c) to MASC+LH* A1B (Fig. 4.3c). The reduction in wind shear over the tropical Indian Ocean is a major difference between SUSI (Figs. 1.2a and 4.2d) and A1B (Figs. 1.2b and 4.3d), due to both RAD and SST effects. Likewise the increased wind shear in the eastern tropical North Pacific is due to the SST patterns. The MASC effect is most pronounced in the Northern Hemisphere, helping maintain the cooling in the deep tropics and the associated wind shear reduction in the North Pacific (Figs. 4.2c, e and 4.3b, c, e). The combined MASC+LH* A1B effect increases the correlation with CM2.1 by 0.2 compared to the LH* A1B effect.

The MASC effect is much weaker than LH in interannual variability of the atmosphere, but the above analyses show that it is important in global warming. Specifically, the MASC mechanism acts to reduce tropical tropospheric meridional temperature gradients and zonal wind shear. Along with the “wet-get-wetter” mechanism (Chou and Neelin 2004), the MASC is an important mechanism of pattern formation when SST warming is spatially uniform.

### 4.3 Overturning circulations in LBM

This section examines the forcing factors for overturning circulation changes based on LBM simulations and comparison with GCMs.

#### 4.3.1 Walker circulation

Upper-tropospheric velocity potential ($\chi$) was used to diagnose the large-scale atmospheric circulation (Gastineau et al. 2009). $\chi$ at the 250 hPa level is calculated to
characterize the change of the Walker circulation, by solving the following Poisson equation on the globe (Tanaka et al. 2004),

$$\nabla \cdot \mathbf{V} = -\Delta \chi,$$

where \( \mathbf{V} \) is wind velocity. Figures 4.4a-h show the horizontal distributions of the changes in 250 hPa \( \chi \) (color shading) and divergent wind (vectors) in response to various forcing factors. The climatological mean (contours) features the Pacific Walker circulation with divergence/convergence (negative/positive \( \chi \)) on the western/eastern basins.

The MASC (Fig. 4.4a) tends to reduce the divergence/upward motion over the Indo-Pacific warm pool, by its adiabatic cooling effect illustrated in Fig. 4.1. By contrast, LH\textsuperscript{* SUSI} (Fig. 4.4b) induces divergence over the eastern Pacific, due to the heating peak there (Fig. 4.1a), and weak convergence at 170°E (Fig. 4.4i), due to the “upped-ante” mechanism. Even without any SST gradient, the MASC and LH\textsuperscript{* SUSI} act together to slow down the Walker circulation over the vast area of the Pacific (Figs. 4.4e, f). High spatial correlation \((r = 0.78, \text{Table 4.3})\) appears between MASC+LH\textsuperscript{* SUSI} and AM2.1. \( \chi \) on the equator (Fig. 4.4i), representing the Walker circulation, shows that the MASC reduces the zonal wind between 130°E and 130°W, while the LH\textsuperscript{* SUSI} reduces it in 130°-90°W. The higher correlation with AM2.1 of MASC (0.74) than of LH\textsuperscript{* SUSI} (0.55) indicates the MASC effect is key to the success of LBM simulation of the weakened Walker circulation \((r = 0.87 \text{ for MASC+LH\textsuperscript{* SUSI}})\).

For changes in A1B, the RAD slightly strengthens the Walker circulation (Figs. 4.4c, j). The peak of SST warming in the mid-equatorial Pacific (Fig. 4.1b), contribute to a divergence center at 175°E in the central Pacific (Fig. 4.4d, j), reducing the Walker circulation over the western Pacific but accelerating it in the eastern basin.
Primarily maintained by the MASC, the zonal wind reduction of MASC+LH* A1B extends to 140°W (Fig. 4.4g, j) and is almost identical to CM2.1 (Fig. 4.4h, j), with very high spatial correlation (0.94) especially on the equator (0.97).

Thus, the MASC is very important for the weakening of the Walker circulation, even more so in the A1B simulation. It has essential effect over the west-to-mid equatorial Pacific. The SST contribution to $\chi$ is strong over the entire Pacific, but its accelerating effect is opposed by the MASC to the west of 140°W. Considering SST patterns of CM2.1 are among the strongest (spatial standard deviation is 0.24 K per K tropical mean warming, as compared to the CMIP3 model ensemble of 0.17 ± 0.05 K per K), the strong MASC effect explains that the weakening of the Walker cell is very robust among the CMIP3 models (Vecchi and Soden, 2007a).

### 4.3.2 Hadley circulation

The Hadley circulation change is represented by the zonal-integrated meridional mass stream function ($\psi$, Fig. 4.5), and by the zonal mean 250 hPa $\chi$ (Figs. 4.4k, l). $\psi$ is calculated as follows,

$$\psi = \frac{1}{g} \int_{100 \text{ hPa}}^{\rho} \vec{V} dp,$$  \quad (4.7)

where $g$ is gravity and $\vec{V}$ is zonal-integrated meridional wind velocity. Figure 4.5 shows the climatological annual mean Hadley circulation in contours, which features a clockwise/anticlockwise tropical cell (positive/negative values of $\psi$) in the Northern/Southern Hemisphere, with upward motion over the ITCZ (~5°N) and subsidence in the subtropics (20°-40°).

The MASC forces a significant weakening of the Hadley circulation (Figs. 4.5a and 4.4k), while LH* SUSI accelerates it (Figs. 4.5b and 4.4k) at a reduced magnitude.
The combined effect follows the MASC patterns ($r = 0.43$ for $\psi$ and $0.66$ for $\chi$) and is quite similar ($r \geq 0.67$) between the LBM (Figs. 4.5e and 4.4k) and AM2.1 (Figs. 4.5f and 4.4k). Thus, the MASC is the major driver for the Hadley circulation reduction in the AM2.1 SUSI experiment. Although the LH*$_{SUSI}$ effect is weak here, its sign is always opposite to the MASC.

