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A model intercomparison of the tropical response to a CO$_2$ doubling in aquaplanet simulations

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Abstract

The present-day Earth’s climate has warmer Northern Hemisphere (NH). This hemispheric asymmetry is expected to be amplified in response to increasing CO$_2$. It is of question whether the tropical precipitation consistently shifts toward the even warmer NH. We employ four different climate models (AM2, AM3, HiRAM, and CAM5) that are coupled to an aquaplanet slab ocean. In simulations, a northward ocean heat transport is prescribed to mimic the present-day climate state of a warmer NH. This reference state is then perturbed by a doubling of CO$_2$ to explore the response of tropical precipitation to a uniform radiative forcing. Even though the forcing is uniform in space, the hemispherically asymmetric response of cloud radiative forcing results in the cross-equatorial atmospheric energy transport change, which induce the tropical precipitation shift. Yet, the sign of cross-equatorial atmospheric energy transport is not robust across models, causing a large spread in the response of tropical precipitation. Furthermore, even in the case of little changes in the atmospheric energy transport, there is a significant shift in the tropical precipitation. It is shown that the total gross moist stability ($\Delta m$) changes, which has been often neglected in previous studies, are critical for understanding the response of tropical precipitation to uniform CO$_2$ forcing. Large uncertainties in $\Delta m$, due to the dependence of the vertical structure of moist static energy on convection schemes and cloud modeling, calls for their improvement to better project tropical precipitation in the future.
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I. Introduction

Global warming that depicts the increase in the average temperature of Earth is the main characteristic of the current climate. Over the past 100 years, the global average temperature has steadily increased as a result of the increase in greenhouse gas concentration, and the globe is projected to continue to get warmer at a rapid rate in the future. Although the exact magnitude of the global-mean increase in temperature and precipitation vary across climate models, all models consistently predict that the atmosphere will become warmer and wetter as the globe warms.

Although the globe has indeed gotten warmer on average, this warming has not occurred uniformly in time and space. For the past 15 years, there has been a slowdown of global warming, referred to as global warming hiatus (Kosaka and Xie 2013). Also, some parts of the world such as near the Andes had experienced cooling from about 1910 to 1980 (Ji et al. 2014). The surface warming is strongest in the Arctic region (e.g. Serreze and Barry 2011), a phenomenon commonly known as the Arctic amplification. There is greater warming over land in the Northern Hemisphere (NH) and less warming over the Southern Ocean (Screen and Simmonds 2010). Hence, the NH is predicted get warmer than the SH, which is to increase the existing hemispheric asymmetry. Even in the pre-industrial period, the warmer NH exists, which is suggested to be a result of a northward cross-equatorial ocean heat transport (Kang et al. 2014). This is also the reason for the band of heavy rainfall in the tropics, called the inter-tropical convergence zone (ITCZ), to be located on the northern side of the equator (Marshall et al. 2013, Frierson et al. 2013).

It is of question whether the ITCZ will be consistently shifted toward the warmer NH in response to increasing CO$_2$. If the Earth climate were to be hemispherically symmetric, the uniform radiative forcing would not be able to shift the ITCZ to make it favor one hemisphere. However, our current climate is hemispherically asymmetric with a warmer NH and the ITCZ in the NH. Then, the uniform radiative forcing can result in a meridional ITCZ shift due to nonlinear cloud radiative feedbacks (Kang et al. 2008, Shaw et al. 2015). Indeed, in an aquaplanet world, when the reference state with the warmer NH is perturbed by a doubling of CO$_2$, the ITCZ is shown to be shifted northward (Merlis et al. 2013). It is of interest to investigate the robustness of the ITCZ response to increased CO$_2$ in the hemispherically asymmetric state. As CO$_2$ concentration is increased, the hemispheric asymmetry in the surface temperature will be strengthened, which may lead to a consistent ITCZ shift to the north. In contrast, the cloud responses may act to preferentially warm the SH, leading to a southward ITCZ shift. Large uncertainties in cloud modeling motivate to investigate the robustness of the ITCZ response. Another complication may arise from the dependence of gross moist stability on sea surface
temperatures (Hill et al. 2015), as we shall see.

To eliminate complications from realistic boundary conditions while keeping the hemispherically asymmetric state, we utilize the atmospheric model coupled to an aquaplanet slab ocean and prescribe the northward ocean heat transport. We first describe the models and experiment setup in Section 2. Then, the detailed backgrounds are introduced in Section 3, followed by the results in Section 4. Finally, we discuss the implication of this study in Section 5.
II. Model description and experimental design

2.1. Models

The experiments are designed to study the mechanism how the off-equatorial ITCZ respond to a doubling of CO$_2$. To investigate the robustness of the results, four different models are employed: AM2 (Anderson et al. 2004), AM3 (Donner et al., 2011), HiRAM (Zhao et al., 2009), and CAM5 (Medeiros et al., 2016). All are coupled to an aquaplanet slab ocean of 2.4 m depth. All simulations are integrated for 10 years, with a spin-up period of 2 years.

