Formation of Precipitation-Circulation-Sea Surface Temperature Patterns under Global Warming

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF
THE UNIVERSITY OF HAWAIʻI AT MĀNOA
IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE
DEGREE OF
DOCTOR OF PHILOSOPHY
IN
METEOROLOGY
May 2015

By Lei Zhang

Dissertation Committee:
Tim Li, Chairperson
Fei-Fei Jin
Niklas Schneider
Duane Stevens
Bin Wang
Acknowledgements

I would like to thank my advisor, Professor Tim Li, for his great mentoring during these years. Also thanks to my committee for their patience and guidance, and the members of my research group for their helpful discussions. I want to give my special thanks to my wife, who has helped me in so many ways.

I acknowledge the World Climate Research Program’s Working Group on Coupled Modeling, which is responsible for CMIP, and I thank the climate modeling groups for producing and making available their model outputs.
Abstract

Most of CMIP5 models projected a weakened Walker Circulation and an El Nino like warming in the tropical Pacific, but what causes such a change is still an open question. We hypothesize that the following three mechanisms are responsible.

The first mechanism is a so called “longwave radiative – evaporative damping” mechanism. A simple analytical model was constructed to understand the formation mechanisms of future SST warming pattern under global warming. It is demonstrated that the future SST warming pattern is primarily determined by present-day mean SST and surface latent heat flux ($Q_{th}$) fields through a longwave radiative – evaporative damping mechanism. Assume a local thermodynamic equilibrium limit without circulation and cloud changes, a uniform GHG forcing would lead to a smaller (larger) warming in the regions where mean SST and $Q_{th}$ are large (small). This longwave radiative – evaporative damping mechanism can explain a much greater warming in high latitudes than in tropics, and an El Nino like warming in tropical Pacific. In addition, cloud decreases (increases) in eastern (central) Pacific due to weakened Walker circulation (warmer SST), which also contributes to the El Nino like warming.

The second mechanism is “the richest get richer”. In response to a uniform surface warming, the Asia-western North Pacific (WNP) monsoon system is enhanced by competing moisture with other large-scale ascending systems. The strengthened WNP monsoon induces surface westerlies in the western-central equatorial Pacific, weakening the Walker circulation. Idealized AGCM experiments that separate the effect of the mean warming and the relative SST warming pattern clearly demonstrate the effect of the mean warming on change of the equatorial wind.
The third mechanism is extra land surface warming, that is, the land obtains a larger warming than the ocean. In particular, a great thermal contrast between American continent and tropical Pacific causes a zonal pressure gradient and westerly anomalies in the eastern equatorial Pacific. This weakens the Walker circulation.

A traditional view of weakened Walker circulation is attributed to a slower increase rate of global mean precipitation than moisture. However, by analyzing a uniform warming experiment in an aqua-planet setting, it is demonstrated that the Walker circulation is strengthened, even though global averaged upward motion is weakened. This result suggests that the global moisture budget argument may not be sufficient to explain the change of the Walker circulation in the tropics. By conducting numerical simulations in a realistic land-ocean distribution, we demonstrated that the weakening of the Walker Circulation is attributed to both the monsoon and land forcing effects. The relative SST warming pattern also plays a role, but it is just an amplifier, not a fundamental cause.
# Contents

Acknowledgements............................................................................................................. i
Abstract.................................................................................................................................. ii
List of Tables.............................................................................................................................. vi
List of Figures............................................................................................................................. vii

Chapter 1. Introduction.............................................................................................................. 1
  1.1. The Weakening of Walker circulation........................................................................... 1
    1.1.1. Enhanced Static Stability ....................................................................................... 3
    1.1.2. Hydrological Cycle Constraint ............................................................................... 3
  1.2. Formation of SST Warming Pattern under Global Warming...................................... 5
    1.2.1. Ocean Thermostat Theory....................................................................................... 5
    1.2.2. Evaporative Damping Mechanism......................................................................... 6
  1.3. Precipitation Change under Global Warming............................................................... 7
    1.3.1. Wet-get-wetter Theory........................................................................................... 7
    1.3.2. Warmer-get-wetter Theory.................................................................................... 8
  1.4. Objectives and Approaches........................................................................................... 9

Chapter 2. An analytical model for understanding formation of SST warming patterns under global warming – a longwave radiative-evaporative damping mechanism................................................................. 12
  2.1. Introduction.................................................................................................................... 13
  2.2. Model description and methodology .......................................................................... 14
  2.3. Meridional distribution of zonal mean warming pattern ............................................. 18
  2.4. The SST warming pattern along the equator .............................................................. 20
  2.5. Conclusion..................................................................................................................... 22

Chapter 3. Effects of mean warming and extra land warming in causing the weakened Walker circulation under global warming......................................................................................... 32
  3.1. Introduction.................................................................................................................... 32
3.2. Relative roles of mean SST warming, SST warming pattern and land surface warming in determining the Walker circulation changes under global warming ................................. 34
  3.2.1. Model and Methodology ........................................................................................................ 34
  3.2.2. Annual mean trade wind changes .......................................................................................... 36
  3.2.3. Weakened Trade Wind in Boreal Summer ............................................................................ 37
  3.2.4. Weakened Trade Wind in boreal winter .................................................................................. 38
3.3. Relative roles of different surface warming patterns in determining precipitation changes .... 39
3.4. Strengthening of Walker circulation in Aqua-Planet simulation ................................................. 40
  3.4.1. Model design .......................................................................................................................... 41
  3.4.2. Changes of the Walker circulation in Aqua-Planet simulation .............................................. 42
3.5. Conclusion .................................................................................................................................. 45

Chapter 4. Conclusions .................................................................................................................... 61
  4.1. Key points .................................................................................................................................. 61
  4.2. Summary and discussion ........................................................................................................... 61
  4.3. Future work .............................................................................................................................. 66

Bibliography ...................................................................................................................................... 72
List of Tables

Table 2.1 17 CMIP5 climate models used in this study. Monthly mean outputs are analyzed for each model, including SST, horizontal winds, surface heat fluxes, specific humidity, surface wind speed, air temperature and total cloud cover. .................................................................24

Table 3.1 Climate models from CMIP5 for analysis. Monthly horizontal winds, air temperature, specific humidity, relative humidity, precipitation, surface evaporation and SST are analyzed. ..................48

Table 3.2(a) The SLP differences between 60°W-0°W and 120°E-180°E averaged over three latitudinal bands in PD, GW and their difference (units: Pa). (b) Same as in (a) except for SLP differences between 70°W-0°W and 110°E-180°E. ..................................................................................................................49

Table 3.3 The percentage change (GW-PD/PD) of intensity of the Walker Circulation (unit: %) calculated based on the original definition of circulation at different longitude-latitude domain from 925hPa to 150hPa. Shown is the result for ensemble mean of convective parameterization schemes “YS”, “AS” and “KF”. ..................................................................................................................49

Table 3.4 The percentage change rate (GW-PD/PD) of global averaged column integrated moisture, upward motion, precipitation rate, net downward radiation at surface (including both longwave and shortwave radiation), and net radiative cooling in the atmosphere. .................................................................50
List of Figures

Figure 1.1 (a) Differences of surface temperature (shading, K) and 850hPa wind (vector, m/s) between PD and GW in CMIP5 models. (b) for precipitation (shading, mm/day) and 500hPa omega changes (contour, Pa/s). .................................................................................................................................................................. 10

Figure 1.2 Differences of SLP (shading, hPa) and surface temperature fields with 2K warming removed (contour, K) between uniform warming experiment and CTRL in (a) boreal summer and (b) boreal winter. (c) and (d) shows changes of 850hPa winds (vector, m/s) and precipitation (shading, mm/day) in JJA and DJF, respectively. .......................................................................................................................................................... 11

Figure 2.1 (a) Ensemble mean of SST change patterns (K) in 17 CMIP5 models. Shown is global warming minus present day value. (b) Black line is the zonal mean SST changes, and shading stands for the uncertainty among different models. (c) Black line is the SST changes averaged between 5°S and 5°N, and shading the uncertainty of the zonal SST changes among 17 models (one standard deviation across the models)................................................................................................................................. 25

Figure 2.2 SST warming pattern (K) predicted by simple analytical model (Eq. (2.8)) with constant forcing in the numerator (12 W/m²). ...................................................................................................................................................... 26

Figure 2.3 (a) Zonal mean SST warming profile derived from CMIP5 ensemble (δT_s, black solid line) and predicted by a simple analytical model (Eq. (2.8)) (δT_s', red dotted line). Shading is the uncertainty of δT_s derived from 17 CMIP5 models. (b) δT_s' and contributions from each of five terms in the numerator (at Eq. (2.8)) divided by the denominator including downward longwave radiation forcing δQ_{lw} (green), sensible heat flux δQ_{sh} (blue) and latent heat flux change δQ_{lh} due to change of atmospheric state (red), oceanic temperature advection change δDo' (orange), and net solar radiation change δQ_{s0} (red). (c) Same as (b) but only for terms in the numerator. (d) Meridional profiles of three terms in the denominator of Eq. (2.8) and the sum of them (black dashed line). Green is associated with distribution of present-day mean SST (4σT³), blue surface wind speed (γ < V) and yellow surface latent heat flux (γ > Q_{05}). .................................................................................................................................................. 27

Figure 2.4 (a) Zonal distribution of logarithm of the surface latent heat flux in the present-day climate averaged between 5°S and 5°N (red) and the contribution to the heat flux zonal profile by surface wind speed (blue) and air-sea specific humidity difference (green). (b) Same as (a) but for zonal mean latent heat flux distribution. ........................................................................................................................................... 28

Figure 2.5 Green and red line denotes distribution of the change of zonal mean clear sky downward longwave radiation and net solar radiation (GW-PD), respectively, units are W/m²................................................. 28

Figure 2.6 Dash line stands for vertical profile of area averaged air temperature (units, K) in the southern high-latitude region (60°S-75°S, zonal mean) and solid line for the northern high-latitude region.
Figure 2.7 Red bars denote the area-weighted global mean SST change and the contributions from five terms in Eq. (2.8). Whiskers stand for spread among different CMIP5 models.

Figure 2.8 Same as in Fig. 2.3 but for the equatorial region averaged between 5°S and 5°N.

Figure 2.9 Ensemble mean cloud cover change (units: %) in (a) and difference between air temperature change at 850hPa and 1000hPa \((T_{a,850hPa}-T_{a,1000hPa})\) (K) in (b) averaged within 5°N and 5°S.

Figure 3.1 Differences of surface temperature (units, K; 1.55K mean tropical warming between 30°N and 30°S is removed) and 850hPa wind (vector, m/s) between RCP4.5 and historical run in boreal summer (a) and winter (b), respectively. (c) 850hPa zonal wind changes averaged between 5°N and 5°S (units, m/s) in boreal summer (red), winter (blue) and annual mean changes (black). Arrows denote areas where westerly anomaly are significant during different seasons.

Figure 3.2 (a) annual mean precipitation (mm/day) in CTRL with (contour) or without prescribed land surface temperature (shading); (b) and (c) are same as in (a), but for boreal summer and winter, respectively.

Figure 3.3 (a) differences of annual mean precipitation (shading, mm/day) and 850hPa wind (vector, m/s) between DW and CTRL, contour for SST differences with tropical mean warming between 30°N and 30°S removed; (b) as in (a), but for the differences between MW and CTRL, contour for precipitation pattern in CTRL (5mm/day interval); (c) differences between LW and CTRL. (d) differences of 850hPa zonal wind averaged between 5°N and 5°S (m/s) in DW (green), MW (blue) and LW (red) compared to CTRL. Black is the sum of the three curves.

Figure 3.4 differences of annual mean SLP fields (hPa) in (a) DW, (b) MW and (c) LW compared to CTRL.

Figure 3.5 as in Figure 3.3, but for the differences in boreal summer (JJA).

Figure 3.6 as in Figure 3.4, but for the differences in boreal summer (JJA).

Figure 3.7 as in Figure 3.3, but for boreal winter (DJF).

Figure 3.8 as in Figure 3.4, but for boreal winter (DJF).

Figure 3.9 Differences of area averaged 850hPa zonal wind changes in DW (green), MW (blue) and LW (red) compared to CTRL. Domain 165°W-90°W, 5°N-5°S is chosen for annual mean changes. 150°E-90°W, 5°N-5°S for boreal summer and 165°W-90°W, 5°N-5°S for changes in boreal winter.

Figure 3.10 Annual mean precipitation changes in (a) “total forcing”, (b) SST’ and (c) “mean warming and land warming” (units: mm/day). Contours denote (b) SST differential warming (K, 0.2K interval) and (c) rainfall fields (mm/day, 5mm/day interval) in CTRL, respectively. Boxes locate at tropical Pacific (120°E-160°W, 5°N-5°S).

Figure 3.11 Horizontal pattern of a prescribed SST field in the PD simulation. The SST field is symmetric about the equator. It peaks at the equator and decreases poleward. Along the equator, highest SST (31°C) is located at 120°E and lowest SST (27°C) is located at 60°W.