In response to GHG increase, the RAD effect causes a weak reduction in the Hadley cell in the Northern Hemisphere (Figs. 4.5c and 4.4l), with the same sign as the MASC. The SST patterns (specifically the equatorial peak, Fig. 4.1b) induce an anomalous cell on either side of the equator (Fig. 4.5d), with a meridional scale of the equatorial Rossby radius of deformation ($\sim 15^\circ$). This corresponds to an acceleration of the Hadley cell near the equator in response to the equatorial peak in SST warming (Liu et al. 2005; Xie et al. 2010). Because the SST warming is greater in the Northern than the Southern Hemisphere (Fig. 4.1b), the Northern SST-induced equatorial cell is weak, especially above 500 hPa, and the Hadley cell is reduced in the Northern off-equatorial region. As a result, only the southern cell intensification is visible in the upper troposphere (Fig. 4.4l).

The total effect of the four factors (MASC+LH*$_{A1B}$, Fig. 4.5g) explains the CM2.1 changes (Fig. 4.5h) quite well ($r = 0.88$). The Hadley cell weakens in the Northern Hemisphere and accelerates in the Southern Hemisphere. For the Northern Hemispheric cell, the SST-induced acceleration is offset by the LH*$_{SUSI}$ while both the MASC and RAD effects contribute to the reduction. In the Southern Hemisphere, the SST-induced cell dominates over the MASC+LH*$_{SUSI}$ effect, accelerating the Hadley cell.

Figures 4.6 and 4.7 show the seasonal cycle of the Hadley circulation change. The boreal summer is represented by mean during June-July-August (JJA), and the boreal
winter by December-January-February (DJF). In both seasons, the MASC effect dramatically reduces the Hadley cell (Figs. 4.6a and 4.7a). The LH* SUSI accelerates the Hadley cell in the Northern Hemisphere and vice versa in the Southern Hemisphere (Figs. 4.6b and 4.7b). In total, the MASC dominates the SUSI change (Figs. 4.6e and 4.7e). The LBM well simulates the weakening of the Hadley cell in AM2.1 in both seasons (Figs. 4.6f and 4.7f), with correlations of 0.84 and 0.83 in JJA and DJF, respectively.

In A1B, the RAD (Figs. 4.6c and 4.7c) and SST pattern (Figs. 4.6d and 4.7d) effects does not show significant seasonal variation and share features common to the annual means (Figs. 4.5c and d). As a result, in JJA, the Hadley cell reduction by the MASC (Figs. 4.6e and f) is offset by the SST pattern contribution near the equator (Fig. 4.6g), which resembles the CM2.1 results (Fig. 4.6h) with $r = 0.65$. For the Hadley cell in DJF, in addition to the reduction induced by the MASC effect (Figs. 4.7e and f), the SST patterns generates an anti-clockwise cell in the Southern Hemisphere (Figs. 4.7g and h), as similarly seen in the annual mean (Figs. 4.5g and h). The correlation between LBM and CM2.1 results reaches 0.94.

The LBM experiments indicate that the SST pattern-induced LH effect explains the difference in the Hadley cell changes between AM2.1 and CM2.1, shedding lights on the inter-model variations in CMIP3 models, which would be discussed in the next chapter.

### 4.4 Summary

A diagnostic framework that employs the LBM is developed to understand the tropospheric circulation change in global warming by decomposing thermodynamic forcing (TF) on the right hand sides of Eqs. (5) and (6) into the following terms
\[
TF = \text{MASC} + \text{LH}^*_{\text{SUSI}} + \text{LH}^*_{\text{RAD}} + \text{LH}^*_{\text{SST}}.
\]

The results demonstrate the LBM’s utility in global warming pattern studies to quantify the relative importance of various forcing factors. Global warming features a nearly horizontal-uniform increase of tropospheric temperature. The MASC is important in tropical circulation adjustment, comparable in magnitude to the LH\textsuperscript{*}_\text{SUSI} and SST pattern effects, while the RAD effect is much weaker.

Based on the LBM experiments, the questions raised in the beginning of this chapter can be answered. The MASC warms the subsidence and cools the convective regions, so that the subtropics are more warmed. The SST patterns also contribute to the tropospheric warming patterns but this effect is overcome by the MASC in zonal mean. The MASC weakens both the Hadley and Walker circulation, but the SST patterns have significant influence, especially on the uncertainty of the Hadley circulation change, as detailed below.

In response to a uniform SST increase without any gradient, the Walker circulation weakens due to both the MASC and LH\textsuperscript{*}_\text{SUSI} effects in the western and eastern Pacific, respectively. The SST patterns in CM2.1 significantly reduce the Walker circulation in the western Pacific but accelerate it in the eastern basin. Because the MASC effect is strong over a vast area of the equatorial Pacific, the Walker circulation slow down is robust among the CMIP3 models under the A1B scenario.

In response to a uniform SST warming, the Hadley circulation slow down due to the MASC effect. In the A1B simulation with CM2.1, however, the slow down is limited to the Northern Hemispheric cell while the Southern cell accelerates. The LBM results show that this asymmetric response is due to SST patterns, specifically an inter-hemispheric asymmetry with greater SST warming in the Northern than the
Southern Hemisphere. The meridional variations of SST warming in CM2.1 are among the strongest of the CMIP3 models. In some coupled models with weak SST patterns, e.g. the CCCMA CGCM3.1 T63, GISS AOM, MIROC3.2 Hires, and MRI CGCM2.3 (not shown), the MASC effect to slow down the Hadley circulation in both hemispheres can be stronger than the SST pattern effect, resulting in a weakened Hadley circulation in both hemispheres (Fig. 5.4a).

During the seasonal march of the Hadley circulation, the MASC and SST pattern effects are shown to be the major factors for the intensity change. As the MASC makes strong contributions to the weakening of the Hadley cell in both seasons, the SST pattern effect explains the difference between the SUSI and A1B changes seen in the GCM simulations.

