Sensitivity of intertropical convergence zone movement to the latitudinal position of thermal forcing

Jeongbin Seo, Sarah M. Kang
School of Urban and Environmental Engineering, UNIST,
Ulsan, Korea

Dargan M. W. Frierson
Department of Atmospheric Sciences, University of Washington,
Seattle, Washington

Corresponding author: Sarah M. Kang, School of Urban and Environmental Engineering, Ulsan National Institute of Science and Technology, 100 Banyeon-ri, Eonyang-eup, Ulsan 689-798, South Korea. E-mail: skang@unist.ac.kr
Abstract

A variety of recent studies have shown that extratropical heating anomalies can be remarkably effective at causing meridional shifts in the intertropical convergence zone (ITCZ). But what latitudinal location of forcing is most effective at shifting the ITCZ? In a series of aquaplanet simulations with the GFDL AM2 model, we show that high latitude forcing actually causes a larger shift in the ITCZ than when equivalent surface forcing is applied in the tropics. We run equivalent simulations with an idealized general circulation model (GCM) without cloud- or water vapor-feedbacks, where the ITCZ response instead becomes weaker the farther the forcing is from the equator, indicating radiative feedbacks must be important in AM2.

In the absence of radiative feedbacks, the tendency for anomalies to decrease in importance the farther away they are from the equator is due to the quasi-diffusive nature of energy transports. Cloud shortwave responses in AM2 act to strengthen the ITCZ response to extratropical forcing, amplifying the response as it propagates towards the equator. These results emphasize the great importance of the extratropics in determining the position of the ITCZ.

1. Introduction

Much of the precipitation in the tropics occurs within a narrow zonal band of high rainfall
known as the intertropical convergence zone (ITCZ). Because small changes in the position of the ITCZ can greatly perturb local precipitation, it is important to understand how the ITCZ may respond to external thermal forcing.

Since the ITCZ is often thought to be controlled by tropical mechanisms (e.g., Xie 2004), recent studies that have demonstrated that the ITCZ can respond to heating well outside the tropics have drawn significant attention. For example, Chiang and Bitz (2005) showed that the marine ITCZ shifts away from the hemisphere with increasing high latitude ice cover. Kang et al (2008, hereafter K08), Kang et al (2009) and Yoshimori and Broccoli (2009) interpreted the sensitivity of the ITCZ to extratropical forcings using an energetic framework. Hwang and Frierson (2013) found that the double ITCZ problem of general circulation models (GCMs) is largely caused by cloud biases over the Southern Ocean. Hwang et al (2013) showed that a global southward shift of precipitation in the late 20th century was driven by sulfate aerosols emissions from the Northern Hemisphere (NH) midlatitudes.

The ITCZ has also been shown to shift in response to changes in the surface flux induced by ocean circulation changes. For example, a shutdown of Atlantic Meridional Overturning Circulation (MOC) results in a large decrease in the ocean-to-atmosphere flux in the high latitudes of the NH, and causes a southward ITCZ shift in coupled GCMs (Zhang and Delworth 2005, Stouffer et al. 2006). Fučkar et al (2013) studied an idealized coupled GCM, and found that the direction of cross-equatorial heat transport by MOC controls the ITCZ location. Frierson et al (2013) built upon this argument, showing that the ocean MOC is the primary reason that the ITCZ exists in the NH in the present climate.

The large sensitivity of the ITCZ to surface heat flux anomalies even in the high latitudes begs the question of what meridional locations are most effective at shifting the ITCZ. In this
paper, we study this question by forcing aquaplanet GCMs with surface heating anomalies at different latitudes, and studying the response of the ITCZ.

2. Data and Methods

We employ two aquaplanet GCMs with different levels of complexity. One is a comprehensive atmospheric GCM developed at the Geophysical Fluid Dynamics Laboratory (GFDL), AM2 (Anderson et al. 2004). The model uses a horizontal resolution of $2^\circ$ latitude $\times$ $2.5^\circ$ longitude with 24 vertical levels, and is run under equinox conditions. The other is the gray radiation moist (GRaM) GCM (Frierson et al. 2006) in which the radiative fluxes are only a function of temperature. Hence, there are no water vapor- or cloud-radiative feedbacks. It has T42 horizontal resolution with 25 vertical levels. Both models are coupled to an aquaplanet slab mixed layer ocean of 2.4m depth. The shallow mixed layer is used to shorten the time to reach equilibrium. The models are run for 8 years with the results shown averaged over the last 6 years.