Figure 3.12 Zonal-vertical cross sections of zonal overturning circulation (vector) and vertical p-velocity
(shaded, unit: Pa/s) averaged between 2°S and 2°N derived from the ensemble average of the “YS”, “AS” and “KF” simulations. (a) PD simulation, (b) GW simulation, (c) GW-PD. (d) zonal distribution of SST (dashed) and precipitation (solid) in the PD simulation. .................................................................58

Figure 3.13 (a) The vertical profile of p-velocity (red, unit: Pa/s) and static stability (green, unit: 0.5 K/hPa) averaged over 110°E-130°E and 2°S-2°N. (b) The vertical profile of Q1 (green, 10^-4 K/s), horizontal temperature advection (blue, 10^-4 K/s) and adiabatic heating (red, 10^-4 K/s) averaged over 110°E-130°E and 2°S-2°N. In both panels, the vertical profiles are derived from the ensemble average of the “YS”, “AS” and “KF” simulations. Solid line denotes GW and dashed line denotes PD. .................................................................59

Figure 3.14 Vertical profiles of zonal mean zonal wind (units: m/s) averaged in (a) 120E-60W, (b) 120E-0W and (c) 120E-30E. (d) Mass weighted column-integrated omega (units: Pa^2 sm^-1) averaged between 2N and 2S. Solid (dashed) curve is for PD (GW) simulations (blue for “YS”, red for “AS” and green for “KF” experiment). .........................................................................................59

Figure 3.15 The vertical p-velocity changes averaged in the upper (100-300hPa, blue, unit: Pa/s) and lower (300-1000hPa, red) troposphere and contributions to the vertical velocity changes by the diabatic heating term (column 2), the static stability term (column 3), and sum of the two terms (column 4). All values are averaged over (110°E-130°E and 2°S-2°N) based on the ensemble average of the “YS”, “AS” and “KF” simulations. Whiskers stand for maximum and minimum values for each term in “YS”, “AS” and “KF” simulations. .................................................................................................60

Figure 4.1 Schematic diagram that illustrates the fundamental processes that give rise to the weakening of the Walker circulation/El Nino like SST warming over tropical Pacific under global warming. .......70

Figure 4.2 (a) Evolution of ensemble mean, global mean surface temperature (units, K), with standard deviation across the models (shading, K). (b) Ensemble mean SST trend during hiatus events, (shading, K/decade). Stippled areas are regions where SST trend in at least 110 events (80% of total 137 events) agree with ensemble mean SST trend ........................................................................................................71
Chapter 1

Introduction

The earth is warming due to the effect of greenhouse gases (GHG) under the influence of anthropogenic activities since industrial revolution, with an increasing rate of global mean surface temperature around 0.1K~0.15K/decade based on Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment report (AR4). Evident changes of the climate system have occurred and will continue to take place based on the future projections of climate models, e.g. sea-level rise, ice melting, changes of regional precipitation pattern, and tropical cyclone activities. The impacts of global warming thus draw lots of attention and have been studied elaborately (Manabe et al., 1991, 1992; Church et al. 2001; Allen and Ingram, 2002; Chou and Neelin, 2004; Held and Soden, 2006; Vecchi and Soden, 2007; Li et al. 2010; Seager et al., 2010; Xie et al., 2010; Hsu et al. 2011; Ma et al., 2012; Murakami et al., 2012; Zhao and Held, 2010, 2012). But the influences of global warming have not been fully understood yet, and more studies are still in demand to achieve more comprehensive understanding of changing conditions of climate system.

1.1. The Weakening of Walker circulation

As the major energy source of the climate system, the changes of coupled atmosphere-ocean system in the
tropics have been widely studied (e.g. Chou et al. 1999; Wang et al. 2000; Xie et al. 2010; Hsu et al. 2012; Murakami et al. 2012). It has been shown that anomalous Walker circulation can influence weather and climate systems globally during El Nino and Southern Oscillation (ENSO) events (e.g. Wang et al. 2000; Wu et al. 2009). How does the Walker circulation change under global warming and what are the causes of such change are of vital importance and being intensively examined (Knutson and Manabe 1995; Clement et al. 1996; Cane et al. 1997; Vecchi et al. 2006; Held and Soden 2006; Vecchi and Soden 2007; Dinezio et al. 2009; Collins et al. 2010; Dinezio et al. 2010; Schneider et al. 2010; Merrifield, 2011). Future projections by most climate models that participate coupled model inter-comparison project 5 (CMIP5) capture a weakened Walker circulation under global warming, with anomalous low-level westerly over tropical Pacific (Fig. 1.1a). Since tropical Pacific is a fully coupled system (Bjerknes 1969), it is not surprising that the multi-model ensemble-mean results exhibit weaker east-west sea surface temperature (SST) gradient in the equatorial Pacific as well (Fig. 1.1a). Consistently, the convection center in the warm pool region shifts eastward and anomalous subsidence occupies the Maritime continent (Fig. 1.1b).

In addition, it is also found that the trade wind in equatorial Pacific becomes weakened even in a uniform warming experiment. Shown in Fig. 1.2 are differences between a 2K uniform SST warming experiment and control run (CTRL). The numerical experiment is conducted using an atmospheric general circulation model (AGCM) ECHAM4.6 (Roeckner et al. 1996). The details of this experiment are described in Chapter 3. It is noted that when the ocean surface is uniformly warmed, low-level westerly anomalies show up over the equatorial Pacific in both summer and winter seasons, which suggests weakening of the Walker circulation (Fig. 1.2). Thus one question arises: what causes the changes of the Walker circulation with no changes in SST gradients? As stated below, this weakening of the Walker circulation in a warmer climate has been attributed to the enhanced static stability (Knutson and Manabe 1995) or changes in hydrological cycle (Held and Soden 2006) in previous studies, and both theories are independent of SST warming pattern. However, it is illustrated
in Chapter 3 that these theories may not be sufficient to explain the changes of the Walker circulation under global warming.

1.1.1. Enhanced Static Stability

Knutson and Manabe (1995) analyzed the responses of the climate system to increased CO₂ concentration in the coupled climate model. As suggested in their study, tropospheric static stability becomes greater due to GHG effect. In the tropics, because of small horizontal temperature gradient, diabatic heating associated with deep convection is primarily balanced by adiabatic cooling associated with ascending motion in the troposphere in either present-day (PD) or global warming (GW) equilibrium states. Therefore a more stable state may result in the weakening of atmospheric circulation, which may further lead to the weakening of the Walker circulation. The upper-tropospheric warming in the tropics mainly results from deeper convection in a warmer climate, which is in turn due to greater moisture release from the surface owing to surface warming. However, it is also noted that with more moisture supplied to the air parcel, the latent heating in the troposphere increases as well, which could lead to a more energetic flow. It is the net effect of changes in both diabatic heating and static stability that determines the change of vertical motion, and thus the zonal overturning circulation, i.e. the Walker circulation.

1.1.2. Hydrological Cycle Constraint

Held and Soden (2006) also examined the changes of the Walker circulation in the future projections of CMIP5 climate models, focusing on the responses of hydrological cycle under global warming. It is argued that with negligible changes in relative humidity, moisture in the troposphere increases at the rate around 6%~7%
per degree surface warming (Clausius-Clapeyron equation), which is indeed found in CMIP5 models. In contrast, the simulated global mean precipitation increases at a much slower rate (1%~2%). This indicates that global mean upward mass flux becomes weakened, which further suggests that atmospheric overturning circulation must become weakened under global warming. The cause of the rainfall-moisture “mismatch” is attributed to the global atmospheric energy balance constraint that precipitation or latent heating increase is to a large extent balanced by the change of radiative cooling, which depends on the square root of moisture in air column (Shine et al. 1990; Vecchi and Soden 2007). However, it should be noted that the hydrological constraint is essentially derived from global averaged moisture budget. Whether such argument is sufficient to explain the weakening of Walker circulation, a zonal overturning circulation confined in the tropical region, needs to be examined. In addition, the diagnosis relationship among changes of global mean precipitation, upward mass flux and moisture only holds when the area occupied by the upward motion does not change under global warming, which may not be always true in climate model projections.

It is important to note that these previous studies focus on the constraints from atmospheric point of view, and independent of SST gradient changes. However, by analyzing a uniform SST warming simulation with aqua-planet setting in Chapter 3, it is illustrated that the Walker circulation is actually strengthened, with similar hydrological constraint and enhanced static stability in the troposphere. Such result suggests the aforementioned mechanisms may not be sufficient to explain the weakening of the Walker circulation under global warming. It is thus speculated that the major factors that affect the equatorial overturning circulation must involve realistic land-ocean distribution: i.e. how it interacts with the monsoon heat sources over South Asia and western Pacific warm pools and how unequal land and ocean warming alters zonal pressure gradients. The relative importance of these factors is further illustrated in Chapter 3 using numerical experiments.
1.2. Formation of SST Warming Pattern under Global Warming

Although it is shown that the Walker circulation is weakened in climate models with no changes in SST gradients, understanding of the formation of future SST warming pattern is still of vital importance, since the unequal SST warming modulates the atmospheric circulation/precipitation changes substantially through air-sea interaction. Indeed, it is shown in Chapter 3 that relaxed trade wind over equatorial Pacific in boreal winter is primarily attributed to the effect of SST differential warming. A large portion of projected precipitation changes under global warming is linked to circulation changes induced by anomalous SST gradient as well (Chapter 3).

Lots of theories about the formation of future SST warming patterns have been put forward, many of which focus on whether the SST differential warming over equatorial Pacific resembles an El Nino-like or La Nina-like warming pattern.

1.2.1. Ocean Thermostat Theory

Clement et al. (1996) presented the “ocean thermostat” mechanism, which predicts a stronger zonal SST gradient in tropical Pacific (La Nina-like warming pattern) with uniformly increased radiative forcing imposed to the ocean surface. The argument is as follows: assume uniform SST warming and no changes in surface wind initially, the stratification in the upper ocean is enhanced because of the surface warming, which leads to stronger dynamical cooling effect associated with the mean oceanic upwelling. This cooling effect is only evident in eastern Pacific (EP) where the mean upwelling is strong, hence the east-west SST gradient across the equatorial Pacific is enhanced. Changes in SST gradients further drive stronger trade wind, which further leads to stronger upwelling in EP (stronger poleward Ekman transport at ocean surface). This positive feedback may eventually lead to stronger Walker circulation. However, the warmed surface water in the subtropical region can be transported to the equatorial subsurface ocean through oceanic meridional circulation in tropical Pacific. As
a result, the vertical temperature gradient change in upper ocean is reduced and “ocean thermostat” mechanism is weakened. It should also be noted that the atmospheric component of ZC model is rather simple, and surface heat flux is simply a Newtonian cooling term. As shown in Chapter 2, different cloud feedbacks in western and eastern equatorial Pacific due to different cloud regimes are important for SST gradient changes in tropical Pacific through influencing surface solar radiation, which is clearly not included in ZC model.

1.2.2. Evaporative Damping Mechanism

The weakened east-west SST gradient in equatorial Pacific has been attributed to evaporative damping mechanism (Knuston and Manabe 1995). It is argued that with uniform SST warming and no changes in surface wind initially, increase of surface latent heat flux would be greater (smaller) in western (eastern) equatorial Pacific where mean state SST is relatively higher (lower), due to the exponential dependence of surface moisture on SST (Clausius-Clapeyron relation). Such differential evaporative cooling effect may lead to weaker zonal SST gradient, and thus weaker trade wind. Both changes may amplify through air-sea interaction.

Xie et al. (2010) argued that future SST warming pattern is primarily determined by the surface evaporation field in PD mean state, which is in turn associated with the distribution of mean surface wind field. The argument is that greater increase in evaporation would occur in the regions where mean surface wind speed is large, given same SST warming initially. This differential cooling effect may result in relatively less (more) SST warming in the region where surface wind is large (small). Indeed, the projected equatorial warming is greater than that in the subtropical region, consistent with the observation that the surface wind speed is greater away from equatorial region. However, the dependence of SST warming on mean-state surface wind profile fails to explain the El Nino-like warming pattern in the equatorial Pacific, where mean surface wind is stronger in eastern part of the basin.
In Chapter 2, a diagnosed model for SST changes under global warming is developed for better understanding of the formation mechanisms of SST differential warming pattern due to GHG effect associated with anthropogenic activities. It will be demonstrated that SST changes under global warming are, to a large extent, determined by SST and latent heat flux profiles in PD mean state.

### 1.3. Precipitation Change under Global Warming

Because of its important socioeconomic impacts, precipitation change under global warming has also been widely studied, especially the regional rainfall changes in the tropical ocean, where the bulk of precipitation locates (Chou and Neelin 2004; Vecchi and Soden 2007; Murakami and Wang 2010; Seager et al. 2010; Xie et al. 2010; Hsu and Li 2012; Liu et al. 2013; Ma and Xie 2013). Two prevailing theories explaining the basic features of precipitation changes under global warming are “wet-get-wetter” theory and the “warmer-get-wetter” theory.

#### 1.3.1. Wet-get-wetter Theory

The so-called “wet-get-wetter” or “rich-get-richer” mechanism is put forward to explain the rainfall changes in a warmer climate (Chou and Neelin 2004; Held and Soden 2006). It is argued that with negligible changes in relative humidity, moisture gradients in the troposphere increases in a uniformly warmed climate (Clausius-Clapeyron relation). Consequently, moisture convergence (divergence) in the troposphere becomes greater as well, if assume tropospheric flow remains unchanged compared to PD climate. This leads to increase (decrease) of rainfall in the region where it is already wet (dry). It has been pointed out in many previous studies that such simple theory can explain zonal mean precipitation change reasonably well: the equatorial region receives more
rainfall while the subtropical region becomes drier in the future projections of climate models (Vecchi and Soden 2007; Seager et al. 2010; Dinezio et al. 2010; Zahn and Allan 2013).