GFDL-AM2 has a horizontal resolution of $2^\circ$ latitude $\times 2.5^\circ$ longitude and 17 vertical levels. The model uses the relaxed Arakawa–Schubert convection scheme of Moorthi and Suarez (1992). And ensemble of entraining plumes which is determined by different lateral entrainment rates consists convection as in Arakawa and Schubert (1974). These lateral entrainment rates are determined by calculating the required values to make plumes to get their neutral buoyancy levels corresponding to model levels. When a lateral entrainment rate ($\lambda$) of the model is smaller than a critical value ($\lambda_0$), deep convection is hard to occur in updrafts (Tokioka et al. 1988). The subcloud layer depth $z_M$ determines a critical value $\lambda_0$ as $\lambda_0 = \alpha/z_M$ where $\alpha = 0.025$ in the standard case. With larger $\alpha$, the deep convection is hard to occur as the more dry air entrains into the plume as it ascends. Then the fraction of tropical precipitation occurred by large-scale condensation increases. To examine the sensitivity to a convective parameter, $\alpha$ is multiplied by a number of factors ($0, 0+, 4, and 5$). The notation $0+$ refers to an infinitesimally small number, so that only the nonentraining deep convective plume is omitted. The standard case is denoted as AM2(1X).

GFDL-AM3 has a horizontal resolution of $2^\circ$ latitude $\times 2.5^\circ$ longitude and 48 vertical levels. As the model is the first global atmospheric model of GFDL including cloud-aerosol interactions, all cloud parameterizations in AM2 were including sub-grid distributions of vertical velocity which need to be considered aerosol activation. They are used to calculate stratiform clouds (Golaz et al., 2011), shallow convection with entraining plumes, buoyancy sorting and vertical velocity (Bretherton et al., 2004), and deep convection represented by an plume ensemble with vertical velocities and mass fluxes (Wilcox and Donner, 2007, and Donner et al., 2001).

GFDL-HiRAM has a 50km horizontal resolution and 32 vertical levels. The model uses mass flux convection scheme with a single bulk plume for both detraining and entrainments. Based on buoyancy sorting, both entrainment and detrainment rate is calculated. The environment can interact with buoyancy sorting dynamically and thermodynamically after Bretherton et al (2004).
CESM1-CAM5 has a resolution of 1.9° latitude × 2.5° longitude and 30 vertical levels. As described in Neale et al (2010), the parameterization of deep convection is a plume ensemble approach with closure which is based on convective available potential energy (CAPE) as computed for an entraining parcel including momentum transport by convection. A shallow convection is parameterized using a mass-flux closure and bulk plume approach and it is separately calculated on all model levels.
2.2. Experiment setup

To locate the ITCZ in the NH, as in the Earth’s current climatology, a given amount of heat within the slab ocean is moved from poleward of 40°S to poleward of 40°N, as used in Kang et al. (2008). The zonal-mean profile of the prescribed forcing is shown in Fig. 2.2.1. In response to the northward ocean heat transport, the ITCZ is located in the northern tropics. Even though the same magnitude of ocean heat transport is prescribed, the exact ITCZ location varies widely across models (Figs. 2.2.2 and 2.2.3). The AM2(0) exhibits the ITCZ farthest away from the equator at 13.78°N and CAM5 exhibits the ITCZ nearest the equator at 7.86°N. The ITCZ location is closely related to the hemispherical asymmetries of SST, which also exhibits a wide range from 11.96 K in CAM5 to 20.05 K in AM3 (Fig. 2.2.2 and 2.2.3). Large spread in the climate response to the same prescribed ocean heat transport is interesting in and of itself, but this is the beyond the scope of our study.

The reference climate in which the warmer NH exists, is then perturbed by a CO$_2$ doubling. The CO$_2$ concentration of the reference climate is 348 ppm. The responses to CO$_2$ doubling are obtained by differencing the climatologies of the reference integration (1xCO2), from that of the doubled CO$_2$ integration (2xCO2), and is denoted as $\delta$.

Fig. 2.2.1. The latitudinal distribution of imposed heating (W/m$^2$).
Fig. 2.2.2. The ITCZ location (°) and the SST difference between the NH and the SH (K) in the reference (1xCO2) integration.
Fig. 2.2.3. The zonal-mean (a) Sea Surface Temperature (SST; K) and (b) precipitation (mm/day) in the reference (1xCO2) integration.
III. Background

3.1 Robust responses of global warming

Global warming refers to the gradual warming of the average temperature of the Earth’s climate system and its related effects. Its primary cause is considered to be the greenhouse gases produced by human activities, as reported in the Intergovernmental Panel on Climate Change (IPCC, 2013). As the concentration of greenhouse gases is increased by human activities, the more terrestrial radiation is trapped within the atmosphere and is radiated back to the surface, resulting in global warming.