For the tropical wind shear and tropospheric temperature changes, the boreal summer simulations by the LBM show similar skills to the annual mean ones in reproducing GCM results. In the boreal winter, LBM’s simulation skills are much lower. This seasonality appears consistent with the theory of Schneider et al. (2010).

An important parameter for this theory is the local Rossby number,

$$ Ro = -\frac{\zeta}{f}, $$

where $\zeta$ is relative vorticity, and $f$ is the Coriolis parameter. When $Ro \to 1$, the angular momentum conservation degenerates and provides no constraint on the mean meridional mass flux, and the tropical circulation change responds directly to the thermal driving. When $Ro \to 0$, circulation is dynamically constrained and change in the eddy momentum flux divergence plays an important role. Figure 5 of Schneider et al. (2010) shows that only in boreal summer the local $Ro$ are larger than 0.5 in much of the upper branch of the Hadley cell in both hemispheres, while in boreal winter, the
local $Ro$ is smallest. The eddy effects (Lu et al. 2007; Schneider et al. 2010) are not considered in this chapter but worth in-depth investigations.
Table 4.1. Descriptions of the LBM experiments.

<table>
<thead>
<tr>
<th>Name</th>
<th>Forcing</th>
<th>Contribution</th>
</tr>
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<tbody>
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<td>MASC</td>
<td>MASC diagnosed from AM2.1 (SUSI) only</td>
<td>SUSI/A1B</td>
</tr>
<tr>
<td>LH*&lt;sub&gt;SUSI&lt;/sub&gt;</td>
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<td>SUSI/A1B</td>
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<tr>
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<td>MASC and LH* diagnosed from AM2.1 (SUSI)</td>
<td>SUSI/A1B</td>
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<tr>
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<tr>
<td>LH*&lt;sub&gt;A1B&lt;/sub&gt;</td>
<td>LH* diagnosed from CM2.1 (A1B) only</td>
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<tr>
<td>MASC+LH*&lt;sub&gt;A1B&lt;/sub&gt;</td>
<td>MASC and LH* diagnosed from CM2.1 (A1B)</td>
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Table 4.2. Spatial correlation coefficient ($r$, 40°S-40°N) of annual mean 300-850 hPa averaged air temperature warming patterns/absolute zonal wind shear ($ash$) change among LBM and GFDL models.

<table>
<thead>
<tr>
<th>$r$ (40°S-40°N)</th>
<th>MASC</th>
<th>LH* SUSI</th>
<th>LH* SST</th>
<th>MASC+ LH* SUSI</th>
<th>MASC+ LH* A1B</th>
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<td>$T_{va}^*$</td>
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<td>0.14</td>
<td>0.13</td>
<td>----</td>
<td>0.34</td>
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<td>----</td>
<td>0.53</td>
</tr>
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<td>----</td>
<td>0.64</td>
</tr>
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<td>-0.32</td>
<td>----</td>
<td>0.69</td>
</tr>
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<td>0.06</td>
</tr>
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<td>0.15</td>
<td>0.43</td>
<td>0.24</td>
</tr>
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<td>0.08</td>
<td>0.59</td>
<td>0.07</td>
</tr>
<tr>
<td>$U_{ash}$</td>
<td>MASC</td>
<td>----</td>
<td>-0.69</td>
<td>-0.37</td>
<td>----</td>
</tr>
<tr>
<td>$T_{va}^*$</td>
<td>MASC</td>
<td>----</td>
<td>-0.85</td>
<td>-0.54</td>
<td>----</td>
</tr>
<tr>
<td>$U_{ash}$</td>
<td>MASC</td>
<td>----</td>
<td>-0.40</td>
<td>-0.23</td>
<td>----</td>
</tr>
<tr>
<td>$T_{va}^*$</td>
<td>MASC</td>
<td>----</td>
<td>-0.66</td>
<td>-0.37</td>
<td>----</td>
</tr>
</tbody>
</table>
Table 4.3. Spatial correlation coefficient ($r$, 40°S-40°N) of annual mean changes of 250 hPa velocity potential ($\chi$) and meridional stream function ($\psi$) among LBM and GFDL models.

\begin{tabular}{|c|c|c|c|c|c|}
\hline
\text{$r$ (40°S-40°N)} & \text{x-y} & \text{em} & \text{zm} & \text{MASC} & \text{LH* SUSI} & \text{LH* SST} & \text{MASC+ LH* SUSI} & \text{MASC+ LH* A1B} \\
\hline
\text{\textit{\chi}_\text{AM2.1}} & 0.62 & 0.45 & ---- & 0.78 & \text{\textit{\chi}_\text{CM2.1}} & 0.03 & 0.67 & 0.65 & 0.53 & 0.94 \\
& 0.74 & 0.55 & ---- & 0.87 & & -0.11 & 0.68 & 0.57 & 0.54 & 0.94 \\
& 0.66 & -0.48 & & 0.68 & & & 0.68 & 0.57 & & \\
& -0.11 & 0.68 & 0.57 & 0.54 & 0.94 & & & & & \\
\text{\textit{\chi}_\text{MASC}} & ---- & 0.05 & -0.48 & ---- & \text{\textit{\chi}_\text{CM2.1}} & -0.20 & 0.48 & 0.81 & -0.09 & 0.94 \\
& ---- & 0.05 & -0.48 & ---- & & -0.14 & -0.54 & & & \\
& ---- & 0.05 & -0.48 & ---- & \text{\textit{\psi}_\text{AM2.1}} & 0.43 & 0.08 & ---- & 0.67 & ---- \\
& ---- & 0.05 & -0.48 & ---- & & -0.81 & -0.69 & & & \\
& ---- & 0.05 & -0.48 & ---- & \text{\textit{\psi}_\text{CM2.1}} & -0.25 & 0.45 & 0.62 & 0.14 & 0.88 \\
& ---- & 0.05 & -0.48 & ---- & & -0.68 & -0.65 & ---- & ---- & \\
\hline
\end{tabular}