However, it is also noted that substantial deviations between regional precipitation changes simulated by general circulation models and rainfall anomaly pattern predicted by “wet-get-wetter” theory still exist, and these discrepancies mainly arise from the neglect of atmospheric circulation change. Hsu and Li (2012) point out that anomalous tropospheric flow alters the regional precipitation change pattern substantially. For instance, there are two convection centers over Indian Ocean (IO) in boreal summer, i.e. Indian summer monsoon (ISM) center in north IO and intertropical convergence zone (ITCZ) in south IO. These two centers may compete with each other through local Hadley circulation change under global warming in the numerical simulations. As a result, only the strongest convective branch (ISM) receives more rainfall in a warmer climate, while the ITCZ over south IO is suppressed. Similarly for rainfall changes over tropical Atlantic in boreal winter. This “richest-get-richer” theory suggests that anomalous circulation pattern and regional precipitation change, especially changes of monsoon heat sources, are closely connected. It is also shown in Chapter 3 that the “richest” convection center over tropical Pacific in boreal summer, western north Pacific summer monsoon (WNPSM) becomes “richer” and drives anomalous low-level cyclonic flow under global warming, which is responsible for the occurrence of westerly anomaly, and thus relaxed trade wind over western equatorial Pacific.

1.3.2. Warmer-get-wetter Theory

Xie et al. (2010) argues that the positive rainfall changes are anchored in the region where SST warming exceeds the tropical mean warming magnitude. In a warmer climate, upper-tropospheric warming is distributed rather uniformly in the tropics due to fast wave adjustment. In contrast, the amplitude of SST differential
warming can be as large as that of the mean warming magnitude at surface. As a result, only in the regions where SST warming exceeds (falls behind) the tropical mean warming is the increase of surface moist static energy (MSE) greater (smaller) than MSE changes in the upper troposphere, which eventually results in rainfall increase (decrease). However, such argument predicts precipitation changes based on a given SST warming pattern, while tropical atmosphere and ocean is a fully coupled system and the formation of SST differential warming itself is related to wind/rainfall changes in the atmosphere.

1.4. Objectives and Approaches

Given the great impacts of global warming on the climate system and deficiencies in the current theories about the formation mechanisms of tropical coupled air-sea system in a warmer climate, in my study, I would like to address how the coupled SST-precipitation-circulation pattern is formed under global warming by analyzing outputs from CMIP5 climate models and conducting numerical experiments.

The target is to reveal the fundamental causes of the El Nino like SST warming (Chapter 2) and weakening of Walker circulation under global warming (Chapter 3) in the tropical Pacific.
Figure 1.1 (a) Differences of surface temperature (shading, K) and 850hPa wind (vector, m/s) between PD and GW in CMIP5 models. (b) for precipitation (shading, mm/day) and 500hPa omega changes (contour, Pa/s).
Figure 1.2 Differences of SLP (shading, hPa) and surface temperature fields with 2K warming removed (contour, K) between uniform warming experiment and CTRL in (a) boreal summer and (b) boreal winter. (c) and (d) shows changes of 850hPa winds (vector, m/s) and precipitation (shading, mm/day) in JJA and DJF, respectively.
Chapter 2

An analytical model for understanding formation of SST warming patterns under global warming – a longwave radiative-evaporative damping mechanism

In introduction, it is demonstrated that the Walker circulation is weakened under global warming, along with a El Nino like SST warming pattern in tropical Pacific in CMIP5 models. What are the fundamental causes of El Nino like SST warming/weakened Walker circulation in tropical Pacific under global warming? Since tropical atmosphere and ocean is fully coupled, we develop a strategy to examine the changes step by step. Firstly, consider a uniformly distributed downward longwave radiation forcing due to greenhouse gases (GHGs), how does the SST change?

A robust feature derived from 17 CMIP5 models is a much greater warming in high latitudes than in tropics, an El Nino like warming over tropical Pacific and Atlantic, and a dipole pattern in the IO. Yet the physical mechanism responsible for formation of such warming patterns remains open. In order to reveal the formation mechanisms of the future warming patterns, a simple theoretical model is constructed in the following sections. The result shows that the SST warming pattern is, to a large extent, determined by the present-day climate mean state, and the atmospheric and oceanic feedback also play a role.
2.1. Introduction

Because of the GHG effect, more longwave radiation is trapped within the earth system, leading to a higher global mean surface temperature. Although the distribution of CO$_2$ increase is rather uniform, the SST warming patterns simulated by state-of-the-art coupled atmosphere-ocean models are non-uniform, which have great impact on regional weather and climate projection (e.g., Allen and Ingram 2002; Vecchi et al. 2008; Li et al. 2010; Murakami and Wang 2010; Zhao and Held 2010, 2012). Figure 2.1 shows the ensemble mean of SST warming pattern simulated by 17 models from CMIP5, using RCP4.5. Greatest warming appears over the Arctic Ocean (called polar amplification), and the equatorial warming is slightly larger than that in the subtropics. Along the equator, an El Nino like warming pattern appears in the tropical Pacific and Atlantic, while a dipole like pattern occurs in the tropical IO.

Polar amplification is commonly attributed to sea ice-albedo positive feedback. Under global warming, sea ice cover in the polar region decreases because of higher surface temperature. This leads to smaller albedo of the surface and thus less reflection of downward solar radiation, which warms the ocean further. However, numerical experiments showed that a greater warming in Arctic Ocean happened even without sea ice feedback effects (Alexeev 2003; Alexeev et al. 2005). This motivates us to investigate other mechanisms for polar amplification.

It has been shown that consistent with reduced east-west SST gradient in tropical Pacific, the Walker circulation is weakened in most of climate model projections (Hsu and Li 2012; Liu et al. 2012; Ma et al. 2012; Vecchi et al. 2006; Vecchi and Soden 2007; Xie et al. 2010). The weakening of the Walker circulation has been attributed to a slower rate of increase in global mean precipitation than in moisture (Held and Soden 2006). An alternative argument is that the atmosphere becomes more stable because of a greater warming in the upper
troposphere due to deeper convection, which may weaken the atmospheric ascending motion (Knutson and Manabe 1995). However, in chapter 1, we have demonstrated that land-sea thermal contrast and monsoon effect might be important for the weakened Walker circulation. From an oceanic dynamics point view, Clement et al. (1996) argued that a La Nina like warming pattern would arise due to the vertical advection of a stronger stratification of the upper ocean under global warming.

In this Chapter, by constructing a simple surface heat budget model that utilizes simulation results from 17 CMIP5 models, we will demonstrate that the future warming patterns described in Fig. 2.1b and 2.1c are, to a large extent, determined by the present-day climate mean state, and modulated by atmospheric wind, moisture and temperature changes.

### 2.2. Model description and methodology

17 CMIP5 model outputs were used for the current analysis. The reason to use only 17 model outputs is that variables we intended to analyze (e.g., separate upward and downward longwave radiation) were not available from all models at the time of this analysis. Because of these reasons, we could only analyze 17 CMIP5 models (see Table 3.1 for list of these models). It has been shown by some previous studies (e.g., Vecchi and Soden 2007; Pithan and Mauritsen 2014) that the basic features of SST warming patterns were quite similar among CMIP models. Monthly surface temperature, surface heat fluxes, surface wind, air temperature and specific humidity in the troposphere from the models listed in Table 2.1 are analyzed. Clear sky downward longwave radiation and shortwave radiation at surface are available from 13 out of 17 models listed in Table 2.1 (marked by asterisk), which are also analyzed. All the variables are derived from table “Amon” in CMIP5 output, where surface temperature denotes SST at the open sea. Model outputs from RCP4.5, in which the radiative forcing reaches 4.5 W/m² (equivalent to 650 ppm CO₂ concentration) in 2100 and stabilizes after that, are used to
examine future SST changes under global warming. The present-day climate state is derived from 20-yr historical simulations (1986-2005). The future warming climate state is derived from another 20-yr period (2081-2100). The difference between the two periods stands for the global warming induced change. To reduce the internal noise of the climate system and derive more clearly the forced signal, we take a multi-model ensemble analysis approach. All analyses are carried out using RCP 4.5 outputs in this paper. RCP 8.5 experiment is also examined, and we find that the results are in general similar to those derived from RCP4.5 (figures not shown).

Assume that both the present-day and future climate states are approximately in an equilibrium state, following Xie et al. (2010). Thus, the future change of surface net heat flux ($\delta Q_{net}$) is balanced by the future change of three dimensional oceanic temperature advection ($\delta D_b$):

$$\delta Q_{net} + \delta D_b = 0,$$

(2.1)

where $Q_{net}$ consists of net solar radiation ($Q_{sw}$), upward and downward longwave radiation ($Q_{lw}^{up}$ and $Q_{lw}^{down}$), latent heat flux ($Q_{th}$) and sensible heat flux ($Q_{sh}$) at surface, and $\delta$ denotes the difference between the future and present-day climate state. By Stefan’s law, the change of upward longwave radiation at the ocean surface may be approximately written as

$$\delta Q_{lw}^{up} = 4\sigma T_s^3 \delta T_s,$$

(2.2)

where a bar denotes the present-day mean state, and $T_s$ the sea surface temperature. Here we assume that the emissivity of ocean surface is unity.

The bulk formulas of surface latent and sensible heat fluxes may be written as:

$$Q_{th} = \rho L C_E V (q_s - q_a) = \rho L C_E V (1 - RH e^{-a\Delta T}) q_s$$

(2.3)

$$Q_{sh} = \rho C_H C_p V (T_s - T_a) = \rho C_H C_p V \Delta T$$

(2.4)

where $\rho$ is air density near surface, $C_E$ and $C_H$ the heat exchange coefficient of $Q_{th}$ and $Q_{sh}$ respectively, $C_p$ the specific heat capacity at constant pressure, $V$ the surface wind speed, RH relative humidity, and $\Delta T$
difference between sea surface and surface air temperature (i.e., $T_s - T_a$). In derivation of Eq. (2.3), the Clausius-Clapeyron equation $\ln \left(\frac{q_{as}}{q_s}\right) = -\frac{L_v(T_s-T_a)}{R T_s T_a}$ has been applied, where $q_{as}$ is saturated specific humidity of surface air. $\alpha = \frac{L_v}{R T_s T_a}$, where $L_v$ is the latent heat of condensation and $R$ is the ideal gas constant for water vapor. Thus the change of $Q_{lh}$ with respect to change of SST may be expressed as

$$\delta Q_{lh} = \frac{\partial Q_{lh}}{\partial T_s} \delta T_s$$

(2.5)

Where $\delta Q_{lh}$ denotes the part of future surface latent heat flux change due to the SST change. $\delta Q_{lh} = \delta Q_{lh} - \delta Q_{lh}^{o}$ denotes the part of surface latent heat flux change due directly to the change of atmospheric wind speed, relative humidity and air-sea temperature difference. In (2.5), $\gamma_1$ is the function of present-day SST field, and for simplicity, it takes a zonal mean value and is only function of latitude, which is about $0.06$ K$^{-1}$ in the tropics and $0.08$ K$^{-1}$ in the polar region. Similarly, the future change of $Q_{sh}$ may be decomposed into

$$\delta Q_{sh} = \frac{\partial Q_{sh}}{\partial T_s} \delta T_s$$

(2.6)

and

$$\delta Q_{sh}^{o} = \delta Q_{sh} - \delta Q_{sh}^{o},$$

(2.7)

where $\frac{\partial \Delta T}{\partial T_s}$ is empirically determined based on the linear relationship between $\Delta T$ and $T_s$ in the present-day climate state, and it is assumed to be only function of latitude. $\gamma_2 = \rho C_H C_p \frac{\partial \Delta T}{\partial T_s}$.

By substituting each of flux terms above into Eq. (2.1), one can obtain

$\delta Q_{lw} - 4\sigma T_s^3 \delta T_s + \delta Q_{lw}^{down} = \gamma_1 \delta Q_{lh} - \gamma_2 \delta T_s - \delta Q_{ah} + \gamma_2 V \delta T_s - \delta Q_{ah} + \delta D_o = 0.$

Thus the future change of SST, $\delta T_s$, can be diagnosed by transforming the equation above into

$$\delta T_s = \frac{\delta Q_{lw}^{down} - \delta Q_{a}^{h} - \delta Q_{ah}}{4\sigma T_s^3 + \gamma_1 Q_{lh} + \gamma_2 V}.$$  

(2.8)

Eq. (2.8) states that the future change of SST is determined by five factors in the numerator, the change of net solar radiation (due to the change of cloud and surface albedo), the change of downward longwave radiation (due to the change of cloud, greenhouse gases and air temperature/moisture), the change of surface latent and sensible heat fluxes associated with the change of atmospheric variables such as wind speed and air specific
humidity, and the change of oceanic temperature advection. In addition, it is also determined by the
denominator, which, as shown later, is primarily controlled by the first two terms that are related to the
horizontal distributions of SST and surface latent heat flux in the present-day climate state.

Physically, this simple model states that with a local thermodynamic equilibrium limit, a uniform incoming
longwave radiative forcing due to greenhouse gases is balanced by changes in outgoing longwave radiation and
latent heat flux from the surface. In the region where the present-day SST and $Q_{th}$ are large (small), the
balance is achieved with a smaller (larger) increase of SST. This process may be referred to as a “longwave
radiative - evaporative damping” mechanism. Consider an idealized climate change scenario in which the
atmospheric and oceanic circulations (including clouds as well as temperature gradients) are kept unchanged
and downward longwave radiative effect due to greenhouse gases is uniformly distributed. Under such a
circumstance, the numerator in Eq. (2.8) becomes a constant. Because the denominator changes markedly in the
meridional direction, one would expect distinctive warming between the tropics and the pole.

Shown in Fig. 2.2 is a SST warming pattern derived when a constant downward longwave radiative forcing
($\delta Q_{lw}^{down} = 12 \, \text{W/m}^2$) is specified. Here the value of $\delta Q_{lw}^{down}$ was derived from the ensemble average of 17
CMIP5 models shown in Table 2.1, and all other terms in the numerator of Eq. (2.8) were set to be zero. Note
that many features such as the greatest warming in the polar region, an El Nino like warming pattern in tropical
Pacific, and a greater equatorial than subtropical warming can be explained by this simple analytical model.
The pattern correlation coefficient between Fig. 2.2 and Fig. 2.1a is 0.7, which is significantly different from
zero (exceeds 99% confidence level based on the student’s t test).