The experiments are designed to examine the efficiency of surface thermal forcing on shifting the ITCZ depending on its meridional position. The control climate with no flux adjustment is perturbed by prescribing heating in the mixed layer in the NH and cooling of equal magnitude in the SH (Fig. 1a). The meridional position of thermal forcing ($H$) with $10^\circ$ latitudinal width is systematically varied as follows:

$$
H = -A \sin \left( \frac{\pi}{10} (\theta + \theta_0) \right) \quad \text{for} \quad -\theta_0 < \theta < -\theta_0 + 10, \\
H = -A \sin \left( \frac{\pi}{10} (\theta - \theta_0) \right) \quad \text{for} \quad \theta_0 - 10 < \theta < \theta_0, \\
H = 0 \quad \text{elsewhere.}
$$

(1)
Since the global mean of $H$ is zero, it can be described in terms of an implied meridional oceanic heat transport $F_o$, with $H = -\frac{1}{2\pi a^2 \cos \theta} \frac{\partial F_o}{\partial \theta}$, as shown in Fig. 1b. In all of the experiments, $F_o$ is positive, equivalent to a south-to-north OHT. We vary $\theta_0$ from 10° to 80° while adjusting the maximum amplitude of the forcing $A$ to ensure that the cross-equatorial flux $F_o$ is fixed to 1.8PW. This also ensures that the heating in the NH (and cooling in the SH) has the same area-integrated value in each case. The sensitivity of the results to the magnitude of the thermal forcing is examined by performing the same set of experiments with half the amplitude, equivalent to the cross-equatorial flux of 0.9PW. Our results are qualitatively insensitive to $A$, so for brevity, all the figures except Fig. 3 are for $F_o = 1.8$PW.

3. Results

The zonal-mean precipitation between 30°S-30°N in both models is shown in Fig. 2a, b. The tropical precipitation shifts toward the warmer NH not only in response to tropical thermal forcing but also to extratropical thermal forcing, consistent with previous studies (e.g. Broccoli et al. 2006; K08). This behavior is different in the two models however. In GRaM, tropical precipitation is perturbed more effectively by tropical thermal forcing (Fig. 2a), while AM2 is perturbed more effectively by extratropical thermal forcing (Fig. 2b). This stark difference is better shown in Fig. 3a, which shows the latitude of the ITCZ for varying $\theta_0$ in both models. The largest displacement of the ITCZ occurs when $\theta_0=10$ in GRaM and when $\theta_0=60$ in AM2. Fig. 3a also shows that the displacement of the ITCZ is larger in AM2 than in GRaM for all cases, a fact that was previously attributed to the presence of water vapor and
cloud feedbacks in AM2 in Kang et al (2009).

The ITCZ response is closely linked to the response of the cross-equatorial atmospheric energy transport (K08). The energy budget for the atmospheric column in steady-state is

\[ R_{TOA} - \nabla \cdot F_o = \nabla \cdot F \]

where \( R_{TOA} \) is the top-of-atmosphere net radiation and \( F = [m] \) is the zonal and time mean meridional transport of moist static energy \( m \). A change in cross-equatorial atmospheric energy transport (AET) is typically accomplished by an anomalous Hadley circulation, which also transports moisture across the equator in the opposite direction. We express the change in deep tropical AET as a compensation \( C \) of the imposed oceanic flux by the AET, i.e.,

\[ C \equiv \frac{(F - F_{ctl})}{F_o} \]