\textit{em} denotes equatorial mean (15°S-15°N).
Fig. 4.1. Annual mean distributions (a-f) of 300-850 hPa averaged terms in Eq. (4.3) in 0.1 K day⁻¹ (CI 0.05 K day⁻¹; zero omitted), along with their equatorial means (g, h, 5°S-5°N) and zonal means (i, j) in SUSI and A1B runs. In (c) and (d), $T^*$ and CC denote the warming patterns and circulation change terms, respectively. In (f), SH* is unavailable in CM2.1 output, so instead, LW_C* is plotted to show the relation between $Q_R^*$ and LH*. In (h), (j), and hereinafter, LH* represents the combined effect of LH* and $Q_R^*$ in A1B run. In (g)-(j), Sum means the summation of MASC, LH* and CC to show their approximate balance (Eq. 4.4).
Fig. 4.2. 300-850 hPa averaged air temperature warming patterns (color shading, K), wind shear (vectors, m s$^{-1}$), and absolute zonal wind shear (contours, CI 0.5 m s$^{-1}$; zero omitted) changes in LBM forced by annual mean (a) MASC, (b) LH* SUSI, and (c) MASC+LH* SUSI, compared with (d) AM2.1. (e), (f) are the zonal means of the warming patterns and absolute shear change, respectively.
Fig. 4.3. 300-850 hPa averaged air temperature warming patterns (color shading, K), wind shear (vectors, m s$^{-1}$), and absolute zonal wind shear (contours, CI 0.5 m s$^{-1}$; zero omitted) changes in (a) RAD, LBM forced by annual mean (b) LH$^*$ SST, and (c) MASC+LH$^*$ A1B, compared with (d) CM2.1. (e), (f) are the zonal means of the warming patterns and absolute shear change, respectively.
Fig. 4.4. Annual mean changes of 250 hPa velocity potential ($10^5$ m$^2$ s$^{-1}$) distribution (a-h, color shading) with the equatorial means (i, j) and zonal means (k, l). In (a)-(h), vectors are the changes of divergent wind (m s$^{-1}$), and contours (CI 20×$10^5$ m$^2$ s$^{-1}$; zero omitted) show the mean velocity potential for reference.
Fig. 4.5. Annual mean changes of the Hadley circulation presented by the zonal-integrated meridional stream function (color shading, $10^{10}$ kg s$^{-1}$) with the contours (CI $2\times10^{10}$ kg s$^{-1}$; zero omitted) showing the mean circulation for reference.
Fig. 4.6. Same as Fig. 4.5, but for JJA mean changes (CI $4 \times 10^{10}$ kg s$^{-1}$).
Fig. 4.7. Same as Fig. 4.6, but for DJF mean changes.
Chapter 5

Tropical overturning circulation change:

CMIP multi-model results

As is clear in Chapters 3 and 4, the MASC and SST patterns are both important for atmospheric circulation change in the coupled model. MASC is responsible for the slow-down of tropical circulation in SUSI, while SST patterns accounts much for the inter-model uncertainty, especially for the Hadley circulation. This chapter investigates what controls the response of tropical overturning circulation to global warming in a large model ensemble, and the role of SST warming patterns in particular. The results are presented based on three datasets: AGCM experiments to understand the CMIP3 ensemble mean response (Section 2.3), CFMIP2 simulations as part of CMIP5 (Section 2.2), and inter-model variability in CMIP3.

5.1 AGCM sensitivity experiments

Figure 5.1 illustrates the effect of CMIP3 ensemble mean SST patterns on the Hadley circulation with the CAM experiments. The Hadley circulation is represented by the zonal-integrated meridional mass streamfunction (Eq. 4.7). The CMIP3 ensemble-mean Hadley circulation change (Fig. 5.1a) features a weakening (slight strengthening) in the Northern (Southern) Hemisphere. The CAM_A1B simulation
(Fig. 5.1b) captures the Northern Hemispheric changes well, but predicts a slight reduction in the southern branch of the Hadley cell. Consistent with the MASC theory of Ma et al. (2012), the SUSI effect (Fig. 5.1c) weakens the Hadley circulation in both hemispheres. Because the southern branch of the Hadley cell is stronger in CAM than in CMIP3 ensemble mean, the reduction may also be stronger, resulting in the inconsistency between Figures 5.1a and b.

The $T^*$ effect (Fig. 5.1d) consists of two components: the equatorial peak (Fig. 5.1f) and inter-hemispheric asymmetry (Fig. 5.1e). The equatorial peak of the SST warming enhances the Hadley cell on either side of the equator. The south-to-north SST warming gradient causes a cross-equatorial circulation with an enhanced southern Hadley cell and a reduced northern one.

In total, both the SUSI and $T^*$ effects contribute to the reduction of the Hadley circulation in the Northern Hemisphere, but $T^*$ is the key factor for the enhancement of the southern Hadley cell. This is consistent with the finding of Ma et al. (2012) that the spread of cross-equatorial SST gradient explain much variability in the Southern Hadley cell change among CMIP3 models, a point to be returned to in Section 5.3.

### 5.2 CMIP5-CFMIP2 simulations

The CFMIP2 experiments isolate the direct radiative, SUSI, and SST pattern effects from the CMIP5 coupled models. The Hadley circulation is represented by the zonal-integrated meridional mass streamfunction ($\psi$) at 500 hPa, and the Walker circulation by equatorial (15°S-15°N) upper-tropospheric (250 hPa) velocity potential ($\chi$).

Figure 5.2 shows the 500-hPa $\psi$ changes due to different factors. The SUSI effect (Fig. 5.2a) tends to reduce the Hadley cell robustly. Since the MASC effect (Ma et al.
2012) is supposed to be identical among models, the inter-model variations come mainly from LH. The RAD effect (Fig. 5.2c) is slight weakening, reinforcing the SUSI. The equatorial peak in SST warming (Fig. 5.2b) accelerates the Hadley cell near the equator. In the subtropics, the south-to-north warming gradient significantly accelerates the southern cell. The combined effect (Fig. 5.2d) makes a robust slow-down of the Hadley circulation in the northern subtropics. Whereas in the region near and south of the equator, the opposing SUSI and SST pattern effects result in weak and highly uncertain ensemble-mean change, subject to inter-model differences brought by SST patterns and model physics. This is consistent with the CMIP3 results (Fig. 5.4a) and CAM experiments (Fig. 5.1).