Although the two SST warming patterns resemble each other, the magnitudes of the SST changes in Fig. 2.2
and Fig. 2.1a are different, indicating that atmospheric feedback also plays a role in determining the actual
warming amplitude. The other notable deficiency is that a relatively weak warming in North Atlantic and the
southern hemisphere circumpolar region are not captured. As revealed by previous studies (Manabe et al. 1990,
1991), convection in upper ocean is strong in these two regions, which would transport heat from the surface into a deeper layer, and thus cool the ocean surface.

2.3. Meridional distribution of zonal mean warming pattern

Applying Eq. (2.8) to the CMIP5 model data, one can diagnose the specific processes that give rise to the zonal mean SST warming pattern shown in Fig. 2.1b. Figure 2.3a shows the sum of all five terms in Eq. (2.8), whereas Fig. 2.3b shows the contributions from each of the five terms (denoted by a prime to distinguish from the terms in the numerator). The solid black line in Fig. 2.3a is the SST change obtained from the ensemble average of 17 CMIP5 model projections, with shading representing the spread among these models (here spread is defined as one standard deviation across the models). All the models simulate a robust polar warming pattern. The dotted line in Figs. 2.3a and 2.3b is future SST change predicted by Eq. (2.8). The simple analytical model predicts the meridional distribution of zonal mean SST change very well. For example, the SST warming over the Arctic Ocean is about four times greater than that in the tropics. For the same latitude band, say 60-75°N vs. 60-75°S, northern hemisphere warming is greater.

Figure 2.3b indicates that the dominant term that determines the meridional distribution of the SST change is \( \delta Q_{lw}^{\text{down'}} \) (green curve). \( \delta Q_{th}' \) also contributes to the SST warming, but its effect is mainly confined in the tropics. \( \delta Q_{sw}' \) is small over tropical and mid-latitude regions, but somehow contributes to the SST warming in the Arctic Ocean, possibly due to the ice-albedo feedback. The contributions from zonal mean \( \delta Q_{sh}' \) and ocean advection are in general small compared to other terms.

The meridional distribution of each numerator (denominator) term is shown in Fig. 2.3c (2.3d). Note that dominant terms in the denominator are the first two terms. While term \( 4\sigma T_s^3 \) depends on present-day SST distribution, term \( \overline{Q_{th}} \) depends on the meridional distribution of present-day surface latent heat flux. Given that
the ratio of the denominator values between the tropics (15°S-15°N) and the Arctic (75°N-90°N) is 2.5:1, even a uniform greenhouse radiative forcing would lead to a warming in the Arctic about 2.5 times as large as that in the tropics. This indicates that present-day distribution of \(4\sigma T_s^3\) and \(\overline{Q_{lh}}\) is critical in determining the future warming pattern. The sharp meridional gradient of the denominator is mainly attributed to zonal mean \(\overline{Q_{lh}}\) field, due to rapid decrease of sea-air specific humidity difference with latitude (Fig. 2.4). Although the meridional distribution of \(4\sigma T_s^3\) is relatively uniform, the inclusion of this term greatly reduces the ratio of the denominator between the tropics and the Arctic. The ratio would increase to 15:1 without term \(4\sigma T_s^3\).

The analysis above implies that about 62.5% of the meridional warming contrast between the Arctic and the tropics is attributed to the meridional distribution of present-day mean SST and surface latent heat flux fields, since the denominator itself can induce 2.5 times greater warming in the polar region when imposed by uniform forcing, while the actual warming is four times greater in the polar region. This suggests that the present-day climate state is a major factor that regulating the warming pattern. This is consistent with the fact that all CMIP5 models project a similar meridional warming pattern in Fig. 2.1b.

In addition, atmospheric longwave radiative feedback (term \(\delta Q_{lw}^{down}\) in the numerator) further modulates the warming contrast. Figure 2.3c shows that term \(\delta Q_{lw}^{down}\) is not uniformly distributed. It peaks in the Arctic Ocean. The ratio of \(\delta Q_{lw}^{down}\) between the Arctic and the tropics is 1.6:1. This implies that atmospheric feedback can enhance the warming contrast. The cause of the differential \(\delta Q_{lw}^{down}\) forcing is related to cloud changes in a warmer climate. As shown in Fig. 2.5, the distribution of clear sky \(\delta Q_{lw}^{down}\) is much more uniform compared to Fig. 2.3c. Thus the contrast of \(\delta Q_{lw}^{down}\) between tropics and the polar region is primarily attributed to greater cloud amount in the polar region under global warming, which traps more \(Q_{lw}^{down}\) and warms the surface. On the other hand, more cloud also reduces \(Q_{sw}^{down}\), which partly cancels out positive \(\delta Q_{sw}^{up}\) associated with sea-ice changes and leads to relatively small \(\delta Q_{sw}^{net}\) in Fig. 2.3c. While in the clear sky condition, \(\delta Q_{sw}^{down}\) is negligible and the distribution of \(\delta Q_{sw}^{net}\) is dominated by \(\delta Q_{sw}^{up}\) (Figure 2.5).
The polar amplification was primarily attributed to sea ice-albedo feedback in previous studies. Here we show that it is attributed to a much smaller \( \overline{Q_{th}} \) (as well as atmospheric longwave radiative feedback) in the polar region. This result is consistent with previous studies that found a greater polar warming even with no sea ice effect (Alexeev 2003; Alexeev et al. 2005). In addition, a recent study by Pithan and Mauritsen (2014) found that the surface albedo feedback is not the main contributor to the Arctic amplification. Instead, they demonstrate that the warming contrast between high latitude and tropics mainly come from a temperature feedback: more energy is radiated back to space in the tropics compared to high-latitude in a warmer climate because of higher mean state temperature in the region, which is consistent with our longwave radiative-evaporative damping mechanism.

It is also noted that there is a hemispheric asymmetry in the SST warming. For instance, the warming over latitude band (60°N-75°N) is greater than that over (60°S-75°S). The difference is primarily caused by the asymmetry in \( \delta Q_{lw}^{down} \) (Fig. 2.3c and Fig. 2.5). It is noted that background air temperature in the present-day climate is higher over 60°N-75°N than 60°S-75°S (Fig. 2.6). Thus, even with same air temperature change initially, the resultant \( \delta Q_{lw}^{down} \) is greater in the northern high-latitude band than the southern counterpart based on the Stefan’s law. This leads to a greater SST warming in the northern band.

Applying an area integral to Eq. (2.8), one may estimate the relative roles of atmospheric and ocean feedback processes in determining the global mean SST change (Fig. 2.7). Note that the ensemble mean of global mean SST warming is 1.7K, about 65% of which is contributed by \( \delta Q_{lw}^{down'} \).

2.4. The SST warming pattern along the equator

One may apply Eq. (2.8) to the equatorial region (averaged between 5°S and 5°N) to understand the cause of an El Nino-like warming in the Pacific and Atlantic. Figure 2.8a shows the warming patterns at the equator...
derived from the ensemble average of 17 CMIP5 models (solid curve) and the simple heat budget model (dashed curve). An El Nino like pattern in the Pacific and Atlantic and a dipole like pattern in the IO are well captured.

Among all the atmospheric forcing terms, \(\delta Q_{lw}^{down} \) again dominates (Fig. 2.8b). \(\delta Q_{th}^{a} \) also contributes to the equatorial warming, but it acts against SST gradient change in eastern Pacific and Atlantic. \(\delta Q_{sh}^{a} \) is generally small in most regions (Fig. 2.8b). \(\delta D_o \) favors a La Nina warming pattern, which is consistent with the ocean thermostat mechanism (Clement et al., 1996). \(\delta Q_{sw}^{a} \) contributes to an El Nino like warming in both the basins, which is attributed to different cloud responses to surface warming between the east and west of both basins due to different cloud regimes. In the warm pool region, more deep convective clouds would form when the ocean surface becomes warmer (Fig. 2.9a). In the cold tongue region, low stratus clouds are pronounced due to a stable condition in the lower troposphere associated with trade wind inversion (Philander et al. 1996; Li and Philander 1996). Under global warming, even though the troposphere becomes more stable (Fig. 2.9b), the subsidence is weakened in eastern Pacific due to weakened Walker circulation. The competition between the two leads to a small negative cloud change there (Fig. 2.9a). The distribution of clear sky \(\delta Q_{sw}^{net} \) is rather uniform over the equatorial Pacific and Atlantic, which also supports our conclusion that changes of net shortwave radiation in Fig. 2.8b and 2.8c are mainly attributed to cloud changes (figure omitted).

Each term of the numerator and denominator in Eq. (2.8) is shown in Fig. 2.8c and 8d respectively. Given that \(\delta Q_{lw}^{down} \) distributes uniformly over the equatorial Pacific and Atlantic (green curve in Fig. 2.8c), the El Nino like warming pattern is attributed to the longwave radiative-evaporative damping mechanism (black dashed line in Fig. 2.8d). The zonal distribution of the denominator is mainly controlled by the present-day surface latent heat flux (\(Q_{th} \)) pattern, which is small in eastern equatorial Pacific despite of larger surface wind speed there. This again indicates that sea-air specific humidity difference controls the \(Q_{th} \) profile in the Pacific.

Term \(4\sigma T_s^{-3} \) is rather uniform along the equator, and the inclusion of this term in denominator leads to a
weaker warming contrast between east and west of the ocean basins. The result suggests that an El Nino like warming results essentially from the present-day mean $\overline{Q_{th}}$ profile that decreases toward the east in both the Pacific and Atlantic basins.

The formation of the warming pattern in the IO differs from that in the Pacific and Atlantic. Note that in the IO the warming increases toward the west. The westward warming profile is attributed to the effects of $\delta Q_{lw}^{down}$, $\delta Q_{th}^{a}$ and $\delta D_0$ in numerator, while the denominator plays a negative role. While a greater value of $\delta Q_{lw}^{down}$ in western IO is associated with increased local moisture and air temperature, the latter two terms in numerator are related to anomalous low-level easterlies in the equatorial IO in response to the El Nino like warming in the Pacific (Xie et al. 2010).

2.5. Conclusion

By constructing a simple heat budget model, we reveal the cause of future SST warming patterns under global warming. We conclude that the most important factor that regulates the meridional distribution of SST warming is the mean SST and latent heat flux profiles in the present-day climate state. In the region where $\overline{Q_{th}}$ and $4\sigma T_s^3$ are small, it requires a greater ocean warming response to cancel out the imposed longwave radiative forcing. Because of this fundamental feature of the present-day mean climate dependence, all CMIP5 models projected a robust meridional pattern with a much greater warming in the pole than in the tropics. The mean-state induced warming contrast is further amplified by atmospheric longwave radiation feedback associated with cloud changes. The occurrence of an El Nino like warming in the equatorial Pacific and Atlantic arises from a similar process but with weaker amplitude and a smaller signal-to-noise ratio (Fig. 2.7). Different cloud feedback in east and west of the two ocean basins also plays a role. The less robustness of the east-west warming contrast along the equator than the meridional warming contrast among the models can also be
explained by different magnitudes of total denominator contrast. The warming mechanism in the IO, on the other hand, differs markedly from the Pacific counterpart. It is primarily attributed to the remote forcing from the equatorial Pacific through anomalous Walker circulation.

To sum up, results shown in this study suggest that the global SST warming pattern is mainly controlled by present-day climate state, although the atmospheric and oceanic feedback might modulate the warming pattern as well. Such conclusion may add confidence to the projection of future SST pattern derived by climate models.
Table 2.1 17 CMIP5 climate models used in this study. Monthly mean outputs are analyzed for each model, including SST, horizontal winds, surface heat fluxes, specific humidity, surface wind speed, air temperature and total cloud cover.