All climate models predict continuous global mean surface temperature increases over the 21st century, even though the amplitude of predicted warming is diverse among models (Boer et al. 2000; Dai et al. 2001; Held and Soden 2000; Delworth and Knutson 2000; Houghton et al. 2001; Meehl et al. 2000; Lucarini and Russell 2002; Yonetani and Gorden 2001; Washington et al. 2000; Watterson and Dix 1999). Despite the uniform increase in greenhouse gases, there are substantial spatial patterns in the surface temperature and rainfall changes. Consistent with observations during the late 20th century, models predict less warming over North Atlantic and the southern oceans and the larger temperature increases over land and at high northern latitudes (IPCC, 2007). As the NH is predicted to warm faster than the SH, hemispherical asymmetries in surface air temperature is projected to become larger (Hill et al, 2015) since the NH is currently warmer than the SH (Kang et al. 2014). Both surface flux adjustments and ocean dynamics are considered as a critical factor to emerge these substantial variations in SST warming. For example, ocean dynamics especially changes in mode water ventilation generate narrow banded SST warming structures in the northern extratropics. Also, a stronger warming in the northern subtropics than in the southern subtropics can be understood by the hemispheric asymmetry in trade wind changes which is due to the strengthening of the southeast trades and weakening of the northeast trades (Xie et al. 2010). Not only there are changes in the magnitude and pattern of surface temperature, but also there are substantial changes in the Earth’s atmosphere under global warming (Meehl et al. 2007). One of the robust features is the intensification of the overall global hydrological cycle. Precipitation is predicted to increase in the tropical precipitation maxima region, over the tropical Pacific, and to decrease both in the subtropics and at high latitudes. This is a common feature among climate models despite imperfect simulation of precipitation distribution, which is explained by “rich-get-richer mechanism” (Held and Soden 2006; Chou and Neelin 2004): the atmospheric water vapor increase under global warming (relatively unchanged relative humidity) is advected by the mean vertical motion to enhance the global hydrological cycle. Also, the spatial variations of the sea surface warming are another factor.
contributing to the distribution of tropical precipitation response (Xie et al. 2010). As the tropical tropospheric temperature is flat, convective instability change locally and then precipitation is determined by relative SST, the deviation of the tropical mean: this is the warmer-get-wetter mechanism. Both mechanisms are at play in determining the spatial distribution of tropical precipitation (Huang et al. 2013).

Although it is inconclusive in observations, the weakening of tropical circulation is also one of the most robust changes in global warming simulations (Mitas and Clement 2005; Shon and Park 2010; Song and Zhang 2009; Tanaka et al. 2005; Vecchi et al. 2006; Young et al. 2011; Vecchi and Soden 2007). The weakening is evident in the zonally asymmetric component, the Walker circulation, whereas the zonally symmetric component, the Hadley circulation, exhibits large uncertainty (Ma et al. 2012, Vecchi et al. 2006). Based on a simple moist budget scaling, the strength of tropical circulation is predicted to weaken as the column-integrated water vapor increases more rapidly than the globally-averaged precipitation (Held and Soden 2006). A cause for the weakening of the tropical circulation is proposed to be the increase in atmospheric stability associated with the deepened convection in a warmer climate (Chou and Chen 2010).
3.2 The shift of the ITCZ

The Inter-Tropical Convergence Zone (ITCZ) is a narrow zonal band of high rainfall in the tropics. Because many tropical societies rely upon these rains, even a small change in the position of the ITCZ can cause large societal impacts.

In the past, the ITCZ was often thought to be determined by tropical mechanisms (e.g., Xie 2004). For example, Xie and Philander (1994) introduced the wind-evaporation-SST (WES) feedback as a possible mechanism to locate the ITCZ on the north of the equator when SST becomes slightly warmer in northern side of the equator than to the south. The impact of stratus-SST feedback which indicates that increased stratus cloud induced by SST cooling acts to make the initial sea surface cooling larger is also suggested to make ITCZ shift (Philander et al. 1996). The effect of ocean upwelling/downwelling change that can generate anomalous SST gradient (Chang and Philander 1994) and the shape of the continents (Philander et al. 1996) are also thought to be the factor. However, recently, the ITCZ location is thought to be determined by the cross-equatorial ocean transport. The ocean transports energy to the north of equator, mostly by the Atlantic Meridional Overturning Circulation (AMOC), resulting in the ITCZ which is located in NH (Marshall et al. 2014, Frierson et al. 2013). Indeed, using an idealized coupled GCM, Fučkar et al. (2013) showed that the direction of cross-equatorial heat transport by AMOC can largely affect to control the ITCZ shift.