The 250-hPa χ (Fig. 5.3) shows a robust slow-down of the Pacific Walker circulation in SUSI (Fig. 5.3a). The RAD (Fig. 5.3c) has a weak enhancing effect. SST patterns (Fig. 5.3b) bring large uncertainty but do not offset the SUSI effect. As a result, the Walker circulation slow-down is robust (Fig. 5.3d), similar to the CMIP3 results (Fig. 5.4b).

The MASC effect (Ma et al. 2012) in SUSI slows down both the Hadley (Fig. 5.2a) and Walker (Fig. 5.3a) circulations. The RAD effect is weak (Figs. 5.2c and 5.3c). The SST pattern effect (Figs. 5.2b and 5.3b) introduces considerable uncertainty to both types of overturning circulations. The ensemble-mean effects of SST patterns are quite different between them. $T^*$ contributes little to the Walker circulation (Fig. 5.3b) so its slow-down remains robust (Figs. 5.3a, d). For the Hadley circulation by contrast, the ensemble-mean $T^*$ counteracts the SUSI effect (Fig. 5.2b) and makes ψ flat and uncertain in the southern subtropics and near the equator (Fig. 5.2d).

**5.3 CMIP3 inter-model variability**
This section investigates the SST pattern effect on inter-model variations of overturning circulation changes in CMIP3. Figure 5.4 shows the climatology and change of 500-hPa $\psi$ (Fig. 5.4a) and 15°S-15°N averaged 250-hPa $\chi$ (Fig. 5.4b), including both the CMIP3 ensemble mean and spread (shading). For the Hadley stream function, there are large inter-model variations in 15°S-15°N, so large that they can alter the sign of the Hadley circulation change in the southern subtropics and equatorial region. In the northern subtropics, the Hadley cell slow-down is robust, consistent with Figs. 5.1a and 5.2d. This asymmetry is due to the spatial variations in ensemble mean SST warming as discussed above. The Pacific Walker circulation (Fig. 5.4b) shows a robust reduction, consistent with previous studies (Vecchi et al. 2006; Vecchi and Soden 2007a). The inter-model variations appear on the same order of magnitude as those for the Hadley circulation. Below investigates the cause of the inter-model variability.

In order to examine the SST pattern effect on the inter-model variations of the overturning circulation changes, two sets of inter-model SVD analyses are performed among the CMIP3 GCMs. One is between the zonal mean SST and 500-hPa $\psi$, and the other is between the equatorial SST and 250-hPa $\chi$. Specifically, a conventional SVD analysis is applied to 22 pairs of variables simulated by the 22 CMIP3 models.

The meridional SST modes (Fig. 5.5) feature the south-to-north gradient and the equatorial peak (Ma et al. 2012). Both modes significantly influence $\psi$ in the region with high inter-model variability (Fig. 5.4a). The first SVD mode (Fig. 5.5a) is anti-symmetric with positive/negative anomalous SST in the Northern/Southern Hemisphere. This causes a cross-equatorial anticlockwise circulation between 15°S and 15°N that represents an enhanced/weakened Hadley cell south/north of the equator. The second mode (Fig. 5.5b) is symmetric and features an enhanced Hadley
circulation on either side of the equator (within 20°S-20°N), driven by the equatorial peak of SST warming. The leading zonal modes (Fig. 5.6) represents inter-model variability in zonal SST gradient across the Pacific, associated with changes in the Walker circulation.

Then, the first two SVD modes of $\psi$ and $\chi$ are removed from the inter-model variability. Inter-model variations of the Hadley and the Pacific Walker circulations are both dramatically reduced (Fig. 5.4), indicating that the SST warming patterns are a major source of uncertainty in changes of overturning circulations. This does not come as a surprise as the leading two modes explain 82% and 69% of the inter-model variability in the Hadley (Fig. 5.5) and Walker (Fig. 5.6) circulations, respectively. The residual uncertainty may be due to differences in model physics.

Figure 5.4 sheds light on the reason why the Hadley circulation change is not as robust as the Walker circulation. Because inter-model variations in SST patterns induce similar amount of uncertainty, the magnitude of the ensemble mean changes becomes important. The slow-down of the Walker and northern Hadley cells are robust because of large ensemble-mean changes. The Hadley circulation change near and south of the equator is not robust, because of small ensemble-mean change. This is illustrated more clearly by the CMIP5-CFMIP2 results (Figs. 5.2 and 5.3).