<table>
<thead>
<tr>
<th>Model</th>
<th>Institute ID</th>
<th>Modeling center (Group)</th>
</tr>
</thead>
<tbody>
<tr>
<td>* CCSM4</td>
<td>NCAR</td>
<td>National Center for Atmospheric Research</td>
</tr>
<tr>
<td>* CESM1-BGC</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CESM1-CAM5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CMCC-CM</td>
<td>CMCC</td>
<td>Centro Euro-Mediterraneo per I Cambiamenti Climatici</td>
</tr>
<tr>
<td>CMCC-CMS</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* CNRM-CM5</td>
<td>CNRM-CERFACS</td>
<td>Centre National de Recherches Meteorologique / Centre Europeen de Recherche et Formation Avancees en Calcul Scientifique</td>
</tr>
<tr>
<td>* CanESM2</td>
<td>CCCMA</td>
<td>Canadian Centre for Climate Modelling and Analysis</td>
</tr>
<tr>
<td>FGOALS-g2</td>
<td>LASG-IAP</td>
<td>LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences</td>
</tr>
<tr>
<td>* IPSL-CM5A-LR</td>
<td>IPSL</td>
<td>Institut Pierre-Simon Laplace</td>
</tr>
<tr>
<td>* IPSL-CM5A-MR</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* IPSL-CM5B-LR</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* MIROC-ESM</td>
<td>MIROC</td>
<td>Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National</td>
</tr>
<tr>
<td>* MIROC-ESM-CHEM</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* MIROC5</td>
<td>MIROC</td>
<td>Atmosphere and Ocean Research Institute (The University of Tokyo), National Institute for Environmental Studies, and Japan Agency for Marine-Earth Science and Technology</td>
</tr>
<tr>
<td>* MPI-ESM-LR</td>
<td>MPI-M</td>
<td>Max Planck Institute for Meteorology</td>
</tr>
<tr>
<td>* MPI-ESM-MR</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* MRI-CGCM3</td>
<td>MRI</td>
<td>Meteorological Research Institute</td>
</tr>
</tbody>
</table>
Figure 2.1 (a) Ensemble mean of SST change patterns (K) in 17 CMIP5 models. Shown is global warming minus present day value. (b) Black line is the zonal mean SST changes, and shading stands for the uncertainty among different models. (c) Black line is the SST changes averaged between 5°S and 5°N, and shading the uncertainty of the zonal SST changes among 17 models (one standard deviation across the models).
Figure 2.2 SST warming pattern (K) predicted by simple analytical model (Eq. (2.8)) with constant forcing in the numerator (12 W/m²).
Figure 2.3 (a) Zonal mean SST warming profile derived from CMIP5 ensemble ($\delta T_s$, black solid line) and predicted by a simple analytical model (Eq. (2.8)) ($\delta T_s'$, red dotted line). Shading is the uncertainty of $\delta T_s'$ derived from 17 CMIP5 models. (b) $\delta T_s'$ and contributions from each of five terms in the numerator (at Eq. (2.8)) divided by the denominator including downward longwave radiation forcing $\delta Q_{lw}'$ (green), sensible heat flux $\delta Q_{sh}'$ and latent heat flux change $\delta Q_{sh}'$ due to change of atmospheric state (blue and yellow), oceanic temperature advection change $\delta D_o'$ (orange), and net solar radiation change $\delta Q_{sw}'$ (red). (c) Same as (b) but only for terms in the numerator. (d) Meridional profiles of three terms in the denominator of Eq. (2.8) and the sum of them (black dashed line). Green is associated with distribution of present-day mean SST ($4\sigma T^3$), blue surface wind speed ($\gamma_2 \overline{\nu}$) and yellow surface latent heat flux ($\gamma_3 \overline{Q_{sh}}$).
Figure 2.4 (a) Zonal distribution of logarithm of the surface latent heat flux in the present-day climate averaged between 5°S and 5°N (red) and the contribution to the heat flux zonal profile by surface wind speed (blue) and air-sea specific humidity difference (green). (b) Same as (a) but for zonal mean latent heat flux distribution.

Figure 2.5 Green and red line denotes distribution of the change of zonal mean clear sky downward longwave radiation and net solar radiation (GW-PD), respectively, units are W/m².
Figure 2.6 Dash line stands for vertical profile of area averaged air temperature (units, K) in the southern high-latitude region (60°S-75°S, zonal mean) and solid line for the northern high-latitude region (60°N-75°N, zonal mean).

Figure 2.7 Red bars denote the area-weighted global mean SST change and the contributions from five terms in Eq. (2.8). Whiskers stand for spread among different CMIP5 models.
Figure 2.8 Same as in Fig. 2.3 but for the equatorial region averaged between 5°S and 5°N.
Figure 2.9 Ensemble mean cloud cover change (units: %) in (a) and difference between air temperature change at 850hPa and 1000hPa \((T_{\text{a,850hPa}} - T_{\text{a,1000hPa}})\) (K) in (b) averaged within 5°N and 5°S.
Chapter 3

Effects of mean warming and extra land warming in causing the weakened Walker circulation under global warming.

In the argument above, we emphasize the local thermodynamic equilibrium limit: with no changes in cloud/wind, we found an unequal SST warming with a uniform downward longwave radiation forcing. This present-day mean state dependence essentially give rise to the El Nino like SST warming in tropical Pacific due to the effects of GHGs. To the first order approximation, SST warming is distributed relatively uniform in the tropical ocean because of uniformly distributed greenhouse gases in the troposphere. Therefore in this chapter, we consider another idealized scenario: assume a uniform SST warming across the global initially, with no changes in SST gradients, how would wind/precipitation change? These two parts are complementary to each other.

3.1. Introduction

The changes of the Walker circulation in a uniform warming experiment are analyzed first. The numerical experiments are conducted using AGCM ECHAM4.6. The model resolution is T42. 2K uniform warming is added to SST compared to the CTRL. For both experiments, model was integrated for 20 years and last 15
years are analyzed. Shown in Fig. 1.2 are 850hPa wind, rainfall, sea level pressure (SLP), and land surface temperature changes with 2K mean warming removed in boreal summer and winter respectively. In both seasons, westerly anomaly shows up over the equatorial Pacific, which suggests that trade wind and thus Walker circulation is weakened. Over western Pacific, westerly anomaly in boreal winter seems to be associated with cyclonic circulation anomalies on both sides of equator, which are further related to enhanced south Pacific convergence zone (SPCZ) and convection center over WNP. While in boreal summer, weakened trade wind over the warm pool region is due to enhanced WNP monsoon trough. Therefore it is speculated changes of monsoon heat sources over equatorial western Pacific is closely related to weakened trade winds in tropical Pacific.

How do monsoon centers respond to GHG effect? One prevailing theory regarding precipitation changes under global warming is the so-called “rich-get richer” (or “wet-get-wetter”) theory (Chou and Neelin, 2004; Held and Soden, 2006), which essentially emphasizes the thermodynamical effect of mean SST warming. However, as revealed by Hsu and Li (2012), not all the rainfall centers becomes stronger even in the future projections of CMIP5 models. Instead, the relatively stronger convection centers become wetter by driving anomalous convergent flow, while the weaker ones are suppressed due to the circulation change (“richest-get richer”). Therefore, the changes of major convection centers go hand in hand with anomalous flow pattern under global warming. They further illustrate that this “richest-get richer” is primarily attributed to mean surface warming. It is speculated that enhanced WNP monsoon trough due to mean SST warming may compete moisture with other ascending centers and lead to a weakening of the Walker circulation.

Another prominent feature of the surface warming pattern in the future projections by the climate models is greater warming over land surface than that over the ocean (Fig. 3.1). Over South America Continent, this additional warming may lead to lower SLP in the region, which can alter the large-scale circulation pattern by affecting the SLP gradients. Such effect might be important for trade wind changes, especially over EP region
adjacent to America continent. It is indeed found that westerly anomalies show up over EP in Fig. 1.2. A recent study by Bayr et al. (2014) also suggests that greater warming over land might be important for the weakening of the Walker circulation in a warmer climate.

3.2. Relative roles of mean SST warming, SST warming pattern and land surface warming in determining the Walker circulation changes under global warming

In addition to the aforementioned mechanisms (mean SST warming and extra land warming), atmospheric response to SST differential warming pattern is also investigated for comparison, since changes in SST gradients can alter tropospheric circulation substantially through air-sea coupling. But it is important to note that such effect is only an amplifier, rather than a fundamental cause. So in this section, the relative roles of mean SST warming (“richest-get-richer”), SST differential warming pattern and land surface warming (excessive land warming) in determining the Walker circulation changes under global warming are examined.

3.2.1. Model and Methodology

To obtain the SST differential warming pattern and determine the magnitudes of mean SST warming and additional land surface warming in the tropics, fifteen CMIP5 coupled climate models are analyzed (Table 3.1). Both historical runs and the representative concentration pathway 4.5 (RCP4.5) experiments are examined, and surface warming pattern is defined as the difference between present-day (1986-2005) and global-warming (2081-2100) climate states. In RCP4.5, the radiative forcing reaches 4.5 W/m² (equivalent to 650ppm CO₂ concentration) in 2100 and stabilizes after that.
The surface temperature warming patterns derived from CMIP5 models in boreal summer and winter are shown in Figs. 3.1a and 3.1b respectively, with 1.5K annual mean tropical mean SST warming between 30°N and 30°S removed. Both warming patterns exhibit greater warming over land than the ocean surface (0.9K additional warming over land in the tropics, 30°N and 30°S), with El Nino like SST warming pattern over tropical Pacific. In the meridional direction, SST anomaly manifests prominent hemispheric warming contrast, which is more evident in boreal summer. Fig. 3.1c shows 850hPa zonal wind changes in the equatorial region. Westerly anomaly occupies the entire equatorial Pacific in summer (June, July, August, JJA), while relaxed trade winds mainly locate in central and EP in winter (December, January, February, DJF). Annual mean trade wind changes over tropical Pacific in Fig. 3.1c is similar to winter season.

The changes of the Walker circulation and low-level trade winds are linked to changes in SST gradients associated with SST differential warming, changes of monsoon heat sources due to mean surface warming and SLP gradients change in EP related to additional land warming. To illustrate the relative importance of different mechanisms, the atmospheric responses to different SST and land surface warming patterns in numerical experiments are analyzed, using AGCM ECHAM 4.6 (Roeckner et. al, 1996). The resolution of the model is T42. For each experiment, the model is integrated for thirty years and last twenty years are analyzed. In addition to CTRL, which is forced by climatological surface temperature patterns over both land and ocean, three comparative experiments are carried out: (1) additional monthly SST differential warming pattern derived from CMIP5 models (Table 3.1) is added to the climatological SST field, while land surface temperature remains the same as in CTRL; (2) 1.55 K warming, which is the annual mean tropical mean SST warming between 30°N and 30°S, is added to both ocean and land surface compared to CTRL. (3) 0.9 K warming is added to the land surface, while SST has no difference from that in CTRL. By comparing the three comparative experiments to CTRL, the effects of SST warming pattern, mean surface warming and land-sea thermal contrast
on driving anomalous flow patterns in tropical Pacific are examined. Hereafter, these three experiments are referred to as DW (differential warming), MW (mean warming) and LW (land warming) respectively.

Shading in Fig. 3.2 are the distributions of precipitation fields with land surface temperature predicted by the model, and contours are for the experiments with prescribed land surface temperature. The land surface temperature in the latter run is derived from the climatological mean state in the previous one. The remarkable similarity between these two experiments validate our approach by nudging the land surface temperature.

3.2.2. Annual mean trade wind changes

The annual mean atmospheric responses to the anomalous surface warming patterns are shown in Fig. 3.3. The most remarkable features of SST differential warming pattern are El Nino-like warming over the equatorial Pacific, greater equatorial warming than the subtropics and hemispheric warming contrast between northern and southern hemispheres. Correspondingly, major rainfall centers shift eastward over the central equatorial Pacific, accompanied by large-scale convergent winds toward equatorial region, while the subtropical region becomes drier. SLP is consistently lower in the equatorial region than the subtropics as well (Fig. 3.4a). Cross-equatorial flow in central and eastern equatorial Pacific agrees with warmer SST and lower SLP in northern hemisphere (Fig. 3.4a). In the zonal direction, low-level westerly shows up in central and eastern equatorial Pacific in Fig. 3.3a, which is in accordance with lower SLP in EP due to El Nino like warming (Fig. 3.4a). Over IO, a dipole-like warming pattern induces low-level southeasterly anomaly in the equatorial region, which further leads to northwest-southeast seesaw-like precipitation change pattern.

The positive rainfall anomaly centers over the ocean due to mean surface warming are generally in agreement with climatological precipitation distribution in the tropical region (Fig. 3.3b). The enhanced rainfall in WNP co-locates with cyclonic gyre anomaly, which leads to westerly anomaly in the Maritime continent
region (Figs. 3.3b and 3.3d). While over central and eastern tropical Pacific, large-scale westerly winds toward America associated with enhanced precipitation over land show up in both northern and southern hemisphere, but away from the equator. Additional land warming drives low-level westerly anomaly over eastern equatorial Pacific through inducing SLP and precipitation anomaly (Figs. 3.3c and 3.4c).

Overall, westerly anomaly mainly locates in eastern equatorial Pacific, which is consistent with annual mean trade wind changes simulated in CMIP5 models (Fig. 3.1c). Such changes are primarily contributed by changes in SST gradient over tropical Pacific (Fig. 3.9), which contributes to around 58% of total westerly anomaly, while mean warming (extra land warming) induced changes lead to 30% and 12% of weakened trade winds. It is noted that although SST differential warming seems to play a dominant role, the formation of SST differential warming itself is already a result of air-sea coupling, therefore it is not a fundamental cause. Instead, the mean SST warming and additional land warming, both of which are robust responses to GHG effect, must play essential roles in the weakening of the Walker circulation.

### 3.2.3. Weakened Trade Wind in Boreal Summer

Changes of wind patterns in boreal summer are shown in Fig. 3.5. Similar to annual mean SST warming pattern, both El-Nino like warming and greater warming over northern hemisphere are evident in JJA (Fig. 3.1a), which leads to low-level westerly anomaly over Maritime continent and southerly cross-equatorial flow over most of equatorial Pacific (Fig. 3.5a and Fig. 3.5d). Over WNP, anomalous anti-cyclonic circulation and negative rainfall occupy Philippine Sea, which might be due to the enhanced Indian summer monsoon (ISM) associated with dipole-like warming over IO. Previous studies suggest that the rainfall change over WNP is dynamically linked with ISM rainfall anomaly, with out-of-phase relationship (Chang and Li, 2000; Wang et al., 2013), which is consistent with the results shown in Fig. 3.5a.
The differences between MW and CTRL in JJA are shown in Fig. 3.5b. Climatologically, the strongest rainfall center locates at WNP over tropical Pacific in boreal summer, owing to the monsoon trough (Fig. 3.5b). When the earth surface becomes uniformly warmer, with no changes in the land-sea thermal contrast, positive precipitation anomaly and anomalous cyclonic flow occupy WNP as well, which is consistent with “richest-get-richer” theory (Fig. 3.5b). Such change leads to westerly anomaly over central and western equatorial Pacific (Fig. 3.5d). Over EP, strong rainfall center over North America in the tropical region is enhanced, while ITCZ is suppressed.

The additional land warming-induced changes in boreal summer are shown in Fig. 3.5c. Rainfall over North America is enhanced due to higher surface temperature over land, which induces lower SLP and drives large scale low-level westerly over EP (Fig. 3.6c and Fig. 3.5d).