Recent studies have shown that the ITCZ can respond to heating well outside the tropics. For example, the increased NH ice cover in the high latitudes is shown to cause a southward shift of the marine ITCZ (Chiang and Bitz 2005). A weakening of the thermohaline circulation that induces cooling over the northern North Atlantic is also shown to shift the ITCZ southward (Zhang and Delworth 2005, Stouffer et al. 2006). The sulfate aerosol emissions from the NH midlatitudes are proposed to have induced global southward precipitation shift in the late 20th century (Hwang et al. 2013, Rotstayn and Lohmann 2002). In contrast, increased concentration of absorbing aerosols such as black carbon leads to the northward ITCZ shift (Roberts and Jones 2004, Yoshimori and Broccoli 2008, Mahajan et al. 2013). Also, interestingly, a factor to cause the double ITCZ problem of general circulation models (GCMs) is proposed to lie in the extratropics, the cloud biases over the Southern Ocean (Hwang and Frierson 2013). Using idealized experiments, Seo et al. (2014) have demonstrated that the extratropical forcings can induce the shift of the ITCZ more efficiently than the tropical forcing due to large cloud radiative feedback. The mechanism for these shifts of the ITCZ is introduced in the next subsection.
3.3 Energetic framework & limitation

The ITCZ latitudinal position can be understood using the atmospheric energy budget (Kang et al. 2008, 2009). The ITCZ is located in the hemisphere with more energy input to the atmospheric column. Even when the hemispheric asymmetry is introduced in the extratropics (as presented in Section 3.2), through eddy energy fluxes, its effect can reach the tropics. Even, the extratropical energy input is shown to be more effective at affecting the ITCZ location than the low latitude energy input derived from nonlinear cloud radiative feedbacks (Seo et al. 2014). The Hadley circulation (HC) reacts to transport energy from the hemisphere with a surplus of energy to the hemisphere with a deficit of energy. In most cases, the energy is transported in the direction of the HC upper branch, so abundant moisture in the lower troposphere will be transported in the opposite direction to the energy. Hence, there is negative correlation between the shift of the ITCZ and the cross-equatorial atmospheric energy transport ($F_A$). This mechanism is shown in the schematic diagram in Fig. 3.3.1. The ITCZ is typically located where the moisture convergence attains its maximum value, so that it is slightly on the equatorward side of the latitude where the mean meridional wind changes sign ($\theta_{v_2=0}$), because the HC in the colder hemisphere is stronger (Fig. 3.3.2).

Adopting the concept of total gross moist stability, that is defined as the amount of total atmospheric energy transport per unit mass transport ($\Delta \equiv F_A/v_2$), as in Kang et al. (2009), the atmospheric energy transport, $F_A$, is related to the lower-level mass flux, $v_2$, as

$$F_A \equiv -v_2\Delta m.$$  \hspace{1cm} (1)

$F_A$ is the vertical integral of meridional moist static energy transport and $v_2 = \int_{p_0}^{p_m} v dp/g$, with $p_0$ as the surface pressure and $p_m$ as the mid-tropospheric level at which the maximum value of a vertically integrated mass flux exists. Positive $v_2$ indicates a northward lower-level mass flux, and vice versa. The negative sign is included in Eq. (1) for positive $\Delta m$ to indicate a larger energy transport in the upper-level than in the lower-level. In the reference climate, $p_m$ generally falls between 800 and 600 mb (Fig. 3.3.4). In response to a doubling of CO$_2$, $p_m$ consistently increases in all integrations, but only by less than 3 % (Fig. 3.3.4). As shown in Fig. 3.3.5, the effect of changing $p_m$ has negligible impact on the estimate of $\Delta m$. The estimate of $\Delta m$ when $v_2$ is computed with the fixed $p_m$ at 700 mb in all integrations is similar to that when $v_2$ is computed with a variable $p_m$ for each integration. Hence, $p_m$ is fixed to be 700 mb for simplicity.