5.4 Summary

This chapter examines the changes of atmospheric overturning circulation in the CMIP simulations. While the slow-down of the Walker circulation is robust across the models, the change of the Hadley circulation is highly uncertain near and south of the equator. As the MASC effect weakens both overturning circulations commonly in the models, the SST patterns are a major source of the inter-model variability, accounting
for four fifth of the total variance. In ensemble mean, the SST pattern effect is weak for the Walker circulation, so that the slow-down due to MASC is strong enough to overcome the inter-model variability. The Hadley circulation change, by contrast, is significantly affected by the SST patterns. In the northern subtropics, the south-to-north gradient reinforces the MASC effect and results in a robust slow-down. Whereas near and south of the equator, both the equatorial peak and meridional gradient accelerate the Hadley cell, counteracting the MASC effect, so that the Hadley circulation change is weak and subject to large inter-model variability induced by differences in the SST patterns across the models.
Fig. 5.1. Annual mean changes of the Hadley circulation in (a) the CMIP3 ensemble mean and (b-f) various CAM simulations. The Hadley circulation is represented by the zonal-integrated meridional streamfunction (color shading, in $10^{10}$ kg s$^{-1}$), with the contours (CI: $2 \times 10^{10}$ kg s$^{-1}$; the zero contour omitted) showing the mean circulation for reference.
Fig. 5.2. Annual mean changes of the 500-hPa zonal-integrated meridional streamfunction ($10^{10}$ kg s$^{-1}$) in CMIP5-CFMP2 simulations. The shading marks the uncertainty (ensemble mean ± standard deviation) among the five models.
Fig. 5.3. Same as Fig. 5.2, but for the 15°S-15°N averaged 250-hPa velocity potential (10^5 m^2 s^{-1}).
Fig. 5.4. Annual mean climatology and changes of the (a) 500-hPa zonal-integrated meridional streamfunction (in $10^{10}$ kg s$^{-1}$), and (b) 15°S-15°N averaged 250-hPa velocity potential (in $10^5$ m$^2$ s$^{-1}$) in CMIP3 A1B simulations. Gray/light red shading marks the uncertainty (ensemble mean ± standard deviation) of the 22 GCMs in climatology/change. The dark red shading marks the reduced uncertainty by removing the first two SVD modes on SST. The figure is scaled by the climatology so that one can compare the Hadley and Walker circulations.
Fig. 5.5. First two modes of the inter-model SVD analysis between the annual mean changes of zonal mean SST patterns and 500-hPa zonal-integrated meridional streamfunction among the 22 CMIP3 GCMs under the A1B scenario. Reproduced from Ma et al. (2012).
Fig. 5.6. Same as Fig. 5.5, but for SST and 250-hPa velocity potential along the equator, averaged in 5°S-5°N and 15°S-15°N, respectively.
Chapter 6

Conclusions

This study investigates the fundamental mechanisms controlling tropical regional rainfall and tropospheric circulation changes in response to global warming and points to the SST patterns as a major source of their inter-model variability. While large uncertainties exist among the CMIP models, two robust mechanisms are found. One is the MASC effect associated with tropical mean SST warming, which by definition is identical among the models, acting to slow down the tropical circulation. And the other is the “warmer-get-wetter” view among coupled models with SST change patterns, casting the SST patterns (including two leading modes) as a major control on regional precipitation change.

While SST increases everywhere in the tropics, precipitation change is to first order variable in space. The model simulations show that if the SST increases uniformly in space, the precipitation change follows a “wet-get-wetter” pattern. However, in coupled models with increased GHG concentration, a “warmer-get-wetter” pattern dominates. Although the SST patterns are only a fraction of the tropical mean SST warming, it takes a major control on the regional precipitation change through atmospheric circulation adjustments. Because equatorial waves flatten the tropospheric warming to a value determined by tropical-mean SST warming, the threshold for tropical convection rises and the SST pattern effect dominates the
rainfall response. In many parts of tropical oceans, convection is reduced despite an increase in local SST because the local warming falls below the tropical average. In other parts of the tropics where relative SST change is positive, precipitation generally increases.

As illustrated in Figure 1.1, take any two global-warming simulations, and the differences in projected rainfall change are obvious and substantial. Differences in patterns of SST warming explains about one third of inter-model variability in tropical rainfall change among CMIP3 models and one fourth in CMIP5 models.

Two major patterns reside in both ensemble mean and inter-model variations of the CMIP3 and CMIP5 models. One is a meridional dipole mode, with a warmer hemisphere anchoring enhanced rainfall and reduced trades. The other is a peak warming along the equator, which is more ENSO like in CMIP5 than in CMIP3. Note that the ensemble mean tends to underestimate the spatial variations. For instance, in individual CMIP3 models, the spatial standard deviations of SST warming in 20°S-20°N are 0.19 ± 0.04 K for a nominal 1K tropical-mean warming, considerably higher than the ensemble mean (0.12 K). If nature evolves as one of the model realizations, the importance of the SST change patterns would be more significant than what is shown here for the ensemble mean.

Based on a series of LBM simulations, this study shows that the MASC mechanism is crucial in weakening the tropical atmospheric circulation (including the Walker, Hadley cells and vertical shear) and reducing the meridional temperature gradients of the tropical troposphere. The MASC effect arises from adiabatic cooling/heating due to the vertical advection by climatological upward/downward motion on a background of global increase in static stability. MASC is an important mechanism for circulation adjustment to global warming.
With the inclusion of MASC, the LBM is able to simulate the effects of various forcing factors of global warming patterns. The MASC effect is shown to be important in tropical circulation adjustment, comparable in magnitude to the SST pattern effect. In interannual variability, the ratio of tropical mean/standard deviation for SST anomaly drops to ~1/7 (see Introduction), and the MASC and LH* SUSI effects become much weaker than the SST pattern effect, and the effect of SST pattern-induced LH dominates the tropical circulation anomalies. This is a fundamental distinction between interannual variability and global warming.

From the energetic point of view, the increased vertical gradients of potential temperature in global warming enhance the efficiency of the Hadley cell in transporting the dry static energy from the equator to the subtropics, assuming the strength of the Hadley cell does not change. This is equivalent to the MASC mechanism. Held and Soden (2006) shows that the increase of moisture convergence (supplying the LH) is insufficient to overcome the dry energy divergence. Therefore, more energy is transported poleward (Zelinka and Hartmann 2012), flattening the tropical tropospheric meridional temperature gradients, reducing the tropical wind shear, pushing the synoptic eddies poleward, expanding and weakening the Hadley circulation. During the warm phase of the ENSO, enhanced poleward energy transport is also found because of the wind change associated with enhanced SST gradient (Lu et al. 2008). In global warming, SST pattern effects on energy transport need further investigations but SST gradient change is much weaker than global mean warming. The MASC, instead of being caused by circulation change, drives the winds to respond differently from the ENSO with same sign of energy transport change.