Both mean warming effect and additional land warming contribute to weakened Walker circulation in boreal summer under global warming (Fig. 3.5d). While the former leads to westerly anomaly over central and western Pacific, land-sea thermal contrast is responsible for the occurrence of low-level westerly anomaly in EP. SST differential warming only plays a minor role in trade wind changes in JJA (Fig. 3.9).

### 3.2.4. Weakened Trade Wind in boreal winter

The changes of atmospheric state in boreal winter are also analyzed (Figs. 3.7 and 3.8). With a El Nino-like warming over tropical Pacific (Fig. 3.1b), suppressed rainfall in the warm pool region shows up in DW compared to CTRL, while the rainfall is enhanced over central Pacific (Fig. 3.7a). Low-level westerly anomaly occupies central and eastern equatorial Pacific, which contributes to weakened trade winds (Fig. 3.7d). Compared to JJA, the cross-equatorial flow in equatorial Pacific is not as evident, which might be due to less hemispheric warming contrast in boreal winter season (Fig. 3.1b).
Similar to climatological rainfall pattern, the mean warming induced major positive precipitation changes occupy in western Pacific and SPCZ in DJF, while negative rainfall anomaly and anomalous subsidence show up in the equatorial region (Fig. 3.7b). The anomalous cyclonic flow associated with enhanced SPCZ and WNP convection center on both sides of equator show up, but away from equator (Fig. 3.7b). Rainfall increases over America in Fig. 3.7b, with the large-scale convergent flow shifting to the southern hemisphere in DJF. Over south IO, where abundant rainfall occurs, positive precipitation together with westerly anomaly occur to the south of equator as well (Fig. 3.7b).

In LW, westerly anomaly due to warmer land surface and lower SLP over South America still shows up over EP (Figs. 3.7c, 3.7d and 3.8c). Large-scale convergent flows toward Australia occurs over south IO and south Pacific due to anomalous land warming and lower SLP there (Figs. 3.7c and Fig. 3.8c).

The westerly anomaly mainly locates in eastern equatorial Pacific in DW, MW and LW in DJF (Fig. 3.8d), which agrees with low-level zonal wind change found in CMIP5 models (Fig. 3.1c). The anomalous westerlies are primarily attributed to the effect of unequal SST warming (Fig. 3.9).

3.3. Relative roles of different surface warming patterns in determining precipitation changes

In the previous sections, the relative roles of different surface warming patterns in driving low-level zonal wind changes are analyzed. In this section, different factors that contribute to precipitation changes over tropical Pacific are examined. One more experiment which includes the monthly warming patterns over both the ocean and land surface in CMIP5 models are conducted (TOT). Therefore the differences between TOT and CTRL are due to total effects of SST differential warming pattern as well as mean warming and additional land warming, and the results can be compared to precipitation changes simulated in CMIP5 models to further validate the performance of our numerical experiments.
The differences between TOT, which includes the effect of total surface warming derived from CMIP5 models, and CTRL are shown in Fig. 3.10a. Overall, the basic features of precipitation changes in CMIP5 models are well simulated (Fig. 1.1), e.g., the dipole-like precipitation changes over IO and enhanced rainfall over central equatorial Pacific are all found in the numerical experiments.

SST differential warming induces eastward shift of convection center over tropical Pacific, corresponding to El Nino like warming pattern (Fig. 3.10b). Positive (negative) rainfall changes over the tropical (subtropical) regions are consistent with enhanced equatorial warming. The dipole-like rainfall changes over IO in Fig. 3.10b also agree with greater warming over north IO. On the other hand, positive precipitation anomaly in Fig. 3.10c generally follows climatologic rainfall distribution, which is primarily due to mean surface warming effect (“richest-get-richer”).

The remarkable similarity between Fig. 3.10a and Fig. 3.10b suggests a large portion of total annual mean precipitation changes is attributed to SST differential warming effect. But other forcing also plays a role in enhancing rainfall over north IO, central and western Pacific (Fig. 3.10c). The total effect of unequal SST warming account for around 63% of total rainfall changes in equatorial Pacific in the region with most significant precipitation anomaly, while other forcing such as mean surface warming together contribute to around 37%. Such result is consistent with that around 60% of total westerly anomaly is due to SST differential warming. But still, it is noted that such effect is only an amplifier, rather than a fundamental cause.

3.4. Strengthening of Walker circulation in Aqua-Planet simulation

In the previous sections, we illustrate that monsoon and land effects are fundamental causes of the weakened trade winds under global warming. Previous studies have attributed the weakening of the Walker circulation to the changes of hydrological cycle or tropospheric static stability (Held and Soden, 2006; Knuston and Manabe,
1995). However, neither of these arguments concern about land-ocean contrast. Are these theories sufficient to explain the Walker circulation change under global warming in an Aqua-Planet model with no land and monsoons? To eliminate the monsoon and land impacts and examine the “pure” Walker circulation response to global warming, we design a set of idealized numerical experiments in an Aqua-planet Earth to eliminate the monsoon and land impacts and examine the “pure” Walker circulation response to global warming. Through the diagnosis of the idealized simulation outputs, we intend to understand the relative importance of the moisture effect versus the stability effect in affecting the Walker circulation strength.

3.4.1. Model design

A wavenumber-1 zonal SST distribution is specified in the tropics (within 20°S-20°N) to simulate the zonal overturning circulation, namely the Walker circulation. The model used for this part of study is an atmospheric general circulation model developed at Japan Meteorological Agency (JMA) Meteorological Research Institute (MRI). For a detailed description of this model, please refer to Yukimoto et al. (2011) and Mizuta et al. (2012). The model resolution is T106, and each simulation was integrated for 10 years. Solar radiation in this idealized experiment is fixed at the equinox.

Figure 3.11 shows the idealized SST pattern in the PD climate state. The SST is symmetric about the equator, with maximum amplitude right on the equator. Along the equator, highest (lowest) SST is located at 120°E (60°W). In the global warming (GW) climate state, a global uniform 2K warming is imposed and CO₂ concentration is doubled.

One of the greatest uncertainties of the future projection by the climate models is cumulus parameterization (Stocker et al., 2001). To test the sensitivity of the simulation results to model convective parameterization scheme, three parallel runs with the Yukimoto (2011) (hereafter “YS”), Arakawa-Schubert (1974) (hereafter “AS”) and Kain-Fritsch (1990, 1993) (hereafter “KF”) convective parameterization schemes were carried out
for both the PD and GW climate simulations. It has been shown that the MRI model with these schemes can simulate realistic climatological mean states of tropical and monsoon precipitation (Endo et al., 2012) and realistic tropical cyclone distributions (Murakami et al., 2012).

### 3.4.2. Changes of the Walker circulation in Aqua-Planet simulation

In the presence of an idealized wavenumber-1 SST distribution, the Walker Circulation is well simulated in both PD and GW states (Fig. 3.12). Maximum ascending motion and precipitation appear over the warmest pool, and minimum rainfall and descending motion appear around 60°W. The most striking difference between the GW and PD simulations lies in the change of vertical motion profile over the ascending branch of the Walker Circulation around 120°E (Fig. 3.12c). The ascending motion is strengthened (weakened) in the upper (lower) troposphere. Consistent with the vertical velocity change, a westerly anomaly appears in the upper and lower levels, while an easterly anomaly appears in between (Fig. 3.12c). As a result, an anomalous “double-cell” vertical overturning circulation pattern with a clockwise (counter-clockwise) cell in the upper (lower) troposphere forms. The separation line between the upper and lower cells appears at 300hPa.

An important issue is how to measure quantitatively the change of strength of the Walker Circulation? The area-averaged change of vertical motion at the ascending branch of the Walker Circulation is plotted in Fig. 3.13a. The ascending motion is strengthened above 300hPa but weakened below 300hPa, and keeps unchanged at the maximum vertical velocity level (300hPa). Thus the vertical motion over the region cannot be used to determine the change of overall strength of the Walker Circulation. In previous study, the difference of SLP between western and eastern hemisphere has also been applied to measure the strength of Walker circulation (Vecchi et al., 2006). However, such a measurement is very sensitive to domain selected. For example, the change of east-west SLP gradient is negative when two domains 60°W-0°W and 120°E-180°E are used, but
becomes positive when domains 70°W-0°W and 110°E-180°E (in which the domains are only expanded westward by 10° longitude) are used (Table 3.2). This suggests that the zonal SLP gradient (which mainly measures low-level zonal wind) is also not a good indicator.

Given the great uncertainty in both the SLP and vertical motion fields, we decide to use the original definition of circulation (Holton, 2004) to measure the strength of the Walker Circulation. The strength of zonal overturning circulation in the longitude-height cross-section may be calculated according to an area integral of meridional vorticity, \( \int \left( \frac{1}{\rho g^2} \frac{\partial \omega}{\partial x} - \frac{\partial u}{\partial p} \right) dx dp \). This circulation definition contains the combined information of both the vertical motion and the zonal wind in a large longitude-height domain. Figure 3.14 shows that climatologically, maximum westerly (easterly) wind locates at 150hPa (925hPa) in this idealized experiment, and strongest upward motion lies at around 120°E. Considering a longitude-height domain of 120°E-60°W and 925hPa-150hPa, we calculated the percentage change of the intensity of the Walker Circulation in PD and GW simulations averaged over three latitudinal zones (Table 3.3). The result shows that the Walker Circulation is strengthened under global warming in all three convective parameterization sensitivity experiments, regardless of which latitudinal band that we choose. We further test the sensitivity of the result to the longitudinal domain. Given that Kelvin wave response length scale (in response to a given heating in the warm pool) is greater than the Rossby wave response length scale, additional calculations with greater longitudinal domains, 120°E-0°W and 120°E-30°E, were performed (Table 3.3). In the latter case, the Kelvin wave response length scale (270°) is exactly three times as large as the Rossby wave response length scale (90°). The result indicates that the Walker Circulation strengthening signal is robust.

Since the change of zonal overturning circulation in the equatorial plane exhibits a “double-cell” pattern, with clockwise (counter-clockwise) circulation change in the upper (lower) troposphere (Fig. 3.12), we further examine which cell dominates the overall strength of the Walker Circulation. Our calculation shows that the change of strength of the Walker Circulation in the idealized Aqua-Planet model is primarily controlled by the
upper cell circulation change. What causes the strengthening of upper tropospheric vertical motion? In the
tropics where the horizontal temperature gradient is small, for both PD and GW equilibrium state, the adiabatic
cooling associated with vertical motion is approximately in balance with the diabatic heating term, thus
\[
\omega S + Q_1 = 0.
\]  
(3.1)

In equation (3.1), \( S \) denotes atmospheric static stability, \( Q_1 \) represents apparent heating (Yanai et al., 1973)
that includes longwave radiation, condensation heating and divergence of eddy static energy transport. Equation
(3.1) implies that the change of vertical motion under global warming is determined by the combined effect of
the static stability and apparent heating changes. It has been shown that global warming leads to the increase of
both static stability and diabatic heating. Thus it is necessary to reveal their relative roles, in particular how
their relative effects change with height. Figure 3.13 illustrates how the static stability parameter and the
diabatic heating change vertically. Whereas the static stability parameter increases throughout troposphere,
apparent heating exhibits a maximum increase in the upper troposphere, which is consistent with previous study
by Huang et al. (2013). The diagnosis of horizontal temperature advection shows that this term is negligible
(Fig. 3.13b). Therefore equation (3.1) is indeed valid in the region of interest.

To measure quantitatively the relative contributions of the static stability and diabatic heating to vertical
motion change in upper and lower troposphere, we transform equation (3.1) into the following form:
\[
\Delta \omega = -\frac{1}{S} \Delta Q_1 + \frac{Q_1}{S^2} \Delta S
\]  
(3.2)

where \( \Delta \) represents the difference between GW and PD state.

Figure 3.15 shows the diagnosis results for upper and lower tropospheric vertical motion based on equation
(3.2). The effect of diabatic heating change exceeds (falls behind) that of static stability change above (below)
300hPa. In other words, the strengthened upward motion in the upper troposphere is primarily caused by the
diabatic heating effect, whereas the weakened upward motion in the lower troposphere is mainly affected by
enhanced atmospheric static stability. The result indicates that a more stable atmosphere doesn’t necessarily
lead to weakened vertical circulation. It is the net effect of diabatic heating and static stability changes that determine the final sign change of the vertical motion.

While the Walker circulation is strengthened under global warming, global average upward motion is weakened in the Aqua-Planet simulation (Table 3.4). This points out that the global moisture budget argument by Held and Soden (2006) is valid even in an idealized Aqua Planet model. It is found that global average water vapor content in the current model increases by 19% with a 2K warming, whereas global mean precipitation and surface latent heat flux increase by only around 4%. The slower rate of rainfall than moisture increase is consistent with the fact that in the model global mean ascending motion decreases. Therefore, the global moisture budget argument seems to be applicable to changes of global mean upward motion, but may not be sufficient to explain the change of the Walker circulation that is confined in the equatorial region.

The fact that the Walker circulation is strengthened in an Aqua Planet simulation but weakened in the future projections of most climate models participated in the CMIP5 (in which realistic land-sea distributions and monsoon flows are presented) suggest that the weakening of the Walker circulation in a real world might be attributed to the influence of change of the monsoon strength and a greater land warming. Therefore, the major factors that affect the equatorial overturning circulation must involve how it interacts with the monsoon heat sources over South Asia and western Pacific warm pools and how unequal land and ocean warming alters zonal pressure gradients. We will investigate the relative importance of these factors in the following sections.