Eq. (1) indicates that the energy flux equator $\theta_e$ where $F_A = 0$ coincides with $\theta_{v_2=0}$. However, because there are some eddy energy fluxes in the tropics and the rising branch of the HC is not
perfectly straight vertical, $\theta_e$ and $\theta_{\psi=0}$ are not identical, but $\theta_e$ is located on the equatorward side of $\theta_{\psi=0}$ (Fig. 3.3.2). However, the relative position between the ITCZ and $\theta_e$ is not constrained. In our experiments, seven out of eight models exhibit the ITCZ in between $\theta_e$ and $\theta_{\psi=0}$ (Fig. 3.3.2), as is the case in Fig. 3.3.1. Although their exact locations are different, we may expect them to react similarly to the heat perturbation. As the NH is warmed relative to the SH, for instance, the anomalous energy needs to be transported southward across the equator, which will be accomplished by displacing the rising branch of the HC northward to make the southern HC stronger. Then, all of the three latitudinal indexes ($\theta_e$, ITCZ, and $\theta_{\psi=0}$) are expected to be shifted northward. However, the response of $F_A$ can result from changes in either $v_2$ or $\Delta m$:

$$\delta F_A = -\Delta m \cdot \delta v_2 - \delta \Delta m \cdot v_2$$

where $\delta$ indicates the response to some thermal forcing (i.e., a doubling of CO$_2$). In one extreme case, where changes in $\Delta m$ are negligible, $\delta F_A$ arises from $\delta v_2$, so that $\theta_e$ can infer $\theta_{\psi=0}$, hence, the ITCZ. Previous studies have shown that $\theta_e$ can be shifted from changes in the cross-equatorial $F_A$ (blue in Fig. 3.3.3; Kang et al. 2008) and the net energy input to the equatorial atmosphere, equivalent to the divergence of $F_A$ (red in Fig. 3.3.3; Bischoff and Schneider 2014). Regardless of the origin of $\theta_e$ response, the ITCZ shift has been related to the $\theta_e$ shift, assuming that changes in $\Delta m$ are small. In the other extreme case, Eq. (2) indicates that $\delta F_A$ can solely arise from $\delta \Delta m$, in which case the $\theta_e$ shift will not be accompanied by the shifts of $\theta_{\psi=0}$ and the ITCZ. In fact, Shaw et al. (2015) shows that the cross-equatorial energy flux is affected by equatorial gross moist stability changes. In the case of $\delta F_A = 0$, there is no $\theta_e$ shift, but changes in $v_2$ and hence the ITCZ shift can arise from changes in $\Delta m$. Thus, we may expect some cases when the previous energy flux perspective fails to predict the ITCZ shift when the changes in $\Delta m$ are substantial. In this study, it is shown that the ITCZ response to a doubling of CO$_2$ cannot be understood without considering the changes in $\Delta m$. 
Fig. 3.3.1. The schematic of the mechanism depicting the ITCZ located in the hemisphere with a surplus of energy. Red arrows indicate the dry static energy transport and blue arrows indicate the moisture transport (in the unit of J/kg). At the energy flux equator ($\theta_e$), the two arrows have the same magnitude. Outside of $\theta_e$, in general, the dry static energy transport is slightly larger than the moisture transport, so that the total gross moist stability is positive indicating that the moist static energy is transported poleward.
Fig. 3.3.2. The ITCZ location (blue) and the location of the energy flux equator (red) with respect to the latitude where the vertically integrated mass flux maxima changes sign ($\theta_{vz=0}$) in the reference (1xCO2) integration.
Fig. 3.3.3. The schematic that describes qualitative shift of the ITCZ location as the northward cross-equatorial atmospheric energy decreases (blue) and as the net energy input to the equatorial atmosphere increase (red), adopted from Bischoff and Schneider (2014). The ITCZ shift is inferred from the energy flux equator ($\theta_e$) shift.
Fig. 3.3.4. The mid-tropospheric level at which the vertically integrated mass transport attains its maximum ($p_m$) and its response to a doubling of CO$_2$ ($\delta p_m$)
Fig. 3.3.5. (a) A comparison of $\Delta m$ with $p_m$ at fixed level ($\Delta m_{700}$) and that with varying $p_m$ in the reference (1xCO2) integration and (b) their responses to a doubling of CO$_2$. 
IV. Results

As \( \text{CO}_2 \) is doubled, due to the greenhouse effect, the global-means of both sea surface temperature (SST) and precipitation increase in all models (Fig. 4.1a). Consistent with previous studies, the global-mean increase of precipitation is smaller than that of specific humidity expected from the Clausius-Clapeyron relationship when the relative humidity is held fixed, which is often invoked to explain the weakening of tropical circulation (Held and Soden 2006). Indeed, the Hadley circulation is weakened by 4.04 % with one standard deviation of 3.27 %, in terms of the response of maximum mean meridional streamfunction (Fig. 4.2). In fact, there is a significant shift of the ITCZ, creating a large precipitation response of opposite signs, as shown in the zonal-mean profiles in Fig. 4.3b. Despite the robust increase in global-mean precipitation, the ITCZ response exhibits a large model spread (Fig. 4.4). Here, the ITCZ is obtained by finding the zero crossing in linearly interpolated differentiation of the precipitation with respect to latitude. The AM2(0+) exhibits a northward ITCZ shift as much as 1.64°, whereas AM2(4X) exhibits a southward shift of 0.19°. Although there is large spread in the ITCZ response, the NH becomes consistently warmer than the SH in all models (Figs. 4.1b and 4.3a). That is, the hemispheric asymmetry in the SST is intensified as a response to 2xCO2. Even with respect to the ITCZ location in the reference state, the SST warms more on the northern side than the southern side (Fig. 4.1b). However, when the SST response is compared 5° to the north and south of the reference ITCZ location, there is no consistent trend (Fig. 4.3b). Seven out of four models exhibit a more warming to the northern flank of the reference ITCZ, whereas three models exhibit an opposite response and one model (AM2(1X)) shows little contrast. That is, there are some models that warm more in the southern flank of the reference ITCZ yet the ITCZ is shifted northward, implying that the local SST response cannot explain the ITCZ response. Hence, we invoke the energy flux perspective to understand the ITCZ response.