In the CMIP ensemble simulations, both the MASC and SST pattern effects are important for the Hadley and Walker circulation response to global warming. If the
SST warming were spatially uniform, the Hadley and Walker circulation would both slow down as articulated in previous studies (Held and Soden 2006; Vecchi et al. 2006; Ma et al. 2012). The equatorial-peak warming accelerates the Hadley circulation on either side of the equator while the reduced warming in the Southern Hemispheric accelerates the southern Hadley cell. As a result, the northern Hadley cell shows a robust slow down while the Hadley circulation intensity change near and south of the equator is weak in the ensemble mean and subject to large inter-model variability. Uncertainty in SST warming patterns, dominated by the equatorial peak and cross-equatorial gradient modes, accounts for a whopping 82% of inter-model variability in Hadley circulation change among CMIP3 models. Compared to the Hadley circulation, the SST pattern effect on Walker circulation is weak, and the SUSI effect dominates. As a result, the slow-down of the Walker circulation is a robust feature across the CMIP models.

This study shows that SST patterns are important for future climate change, in the “warmer-get-wetter” pattern and Hadley circulation response. In addition, the SST patterns can explain the inter-model variations in TC frequency response to global warming (Zhao and Held 2012). The tropical SST effect is not limited to the tropics as shifts in tropical convection and circulation have major remote effects on climate change elsewhere, via atmospheric teleconnection (Shin and Sardeshmukh 2011) and by affecting modes of climate variability (Collins et al. 2010; Zheng et al. 2010).

Recent observational studies reveal robust evidence on air-sea coupling associated with slow-down of the Walker cell and coherent patterns of climate change over tropical oceans (Tokinaga and Xie 2011a; Tokinaga et al. 2012a, 2012b). Innovative model experimentations show that ocean-atmospheric interactions, specifically the WES feedback, are indeed at work in the formation of major SST warming patterns.
(Lu and Zhao 2012). The Bjerknes feedback might behave differently in CMIP3 and CMIP5 models. To the extent that changes in surface winds and ocean circulation are important for SST patterns, this study calls for investigations into ocean-atmosphere interactions that shape SST and precipitation changes, which is demonstrated to be important for inter-model variability (an important measure of uncertainty) in a large model ensemble.
Chapter 7
Discussion and outlook

7.1 Impact of mean SST biases

The climatological SST biases in the coupled models from observations may induce significant error in projections for global warming. This section examines the impact of these biases on rainfall prediction and the MASC effect.

7.1.1 Effect of SST biases on precipitation change

This section examines the influence of SST biases on precipitation change in the 5 CMIP5 models under 1pctCO2 and AMIPFuture scenarios. The 1pctCO2 means coupled model forced by CO$_2$ increase at 1% per year until quadrupling. The AMIPFuture experiment runs with AGCM forced by observational SST plus ensemble-mean change in CMIP3.

Fig. 7.1 shows the biases of climatological SST and precipitation between coupled models (1pctCO2) and observations (AMIP). Two errors in the coupled models stand out, including a stronger equatorial cold tongue with further westward extension and the double ITCZ with positive SST anomalies in the southeastern Pacific. This forms a zonal dipole bias in precipitation with an eastward gradient.
Two methods can be adopted to estimate the biases of precipitation prediction in the coupled models. One is based on the linear relationship between percentage rainfall change and SST patterns, and the other by directly comparing with the AMIP experiments.

From Eq. 3.1, one could derive the formula for rainfall change bias, i.e.,

\[ \nabla \delta P = \nabla P (C + \alpha T^* + \beta T) \]

(7.1)

where \( \nabla \) denotes the difference between coupled models and observations. Here the AMIPFuture run is used to train the linear model (\( \alpha = 56\% \ K^{-1} \), and \( \beta = 6\% \ K^{-1} \)) and the mean rainfall error in Fig. 7.1 is applied to calculate the bias in change. However, this method cannot resolve non-linear processes actually involved (Fig. 3.3).

In addition, difference of rainfall change between 1pctCO2 and AMIPFuture experiments can be used as an estimation of rainfall change bias. However, the SST pattern change in these two scenarios is somewhat different (not shown), which may interfere the bias signals.

Fig. 7.2 shows the rainfall change bias estimation by both methods compared with AMIPFuture rainfall change. The rainfall change (Fig. 7.2a) features an equatorial increase and subtropical reduction, as seen in the CMIP3 ensemble mean (Fig. 3.2). Between Figs. 7.2b and c, robust patterns in rainfall change biases resemble the mean biases (Fig. 7.1), i.e. the equatorial zonal dipole. However, because neither method is accurate, the attempt here calls for experimental designs specifically for detecting the rainfall change biases, e.g., AMIP tests with different SST climatology (coupled models vs. observations) but same SST pattern change.

### 7.1.2 Influence of SST biases on MASC effect
As illustrated in Figures 4.1a and b, there is slight difference in the MASC forcing between the SUSI and A1B, due to discrepancy in climatological vertical motion. The large-scale vertical motion is basically determined by the SST distribution, so the climatological SST biases have significant effect on this MASC difference.

Figures 7.3 and 7.4 show the influence of the SST biases on the MASC forcing (Eq. 4.3) in the GFDL models and the atmospheric response to this bias in MASC effect calculated by the LBM, respectively. The major large-scale pattern of the SST biases (Fig. 7.3a) includes a westward extension of the equatorial cold tongue and a 2° northward shift of the ITCZ. The former reduces the MASC forcing (Figs. 4.1a and b) in the western equatorial Pacific (Fig. 7.3c) and results in a slightly weaker slowdown of the Walker circulation (Fig. 7.4b) in A1B than in SUSI. The latter (Fig. 7.3b) enhances the Hadley circulation a bit to the north of the equator (Fig. 7.4c). The tropospheric temperature pattern due to the MASC bias includes a zonal dipole over the South Pacific, along with wind shear reduction over the southeastern Pacific and South America and enhancement in the equatorial Atlantic (Fig. 7.4a).