3.5. Conclusion

Idealized Aqua planet simulations were performed in an aim to understand how the Walker circulation changes in an idealized world with no competition of the monsoon circulation or excessive land warming. A zonal wavenumber-1 SST distribution is prescribed in the tropics in the present-day simulation. It is found that
the Walker Circulation is strengthened under such an idealized world, given a uniform 2K SST warming. A diagnosis shows that the ascending branch of the Walker cell is enhanced in the upper troposphere but weakened in the lower troposphere. As a result, a “double-cell” circulation change pattern forms, with a clockwise (anti-clockwise) circulation anomaly in the upper (lower) troposphere. The upper tropospheric circulation change dominates the strength change of the Walker circulation. The mechanism for the formation of the “double cell” circulation pattern is attributed to a greater (smaller) increase rate of diabatic heating than static stability in the upper (lower) troposphere.

The strengthened Walker circulation in aqua planet simulation is in contrast with weakened Walker circulation in both future projections of CMIP5 models and uniform warming experiment (Figs. 1.1 and 1.2), which suggests that the major factors that affect the equatorial overturning circulation must involve how it interacts with the monsoon centers and additional land warming effect in a realistic land-ocean distribution. Numerical experiments are conducted to examine the causes of the weakening of the Walker circulation under global warming. Shown in Fig. 3.9 are the area averaged zonal wind changes due to SST differential warming pattern, mean surface warming and additional land warming, respectively. The area is chosen based on the domain with most significant westerly anomaly over equatorial Pacific in different seasons in both CMIP5 model simulations (Fig. 3.1c) and numerical experiments carried out in this study (Figs. 3.3d, 3.5d and 3.7d). While in boreal winter and for annual mean changes, relaxed trade winds locate at eastern equatorial Pacific, westerly anomaly occupies nearly the entire equatorial Pacific during boreal summer.

SST differential warming pattern plays a dominant role in driving the anomalous low-level westerly over the equatorial Pacific in boreal winter and for annual mean changes (Fig. 3.9). In JJA, however, weakened trade winds over equatorial Pacific are primarily caused by enhanced WNP summer monsoon and greater rainfall over North America (Figs. 3.5b, 3.5c and 3.9), which are related to “richest-get-richer” theory and additional land warming, respectively.
It is noted in Chapter 2 that the El Nino like SST differential warming pattern over tropical Pacific is attributed to longwave radiative – evaporative damping mechanism associated with PD mean state, which is then amplified by the weakened Walker circulation. Thus the roles of SST differential warming in Walker circulation changes are partly a consequence of air-sea coupling. Furthermore, it is shown that the weakening of the Walker circulation can even be captured in a uniform warming experiment (Fig. 1.2). Therefore the mean surface warming (“richest-get-richer”) and additional land warming effects, both of which are independent of SST gradient changes, must play essential roles in the relaxed trade winds over the equatorial Pacific in a warmer climate.
Table 3.1 Climate models from CMIP5 for analysis. Monthly horizontal winds, air temperature, specific humidity, relative humidity, precipitation, surface evaporation and SST are analyzed.

<table>
<thead>
<tr>
<th>Model</th>
<th>Institute ID</th>
<th>Modeling center (Group)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACCESS1-0</td>
<td>CAWCR</td>
<td>The Centre for Australian Weather and Climate Research</td>
</tr>
<tr>
<td>BCC-CSM1.1</td>
<td>BCC</td>
<td>Beijing Climate Center, China Meteorological Administration</td>
</tr>
<tr>
<td>CanESM2</td>
<td>CCCMA</td>
<td>Canadian Centre for Climate Modelling and Analysis</td>
</tr>
<tr>
<td>CNRM-CM5</td>
<td>CNRM-CERFACS</td>
<td>Centre National de Recherches Meteorologiques / Centre Europeen de Recherche et Formation Avancees en Calcul Scientifique</td>
</tr>
<tr>
<td>GFDL-CM3</td>
<td>NOAA GFDL</td>
<td>NOAA Geophysical Fluid Dynamics Laboratory</td>
</tr>
<tr>
<td>GFDL-ESM2M</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GISS-E2-R</td>
<td>NASA GISS</td>
<td>NASA Goddard Institute for Space Studies</td>
</tr>
<tr>
<td>HadGEM2-CC</td>
<td>MOHC</td>
<td>Met Office Hadley Centre</td>
</tr>
<tr>
<td>HadGEM2-ES</td>
<td></td>
<td></td>
</tr>
<tr>
<td>IPSL-CM5A-LR</td>
<td>IPSL</td>
<td>Institut Pierre-Simon Laplace</td>
</tr>
<tr>
<td>IPSL-CM5A-MR</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MIROC-ESM</td>
<td>MIROC</td>
<td>Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies</td>
</tr>
<tr>
<td>MIROC-ESM-CHEM</td>
<td>MIROC</td>
<td></td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>MPI-M</td>
<td>Max Planck Institute for Meteorology</td>
</tr>
<tr>
<td>NorESM1-M</td>
<td>NCC</td>
<td>Norwegian Climate Centre</td>
</tr>
</tbody>
</table>
Table 3.2(a) The SLP differences between 60°W-0°W and 120°E-180°E averaged over three latitudinal bands in PD, GW and their difference (units: Pa). (b) Same as in (a) except for SLP differences between 70°W-0°W and 110°E-180°E.

<table>
<thead>
<tr>
<th>d SLP (Pa)</th>
<th>PD</th>
<th>GW</th>
<th>(GW-PD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5°S-0.5°N</td>
<td>446.295</td>
<td>444.078</td>
<td>-2.217</td>
</tr>
<tr>
<td>2°N-2°S</td>
<td>446.411</td>
<td>444.207</td>
<td>-2.204</td>
</tr>
<tr>
<td>5°N-5°S</td>
<td>445.952</td>
<td>443.947</td>
<td>-2.005</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>d SLP (Pa)</th>
<th>PD</th>
<th>GW</th>
<th>(GW-PD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5°S-0.5°N</td>
<td>428.057</td>
<td>428.726</td>
<td>0.669</td>
</tr>
<tr>
<td>2°N-2°S</td>
<td>428.3</td>
<td>429.05</td>
<td>0.75</td>
</tr>
<tr>
<td>5°N-5°S</td>
<td>428.223</td>
<td>429.237</td>
<td>1.014</td>
</tr>
</tbody>
</table>

Table 3.3 The percentage change \( \frac{GW-PD}{PD} \) of intensity of the Walker Circulation (unit: %) calculated based on the original definition of circulation at different longitude-latitude domain from 925hPa to 150hPa. Shown is the result for ensemble mean of convective parameterization schemes “YS”, “AS” and “KF”.

<table>
<thead>
<tr>
<th></th>
<th>120E-60W</th>
<th>120E-0W</th>
<th>120E-30E</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5N-0.5S</td>
<td>1.33%</td>
<td>2.04%</td>
<td>2.85%</td>
</tr>
<tr>
<td>2N-2S</td>
<td>1.25%</td>
<td>1.96%</td>
<td>2.75%</td>
</tr>
<tr>
<td>5N-5S</td>
<td>1.15%</td>
<td>1.98%</td>
<td>2.80%</td>
</tr>
</tbody>
</table>
Table 3.4 The percentage change rate ($\frac{GW-PE}{PD}$) of global averaged column integrated moisture, upward motion, precipitation rate, net downward radiation at surface (including both longwave and shortwave radiation), and net radiative cooling in the atmosphere.

<table>
<thead>
<tr>
<th></th>
<th>YS</th>
<th>AS</th>
<th>KF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture</td>
<td>+19.2%</td>
<td>+19.5%</td>
<td>+18.5%</td>
</tr>
<tr>
<td>Upward motion</td>
<td>-8.45%</td>
<td>-9.90%</td>
<td>-6.70%</td>
</tr>
<tr>
<td>Precipitation</td>
<td>+3.98%</td>
<td>+4.57%</td>
<td>+4.55%</td>
</tr>
<tr>
<td>Surface radiation (downward)</td>
<td>+2.82%</td>
<td>+2.87%</td>
<td>+1.92%</td>
</tr>
<tr>
<td>Radiative cooling</td>
<td>+3.13%</td>
<td>+2.71%</td>
<td>+2.78%</td>
</tr>
</tbody>
</table>
Figure 3.1 Differences of surface temperature (units, K; 1.55K mean tropical warming between 30°N and 30°S is removed) and 850hPa wind (vector, m/s) between RCP4.5 and historical run in boreal summer (a) and winter (b), respectively. (c) 850hPa zonal wind changes averaged between 5°N and 5°S (units, m/s) in boreal summer (red), winter (blue) and annual mean changes (black). Arrows denote areas where westerly anomaly are significant during different seasons.
Figure 3.2 (a) annual mean precipitation (mm/day) in CTRL with (contour) or without prescribed land surface temperature (shading); (b) and (c) are same as in (a), but for boreal summer and winter, respectively.
Figure 3.3 (a) differences of annual mean precipitation (shading, mm/day) and 850hPa wind (vector, m/s) between DW and CTRL, contour for SST differences with tropical mean warming between 30°N and 30°S removed; (b) as in (a), but for the differences between MW and CTRL, contour for precipitation pattern in CTRL (5mm/day interval); (c) differences between LW and CTRL. (d) differences of 850hPa zonal wind averaged between 5°N and 5°S (m/s) in DW (green), MW (blue) and LW (red) compared to CTRL. Black is the sum of the three curves.
Figure 3.4 differences of annual mean SLP fields (hPa) in (a) DW, (b) MW and (c) LW compared to CTRL.

Figure 3.5 as in Figure 3.3, but for the differences in boreal summer (JJA).
Figure 3.6 as in Figure 3.4, but for the differences in boreal summer (JJA).

Figure 3.7 as in Figure 3.3, but for boreal winter (DJF).
Figure 3.8 as in Figure 3.4, but for boreal winter (DJF).

Figure 3.9 Differences of area averaged 850hPa zonal wind changes in DW (green), MW (blue) and LW (red) compared to CTRL. Domain 165°W-90°W, 5°N-5°S is chosen for annual mean changes. 150°E-90°W, 5°N-5°S for boreal summer and 165°W-90°W, 5°N-5°S for changes in boreal winter.
Figure 3.10 Annual mean precipitation changes in (a) “total forcing”, (b) SST’ and (c) “mean warming and land warming” (units: mm/day). Contours denote (b) SST differential warming (K, 0.2K interval) and (c) rainfall fields (mm/day, 5mm/day interval) in CTRL, respectively. Boxes locate at tropical Pacific (120°E-160°W, 5°N-5°S).

Figure 3.11 Horizontal pattern of a prescribed SST field in the PD simulation. The SST field is symmetric about the equator. It peaks at the equator and decreases poleward. Along the equator, highest SST (31°C) is located at 120E and lowest SST (27°C) is located at 60°W.
Figure 3.12 Zonal-vertical cross sections of zonal overturning circulation (vector) and vertical p-velocity (shaded, unit: Pa/s) averaged between 2°S and 2°N derived from the ensemble average of the “YS”, “AS” and “KF” simulations. (a) PD simulation, (b) GW simulation, (c) GW-PD. (d) zonal distribution of SST (dashed) and precipitation (solid) in the PD simulation.
Figure 3.13 (a) The vertical profile of p-velocity (red, unit: Pa/s) and static stability (green, unit: 0.5 K/hPa) averaged over 110°E-130°E and 2°S-2°N. (b) The vertical profile of $Q_1$ (green, $10^{-4}$ K/s), horizontal temperature advection (blue, $10^{-4}$ K/s) and adiabatic heating (red, $10^{-4}$ K/s) averaged over 110°E-130°E and 2°S-2°N. In both panels, the vertical profiles are derived from the ensemble average of the “YS”, “AS” and “KF” simulations. Solid line denotes GW and dashed line denotes PD.

Figure 3.14 Vertical profiles of zonal mean zonal wind (units: m/s) averaged in (a) 120E-60W, (b) 120E-0W and (c) 120E-30E. (d) Mass weighted column-integrated omega (units: Pa sm$^{-1}$) averaged between 2°N and 2°S. Solid (dashed) curve is for PD (GW) simulations (blue for “YS”, red for “AS” and green for “KF” experiment).
Figure 3.15 The vertical p-velocity changes averaged in the upper (100-300hPa, blue, unit: Pa/s) and lower (300-1000hPa, red) troposphere and contributions to the vertical velocity changes by the diabatic heating term (column 2), the static stability term (column 3), and sum of the two terms (column 4). All values are averaged over (110°E-130°E and 2°S-2°N) based on the ensemble average of the “YS”, “AS” and “KF” simulations. Whiskers stand for maximum and minimum values for each term in “YS”, “AS” and “KF” simulations.
Chapter 4

Conclusions

4.1. Key points

Because of the emission of greenhouse gases due to anthropogenic activities, more downward longwave radiation is trapped within the earth system, leading to a higher global mean surface temperature. Prominent changes of the climate system have taken place and will continue to occur based on the climate model projections, e.g. weakening of the Walker circulation and El Nino like warming pattern over tropical Pacific; alternation of rainfall patterns (Fig. 1.1). In my PhD dissertation, I aim to figure out the formation mechanisms of the coupled SST-circulation-precipitation patterns under global warming, by analyzing the climate model simulations. Especially, I focus on analyzing the fundamental causes of the weakened Walker circulation/El Nino like SST warming over tropical Pacific.