As reviewed in Section 3.3, previous studies used the energy flux equator \((\theta_e)\), which is the latitude of vanishing total atmospheric energy flux, as a proxy for the ITCZ. Figure 4.4 compares the responses of the energy flux equator and the ITCZ. In AM2(1X), the ITCZ is shifted to the north, but the energy flux equator is shifted to the south. Some models (AM2(4X), AM2(5X), and CAM5) accompany a significant ITCZ shift but with little shift of the energy flux equator. Thus, the response of energy flux equator cannot explain the ITCZ response.
Fig. 4.1. (a) The global-mean SST response (in K) and the global-mean precipitation response normalized by the global-mean precipitation in the respective reference integration and the global-mean SST response (in %/K). (b) The difference of SST response between the NH and the SH ($\delta T_{eq}$) and that relative to the ITCZ location in the respective reference integration ($\delta T_{ITCZ}$). The response indicates the difference between 2xCO2 and 1xCO2 integrations.
Fig. 4.2. The zonal-mean meridional streamfunction ($10^{10}$ kg/s) in the reference integration (contours) and its response to 2xCO2. The contour interval is $2 \times 10^{10}$ kg/s.
Fig. 4.3. The response of the (a) SST and (b) precipitation to 2xCO2. Asterisks indicate the ITCZ location in the reference (1xCO2) integration.
Fig. 4.4. The ITCZ response and the energy flux equator response, with the 90% confidence level in bars.
The ITCZ response is tightly related to the response of low level mass flux $\delta v_2$ at the reference ITCZ location (Fig. 4.5). The models with a northward $\delta v_2$ at the reference ITCZ indicates a ITCZ shift to the north, and vice versa. This is because the precipitation response mostly due to the changes in mean moisture convergence rather than the evaporation response. Eq. (2) indicates that $\delta v_2$ can arise from change in atmospheric energy transport $\delta F_A$ and change in total gross moist stability $\delta \Delta m$. In previous studies, $\delta \Delta m$ was neglected, so that it was assumed that $\delta v_2$ is determined by $\delta F_A$. Hence, the energy flux equator response was used to explain the ITCZ response. We proceed to examine the relative importance of $\delta \Delta m$ in causing $\delta F_A$.

Fig. 4.5. The relationship between the ITCZ response (in degrees) and the change in lower level mass flux at the reference ITCZ location ($\delta v_2$; in kg/s).
The total gross moist stability ($\Delta m$) is calculated by Eq. (1) in Section 3.3. As mentioned before, the variations in $p_m$ have little effect on the estimate of $\Delta m$, so we fixed $p_m$ to be 700 mb for simplicity (Fig. 3.3.5). Because the aim of this study is to explain the ITCZ shift, we performed the analysis at the ITCZ. As stated in Section 3.3.3, $\theta_e$ does not coincide with $\theta_{v_2=0}$, so that $\Delta m$ is negative between $\theta_e$ and $\theta_{v_2=0}$. In all cases except AM2(0), the ITCZ in 1xCO2 is located between $\theta_e$ and $\theta_{v_2=0}$, so that $\Delta m$ at the ITCZ is negative. In AM2(0), the ITCZ is located to the south of $\theta_e$, so that $\Delta m$ at the ITCZ is positive. (Fig. 4.6). The model with a positive $\Delta m$ in 1xCO2 is represented by red color and other cases are represented by blue color in Fig. 4.7. Regardless of the sign of $\Delta m$, a positive $\delta \Delta m$ indicates a larger energy transport increase in the upper-level than in the lower level, and vice versa.