7.2 MASC in observations (AMIP)

Chapters 4 and 5 illustrate the importance of the MASC effect to atmospheric circulation change in global warming. However, the analyses are limited to models for future climate prediction. The section will examine the MASC effect in observations (based on AMIP experiment with GFDL AM2.1 forced with observed SST), taking the Walker circulation as an example. The MASC effect is calculated according to Eq. (4.4) with the mean pressure velocity and anomaly of potential temperature in the AMIP experiment. An EOF analysis is applied to separate the forcing patterns and temporal trend. The latter is then compared with the Walker circulation change in the
AMIP runs and observations, which is represented by an index defined as the anomaly of the surface zonal wind averaged in 5°S-5°N, 140°E-80°W.

The AMIP experiment with GFDL AM2.1 is forced with monthly SST and sea ice of HadISST. The experiment consists of 9 members for 1950-2010 and the ensemble and annual mean is shown here. Greenhouse gases, aerosols, ozone concentration, land cover and solar radiation are fixed to 1990 climatological values.

The observed surface wind used here is the wave and anemometer-based sea-surface wind (WASWind) dataset (Tokinaga and Xie 2011b). This new dataset combines ship observations of surface wind velocity and wind wave height archived in the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) to correct the spurious upward trend due to increases in anemometer height. The dataset is at a monthly resolution of 4° by 4° from 1950 to 2009. It substantially reduces the upward trend in wind speed through height-correction for anemometer-measured winds, rejection of spurious Beaufort winds, and use of estimated winds from wind wave height. It has been utilized for climate trend analysis (Tokinaga et al. 2012a).

Figure 7.5 shows the results. The MASC forcing patterns (Fig. 7.5a) are identical to those in Figure 4.1a, with cooling in the convective regions and warming in the subtropics. In temporal evolution (Fig. 7.5b), the long-term weakening of the Walker circulation is observed in the WASWind, consistent with the MASC forcing. However, the AMIP wind is much weaker than the WASWind during 1950-1975, possibly due to the bias in patterns of SST forcing (Tokinaga et al. 2012a, 2012b). The interannual variability of the two wind data agrees qualitatively and is inconsistent and much stronger than that in MASC, providing evidence for the scale analysis in Section 4.1.2 that MASC is much weaker than SST pattern effect in interannual variability. In both wind datasets, the Walker circulation strengthens
during the last two decades, but this is irrelevant to the MASC forcing, indicating the importance of other mechanisms, such as the decadal variation of the SST patterns.

7.3 Feedback processes stabilizing SST warming

Global warming involves climate feedback processes determining SST and associated circulation changes. Detailed examination of these feedbacks is necessary for better understanding the magnitude and form of such changes, in both global mean and regional patterns. Namely, positive feedbacks include the water vapor feedback enhancing the global SST warming, the WES feedback contributing to the inter-hemispheric gradient, and the Bjerknes feedback forming the eastern equatorial Pacific warming peak. Negative feedbacks include the Newtonian cooling in surface latent heating and longwave radiation.

Specifically over the equatorial cold tongue and in the subtropics, low-cloud feedbacks are important. As SST warms, water vapor increases and the atmosphere stabilizes, which increases the low-cloud amount in these regions. This triggers both longwave and shortwave feedbacks. The increased downward longwave cloud radiation enhances SST warming but reduced shortwave radiation cools the surface. Thus, whether the total low-cloud feedback is positive or negative depends on the warming/cooling rates in specific regions. This open question needs further investigation.

7.4 Changes of sea level and ocean circulation

SST patterns and associated surface wind changes may have significant influence on regional patterns of sea level and ocean circulation changes (Timmermann et al.)
2010). For instance, reduced SST warming in the South Pacific than on the equator and in the North Pacific triggers the WES feedback that enhances trades in the South Pacific. The resulted wind curl generates Ekman pumping/suction in the southwestern/northeastern South Pacific, and hence anomalous sea level rising/sinking. The geostrophic flow forced by this pattern of sea level change would probably be along the wind direction. The regime of coupled response from SST change patterns to atmospheric feedback and hence dynamical ocean response needs to be investigated in future study.
Fig. 7.1. Ensemble-mean biases of climatological SST (contours, CI 0.5 K; zero omitted) and precipitation (color shading, mm day$^{-1}$) between 1pctCO2 and AMIP experiments in CMIP5.
Fig. 7.2. Comparison between ensemble-mean precipitation change based on observational SST and biases in the coupled models in CMIP5. (a) Rainfall change in AMIPFuture run. (b) Biases in rainfall change predicted by the linear regression (Eq. 7.1). (c) Difference of rainfall change between 1pctCO2 and AMIPFuture experiments.
Fig. 7.3. (a) Horizontal distribution of the difference in 300-850 hPa averaged MASC forcing term (Eq. 4.3, in 0.1 K day\(^{-1}\)) between SUSI and A1B runs with AM/CM2.1, along with the (b) zonal mean, and (c) equatorial mean (5°S-5°N).
Fig. 7.4. Difference in atmospheric response to the MASC forcing between SUSI and A1B calculated in LBM. (a) 300-850 hPa averaged air temperature warming patterns (color shading, K), wind shear (vectors, m s$^{-1}$), and absolute zonal wind shear (contours, CI 0.5 m s$^{-1}$; zero omitted). (b) 250-hPa velocity potential (color shading, $10^5$ m$^2$ s$^{-1}$) with divergent wind (vectors, m s$^{-1}$). Contours (CI $2\times10^5$ m$^2$ s$^{-1}$; zero omitted) show the mean velocity potential for reference. (c) Zonal-integrated meridional stream function (color shading, $10^{10}$ kg s$^{-1}$) with the contours (CI $2\times10^{10}$ kg s$^{-1}$; zero omitted) showing the mean circulation for reference.
Fig. 7.5. EOF (a) and PC (b; black) of the MASC forcing in AMIP experiment with GFDL AM2.1. The PC is compared with the Walker circulation index in the AMIP experiment (blue) and observations (red) with 9 years running mean.
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