4.2. Summary and discussion

Most climate models that participate CMIP5 project weakened Walker circulation under global warming due to effect of greenhouse gases (GHG), along with El Nino like SST warming over tropical Pacific. It is
further illustrated that trade winds are weakened even in the uniform warming experiment, with no changes in SST gradients (Fig. 1.2). Previous studies suggest that these changes are attributed to the slower increase rate of global mean precipitation than moisture, which requires weakened upward motion (Held and Soden 2006). This argument is essentially derived from global averaged moisture budget, and inexplicitly assumes that the area of ascending motion does not change under global warming. An alternative argument is that because of greater warming in the upper troposphere compared to the lower level, static stability in the troposphere is enhanced, which leads to the weakened Walker circulation (Knuston and Manabe 1995). Yet in a warmer climate, greater latent heating in the troposphere due to more moisture released from the surface could support deeper convection over tropical Pacific. These two competing processes together determine the changes of vertical motion.

Given the deficiencies in the previous studies, the fundamental causes of the weakened Walker Circulation and an El Nino like warming in the tropical Pacific under global warming in CMIP5 models are investigated, and it is illustrated that the following three mechanisms are responsible.

The first mechanism is a so called “longwave radiative – evaporative damping” mechanism. A simple analytical model was constructed to understand the formation mechanisms of future SST warming pattern under global warming. It is demonstrated that the future SST warming pattern is primarily determined by present-day mean SST and surface latent heat flux ($Q_{lh}$) fields. Assume a local thermodynamic equilibrium without circulation and cloud changes, a uniform GHG forcing would lead to a smaller (larger) warming in the regions where mean SST and $Q_{lh}$ are large (small). Because of this fundamental feature of the present-day mean climate dependence, all CMIP5 models projected a robust meridional pattern with a much greater warming in the pole than in the tropics. The mean-state induced warming contrast is further amplified by atmospheric longwave radiation feedback associated with cloud changes.

The occurrence of an El Nino like warming in the equatorial Pacific and Atlantic arises from a similar
process (longwave radiative – evaporative damping mechanism) but with weaker amplitude and a smaller signal-to-noise ratio. Different cloud feedback in east and west of the two ocean basins also plays a role. In the warm pool region, more deep convective clouds would form when the ocean surface becomes warmer. In the cold tongue region, low stratus clouds are pronounced due to a stable condition in the lower troposphere associated with trade wind inversion. Under global warming, even though the troposphere becomes more stable, the subsidence is weakened in eastern Pacific due to weakened Walker circulation. The competition between the two leads to a small negative cloud change there. This different cloud feedback due to different cloud regimes lead to less (slight more) surface solar radiation in west (east) of the ocean basin and thus favors an El Nino like warming. However, it is noted that such process is associated with weakened Walker circulation, therefore this feedback is essentially a feedback that may amplify the changes in both SST gradients and trade winds, but it is not a fundamental cause. The less robustness of the east-west warming contrast along the equator than the meridional warming contrast among the models can also be explained by different magnitudes of total denominator contrast. The warming mechanism in the IO, on the other hand, differs markedly from the Pacific counterpart. It is primarily attributed to the remote forcing from the equatorial Pacific through anomalous Walker circulation.

The second mechanism is “the richest get richer”. As revealed by Hsu and Li (2012), not all the rainfall centers becomes stronger even in a uniform warming experiment. Instead, the stronger convection centers become wetter by driving anomalous convergent flow, while the relatively weaker ones are suppressed due to the circulation change (“richest-get-richer”). Therefore, the changes of major convection centers go hand in hand with anomalous flow pattern under global warming. How monsoon heat sources change due to mean surface warming, and how these changes influence the Walker circulation are examined. It is found that in response to a uniform surface warming, the Asia-western North Pacific (WNP) monsoon system is enhanced by competing moisture with other large-scale ascending systems. The strengthened WNP monsoon induces surface
westerlies in the western-central equatorial Pacific, weakening the Walker circulation. Idealized AGCM experiments that separate the effect of the mean warming and the relative SST warming pattern clearly demonstrate the effect of the mean warming on change of the equatorial wind.

The third mechanism is extra land surface warming. One prominent feature of the surface warming pattern in the future projections by the climate models is greater warming over land surface than that over the ocean. The additional land warming may alter the large-scale circulation pattern by affecting the SLP gradients. Such effect might also be important for trade wind changes. In particular, a great thermal contrast between American continent and tropical Pacific causes a zonal pressure gradient and westerly anomalies in the eastern equatorial Pacific. This weakens the Walker circulation.

In addition to the aforementioned mechanisms, atmospheric response to SST differential warming pattern is also investigated for comparison, since changes in SST gradients can alter tropospheric circulation significantly through air-sea coupling. It is found that SST differential warming indeed plays an important role in driving the weakened trade winds (around 58%), but it is also noted that such process is just an amplifier through positive feedback between atmosphere and ocean, rather than a fundamental cause.

The relative roles of different surface warming patterns in driving precipitation changes is also investigated, and the result is consistent with the aforementioned results: around 60% of total rainfall change over tropical Pacific is due to changes in SST gradients, while other factors are also important.

To sum up, by conducting numerical simulations in a realistic land-ocean distribution, we demonstrated that the weakening of the Walker Circulation is attributed to both the strengthened WNP monsoon heat source and a greater land warming over America. The relative SST warming pattern also plays a role, but it is just an amplifier, not a fundamental cause. A schematic picture is shown as Fig. 4.1 to illustrate the three mechanisms, which we think are essential processes, that give rise to the weakened Walker circulation/El Nino like SST warming in the first place. Changes in SST gradients and trade winds are then amplified through positive
feedback between atmosphere and ocean in the tropical Pacific.

A traditional view of weakened Walker circulation is attributed to a slower increase rate of global mean precipitation than moisture, which does not concern about land-ocean distribution, while we emphasize the land and monsoon effects. We design a set of idealized numerical experiments in an Aqua-planet Earth to eliminate the monsoon and land impacts and examine the “pure” Walker circulation response to global warming. This way, we can examine whether theories put forward by previous studies are sufficient to explain the changes of Walker circulation or not.

A zonal wavenumber-1 SST distribution is prescribed in the tropics in the present-day simulation. It is found that the Walker Circulation is strengthened under such an idealized world, given a uniform 2K SST warming. A diagnosis shows that the ascending branch of the Walker cell is enhanced in the upper troposphere but weakened in the lower troposphere. As a result, a “double-cell” circulation change pattern forms, with a clockwise (anti-clockwise) circulation anomaly in the upper (lower) troposphere. The upper tropospheric circulation change dominates the strength change of the Walker circulation. The mechanism for the formation of the “double cell” circulation pattern is attributed to a greater (smaller) increase rate of diabatic heating than static stability in the upper (lower) troposphere.

While the Walker circulation is strengthened under global warming, global average upward motion is weakened in the Aqua-Planet simulations, which is consistent with the slower increase rate of global mean rainfall than moisture in the model. This points out that the global moisture budget argument by Held and Soden (2006) is valid even in an idealized Aqua Planet model.

The fact that the Walker circulation is strengthened in an Aqua Planet simulation but weakened in the future projections of CMIP5 climate models (in which realistic land-sea distributions and monsoon flows are presented) suggest that the global moisture budget argument seems to be applicable to changes of global mean upward motion, but may not be sufficient to explain the change of the Walker circulation that is confined in the
equatorial region. Therefore, the major factors that affect the equatorial overturning circulation must involve how it interacts with the monsoon heat sources over South Asia and western Pacific warm pools and how unequal land and ocean warming alters zonal pressure gradients.

4.3. Future work

The impacts of global warming are analyzed in this study, using outputs from numerical experiments. However, it is noted in many previous studies that great uncertainties still exist among different climate models, which mainly arises from cumulus parameterization (e.g. Stocker et al. 2001; Allen et al. 2002; Ma and Xie 2013). Yet it is demonstrated in Chapter 3 that cloud feedbacks in the models play an important role in the formation of SST differential warming pattern, which further drives circulation change and alters rainfall pattern. This discrepancy of climate models remains a tough problem to be dealt with, and this may require the development of better cumulus parameterization or very high-resolution (“cloud - resolving”) models. Although there is under 1-km climate models nowadays, it is not likely to run such model for several decades to obtain the GHG induced changes due to limit of the computer power. The mean state in climate models also suffers lots of discrepancies, such as “cold tongue bias” (cold tongue in eastern Pacific is too strong in the model climatologies compared to the observation), double ITCZ etc. Thus there is still great room for the improvement of the climate models.

Evaluating the impacts of global warming in the observational datasets is a challenging task as well because of the relatively short “reliable” observations. It is hard to distinguish the internal variability of the climate system (e.g. decadal variability) from the forced signals due to GHG effect associated with anthropogenic activities (e.g. Dinezio et al. 2009). This problem could be solved by analyzing multi-model ensemble mean outputs so that the internal variability in different models or in different ensemble runs can be eliminated. This
way, the responses of the climate system to the external forcing due to GHG effect can be examined (Fig. 5.1a).

However, it is also found that the surface warming trend has slowed down since late 1990s such that the linear warming trend over the past 15 years is nearly zero (Easterling and Wehner 2009; Kaufmann et al. 2011; Meehl et al. 2011; Trenberth and Fasullo 2013), and some previous studies have attributed this phenomenon to the internal decadal variability of the climate system (Easterling and Wehner 2009; Meehl et al. 2011; Kosaka and Xie 2013; Chen and Tung 2014). In my future research, I’d like to examine what causes this slow down of global warming.

Kosaka and Xie (2013) showed that by forcing their model with observed tropical EP SST, the trend of which exhibits a La-Nina like pattern since the year 2000, the evolution of global mean surface temperature is very well simulated, including global warming hiatus. This provides strong evidence that the emergence of EP cooling is critical for the global warming hiatus. What is the mechanism that causes such strong cooling in EP region on decadal time scales? Another question is with net incoming top-of-atmosphere (TOA) radiative forcing in the past 15 years, if the heating is not warming the surface, where does the heat go? Previous study suggests that at least 90% of incoming radiative forcing goes into the ocean (e.g. Trenberth and Fasullo 2013). Meehl et al. (2011) found that the surplus heat is mainly being restored in the ocean below 300m. How the decadal variability of the ocean influences the distribution of ocean heat content needs to be examined as well.

Because of short length of reliable global observational datasets, analyzing decadal variability is difficult. The observational data of deep ocean is quite limited as well. On the other hand, it is suggested by previous studies that hiatus events are captured in the future projections of some climate models forced by increasing greenhouse gases. Therefore, I propose to analyze the climate models that participate CMIP5 to examine the causes of hiatus events. Shown in Fig. 5.1 are ensemble mean results derived from 34 climate models under representative concentration pathway 4.5 (RCP4.5). Overall, global mean surface temperature (ST) increases at a rate around \(0.16K \pm 0.05K\) per decade (Fig. 5.1a). By calculating ten-year running trend of global mean ST, it
is found that near zero or even negative trends show up at times in a 95-year simulation in each model. But the frequency of such events varies significantly from model to model, e.g. only 2 and 3 events with ten-year trend smaller than zero show up in CMCC-CMS and IPSL-CM5A-MR, while 23 and 46 events are captured in BNU-ESM and GISS-E2-R-CC. To obtain more robust signals of the hiatus events, only the periods with decreasing trend of global mean ST smaller than -0.1K per decade are picked out, and 26 out of 34 models can simulate such events. Shown in Fig. 5.1b is the ensemble mean SST trend of 137 hiatus events, and stippled area are the regions where more than 110 events (80% of the total 137 events) agree with the sign of the ensemble mean SST trends (shading). The most prominent and robust feature of SST trends during these hiatus events is that significant cooling trend shows up in EP, which is consistent with observation in the past decade. This indicates that the hiatus events can be simulated by the climate models.

In my research, using multi-model ensemble mean approach, I would like to address what possible mechanism is responsible for the cooling in EP region on decadal time scales in the climate models, and how the ocean heat uptake is affected during hiatus events. The SST trends shown in Fig. 1b are remarkably similar to SST anomaly pattern associated with PDO or IPO. Trenberth and Fasullo (2013) showed that PDO experiences phase transition around 2000. Jin (2001) found that low frequency modes (LFM) with zonally uniform deepening and shoaling of equatorial thermocline on decadal time scales are possible, which is due to off-equatorial Rossby waves. The physical mechanism of this LFM is similar to recharge oscillator theory for ENSO, but with broader latitudinal band and lower frequency, which is also consistent with results shown in Fig. 5.1b. How the free LFM is transformed into coupled mode is still unclear. In my future study, I will examine whether this LFM is related to the EP cooling during hiatus events.

The other issue is the changes in ocean heat storage during global warming hiatus events. The deep ocean heat uptake is crucial in slowing down of surface warming trend. Although Chen and Tung (2014) argues that Atlantic Ocean plays a pivotal role in storing heat in deep ocean, study by Meehl et al. (2011) et al. suggests
that changes in ocean heat storage is more evident in Pacific. This hot debated issue will also be examined in my future research.

Previous studies also point out that hiatus is more prominent during boreal winter (e.g. Trenberth and Fasullo 2013), determination of the seasonality of the hiatus will also be studied in my research. If possible, I would like to show the typical duration times of these events along with their uncertainty across different models.
Figure 4.1 Schematic diagram that illustrates the fundamental processes that give rise to the weakening of the Walker circulation/El Nino like SST warming over tropical Pacific under global warming.
Figure 4.2 (a) Evolution of ensemble mean, global mean surface temperature (units, K), with standard deviation across the models (shading, K). (b) Ensemble mean SST trend during hiatus events, (shading, K/decade). Stippled areas are regions where SST trend in at least 110 events (80% of total 137 events) agree with ensemble mean SST trend.
Bibliography


Chen X. and Tung K., 2014: Varying planetary heat sink lead to global-warming slowdown and acceleration. *Science*, **345**. DOI:10.1126/science.1254937


Erich Roeckner, Klaus Arpe, Lennart Bengtsson, Michael Christoph, Martin Claussen, Lydia Dümenil, Monika