Fig. 4.6. The schematic that explains the sign of total gross moist stability ($\Delta m$) at the ITCZ in the reference (1xCO2) integration. It is sensitive to the relative ITCZ location to the energy flux equator ($\theta_e$).
Now, we decompose the anomalous atmospheric energy transport ($\delta F_A$, black) into the term due to the change in lower level mass flux ($-\delta v_2 \cdot \Delta m$, blue) and that due to the change in total gross moist stability ($-\nabla_2 \delta \Delta m$, red) using Eq. (2) in Section 3.3 (Fig. 4.7). In all cases, $\delta \Delta m$ (red) is more important at determining $\delta F_A$ (black) than $\delta v_2$ (blue), indicating that the assumption of negligible $\delta \Delta m$ in previous studies is not valid in our 2xCO2 experiments. Indeed, the correlation between $\delta F$ and $\delta \Delta m$ is over 0.9 whereas the correlation between $\delta F$ and $\delta v_2$ is only 0.5 (Fig. 4.8). It indicates that the shift of the ITCZ in response to a doubling of CO$_2$ cannot be understood without accounting for the total gross moisture stability change in the tropics.

The sign of total gross moisture stability change at the ITCZ differs across models. In the two models (AM2(1X) and AM2(4X)), there is a reduction in $\Delta m$, and the rest exhibits an increase (refer to Fig. 3.3.5 and Fig. 4.7). As the sign of $\delta \Delta m$ is determined by the ratio between the changes in the total energy transport in the upper-level and that in the lower-level, how the convection responds to 2xCO2 will have an impact. For instance, if the convection becomes deeper, more energy can be transported by the upper-level branch, potentially leading to an increase of $\Delta m$. However, changes in cloud radiative properties may overwhelm the effect of convection deepening. Hence, uncertainties in cloud modeling and convection schemes can cause uncertainties in $\delta \Delta m$, creating a large model spread (Fig. 3.3.5).

Let us first examine the cases with negligible $\delta F_A$: AM2(4X), AM2(5X), CAM5 (Fig. 4.7). In previous studies where $\Delta m$ is held fixed, $\delta F_A = 0$ is thought to cause no shift in the energy flux equator, hence, the ITCZ. However, if there were significant changes in $\Delta m$, which is the case in our set of experiments, $\delta F_A = 0$ can still result in changes $v_2$ and accompany a shift in the ITCZ. In the case of $\delta F_A = 0$, $\delta v_2 = -\frac{\delta \Delta m}{\Delta m} v_2$. Note that $v_2 > 0$ at the ITCZ. The sign of $\delta \Delta m$ is critical for determining the sign of $\delta v_2$. For example, AM2(4X) case has a negative $\Delta m$ at the reference ITCZ and $\delta \Delta m$ is negative, so that $\delta v_2$ is negative, indicating a southward ITCZ shift. In contrast, AM2(5X) and CAM5 exhibit a positive change in $\Delta m$, hence, $\delta v_2$ is positive, indicating a northward ITCZ shift. These relationships are summarized in Fig. 4.10. When $\delta F_A = 0$, changes in $\Delta m$ must be accompanied by changes in $v_2$, causing the ITCZ shift.

Let us now examine the cases with nonzero $\delta F_A$. The AM2(1X) exhibits a significantly positive $\delta F_A$, whereas AM2(0), AM2(0+), AM3 and HiRAM cases exhibit a significantly negative $\delta F$. These differences are thought to arise from the differences in cloud response. For instance, as shown in Fig. 4.9a, AM2(1X) with the largest northward $\delta F_A$ exhibits a substantial warming from the cloud responses in the southern hemisphere’s subtropics relative to the northern hemisphere’s, whereas
AM3 with the largest southward $\delta F_A$ exhibits nearly zero warming in the southern subtropics. This contrast mostly results from differences in the shortwave component of cloud radiative forcing (Fig. 4.9). In all of $\delta F_A \neq 0$ cases, $\delta F_A$ and $\nu_2 \delta m$ have same signs, so that unless their relative magnitudes are known a priori the sign of $\delta \nu_2$ cannot be determined from Eq. (2), unlike the $\delta F_A=0$ case where the sign of $\delta \nu_2$ can be determined solely from the sign of $\delta m$. The signs of $\delta F_A$ (black) and $-\delta \nu_2 \cdot \Delta m$ (blue) are the same in AM2(1X) and AM2(0), indicating that $\delta F_A$ cannot determine the sign of $\delta \nu_2$, as in previous studies. It again illustrates the importance of accounting for the changes in $\Delta m$ to understand the tropical precipitation response to a uniform CO$_2$ radiative forcing.
Fig. 4.7. Decomposition of the anomalous atmospheric energy flux ($\delta F_A$; black), with the range of 90% confidence level, into the term due to the change in lower level mass flux ($-\delta v\cdot \Delta m$; blue) and that due to the change in total gross moist stability ($-v_2\cdot \delta \Delta m$; red). All in the unit of PW. The case with a positive $\Delta m$ in 1xCO2 (red shading) is distinguished with the cases with a negative $\Delta m$ in 1xCO2 (blue shading).
Fig. 4.8. The relationship between the anomalous atmospheric energy flux ($\delta F_A$ in W) and (a) the anomalous lower level mass flux ($\delta v_2$ in kg/s) and (b) the total gross moist stability change ($\delta \Delta m$). The correlation coefficient between the two variables is indicated at the upper right corner of each panel.
Fig. 4.9. The zonal-mean response of (a) total cloud radiative forcing at top-of-atmosphere, (b) shortwave cloud radiative forcing, and (c) longwave cloud radiative forcing. Positive values indicate downward fluxes.
Fig. 4.10. The flow chart that summarizes the sign of lower level mass flux changes can be determined from the sign of changes in atmospheric energy transport and total gross moist stability.
V. Conclusion

In response to increasing CO\(_2\), the globe on average will certainly get warmer and experience more rainfall. However, their pattern changes are subject to large uncertainties despite their importance for regional climate projection, on which the climate adaptation and policy are based. In this study, we investigate how the pattern of tropical precipitation would change in response to a doubling of CO\(_2\) under an idealized setup. If the current climate state were completely symmetric about the equator, no shift in the ITCZ is expected. Hence, among many factors that can contribute to the ITCZ shift, we consider the effect of a hemispheric asymmetry in the climate state. We introduce the hemispheric asymmetry by prescribing a northward ocean transport in the aquaplanet slab ocean, so that the ITCZ is located in NH. Then, the CO\(_2\) concentration is doubled to the reference climate to examine the response to a CO\(_2\) doubling. The same set of experiments is performed with AM2 (with five different convection scheme parameters), AM3, HiRAM, and CAM5 to test the robustness of the results.

As the NH is warmer in the reference state, the positive water vapor feedback causes the NH to become even warmer than the SH in response to a CO\(_2\) doubling. However, although most models show the ITCZ to be shifted further northward, one model shows a southward ITCZ shift. In previous studies, the tropical precipitation shift has been understood based on the atmospheric energy budget. The key assumption was that the total gross moist stability ($\Delta m$), that indicates the total moist static energy transport per unit mass transport, stays constant. Then, the mass transport change can only arise from the cross-equatorial atmospheric energy transport changes or the divergence of equatorial energy input. However, this study reveals that when accounting for the response to a uniform CO\(_2\) radiative forcing, the changes in $\Delta m$ cannot be neglected, but in fact are more dominant over the atmospheric energy transport ($\delta F_A$) change to determine the changes in the mass transport. For instance, in an extreme case with no change in the mean circulation, the changes in $\Delta m$ can lead to the changes in $F_A$. Then, despite $\delta F_A \neq 0$, there will be no or little, if any, ITCZ shift. Let us consider the other extreme case of $\delta F_A = 0$. If there were large changes in $\Delta m$, to keep $F_A$ constant there must be substantial changes in the mass transport, which will be accompanied by a shift in the tropical precipitation. However, previous studies in which $\delta \Delta m$ is neglected will predict no change in the mass transport if $\delta F_A = 0$. Hence, the shift of the energy flux equator, which is the latitude where $F_A$ vanishes, is not always a good indicator for the ITCZ shift.

In many previous studies (Donohoe et al. 2013; Donohoe et al. 2014; McGee et al. 2014; Voigt et al. 2014), the shift of the energy flux equator has been successfully used to infer the ITCZ shift. In most cases, a hemispherically asymmetric forcing was applied to perturb the climate, which brings about a
substantial shift of the Hadley circulation. This is why the effect of $\delta \Delta m$ was negligible compared to the effect of the mass transport changes for accomplishing the cross-equatorial atmospheric energy transport. Otherwise, the uniform radiative forcing induces a rather subtle shift of the Hadley circulation compared to the hemispherically asymmetric forcing. In fact, if the Earth were completely hemispherically symmetric, the uniform forcing would not cause any latitudinal shift. It is the hemispheric asymmetry of the present-day Earth that causes a latitudinal shift in response to uniform radiative forcing. It turns out that the effect of the changes in the mass transport due to the shift of the Hadley circulation is much less than the effect of $\delta \Delta m$ for inducing changes in the atmospheric energy transport in response to increasing CO$_2$. Therefore, the present study addresses the importance of accounting for the changes in the total gross moist stability $\delta \Delta m$ in order to properly understand the response of tropical precipitation to uniform radiative forcing. There, however, are large uncertainties in $\delta \Delta m$ as the vertical structure of moist static energy is affected by convection and clouds. Thus, not only from the local physics point of view but also from the large-scale dynamics point of view, improving convection schemes and cloud modelling is necessary for a more precise projection of tropical precipitation.
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