Fig. 3b shows the degree of compensation \( C \) averaged between 5°S and 5°N. In GRaM, \( C \) becomes smaller as the external thermal forcing is located farther away from the tropics, consistent with the smaller ITCZ response. In the tropics, thermal forcing is redistributed efficiently by the Hadley circulation so as to maintain small tropical temperature gradients (e.g. Sobel et al. 2001, Yano and Bonazzola 2009). Hence, \( C \) is over 60% for cases with \( \theta_0 < 30 \). The farther away the thermal forcing is from the equator, the less the tropical AET response, consistent with a quasi-diffusive response of energy that is fluxed to space as it spreads equatorward, as we describe later. In contrast, \( C \) in AM2 becomes larger as the external thermal forcing is located outside of the tropics, consistent with the displacement of the ITCZ. When AM2 is compared with GRaM, one can find that \( C \) is much greater in AM2 for all cases with one exception when \( \theta_0 = 10 \).

To understand the dependence of \( C \) on \( \theta_0 \), we express \( R_{TOA} \) as the divergence of a flux \( F_{TOA} \) such that \( R_{TOA} = -\nabla \cdot \delta F_{TOA} \). Further, after subtracting the global mean, \( F_{TOA} \) can be divided into a clear-sky component, \( F_{clr} \), and a component from cloud radiative forcing (CRF), \( F_{CRF} \), i.e.,

\[ F_{TOA} = F_{clr} + F_{CRF} \]

with \( CRF = -\nabla \cdot \delta F_{CRF} \). Then the anomalous energy
budget, with the global mean removed, can be represented as

\[-\nabla \cdot \delta F_{\text{clr}} - \nabla \cdot \delta F_{\text{CRF}} - \nabla \cdot \delta F_{\alpha} = \nabla \cdot \delta F\]

(2)

where \(\delta\) denotes the difference from the control run \((A=0)\). Each term in Eq. (2) for the cases with \(\theta_0 = 10\) (tropical forcing) and 80 (extratropical forcing) are shown in Fig. 4 for GRaM (upper) and AM2 (lower). As stated earlier, since large temperature gradients cannot be sustained in the tropics, the tropical thermal forcing in GRaM, where \(R_{\text{TOA}}\) is a function of temperature only, is mostly compensated by atmospheric energy transport divergence with only a small change in TOA radiative fluxes (Fig. 4a). That is, there is large compensation \(C\) in response to tropical thermal forcing. In contrast, extratropical thermal forcing induces large temperature changes (Fig. 2c) and large meridional gradients of \(\delta F_{\text{TOA}}\) in the extratropics (Fig. 4b). Locally, the imposed warming over 70°-80°N is balanced mostly by the divergence of atmospheric energy transport (\(\nabla \cdot \delta F_A\) is large positive) and partly by increasing outgoing longwave radiation (OLR) via warming the atmosphere (\(\nabla \cdot \delta F_{\text{TOA}}\) is modestly positive). The atmospheric column outside of the forced region (e.g., 20°-70°N) is warmed by quasi-diffusive AET, but loses energy by increased OLR from the warmer atmosphere. That is, \(\nabla \cdot \delta F_A < 0\) and \(\nabla \cdot \delta F_{\text{TOA}} > 0\) in order to have the sum \(\delta F_A + \delta F_{\text{TOA}}\) meridionally flat. Due to weak tropical temperature gradients, \(\delta F_{\text{TOA}}\), which is a function of temperature only in GRaM, must be flat within the deep tropics (20°S-20°N), and so is \(\delta F_A\). In other words, as the NH subtropics become warmer and the SH subtropics become colder, the Hadley circulation tries to remove this inter-hemispheric temperature contrast in the tropics by transporting energy toward the colder SH. Hence, the compensation \(C\) in the tropics is determined by \(\delta F_A\) at the poleward edge of the Hadley circulation. The magnitude of the signal reaching the subtropics is controlled by the quasi-diffusive eddy energy fluxes, and is
thus larger when the forcing is closer to the tropics. This explains why the compensation $C$, and accordingly the ITCZ response, is greater in response to tropical thermal forcing than extratropical thermal forcing in GRaM.

By comparing Figs. 4b and 4d, one can see that changes in CRF in AM2 act to amplify the imposed extratropical thermal forcing by more than 100% in the tropics. This is because, as shown in Fig. 5, the extratropical thermal forcing increases (decreases) low-level cloud amount in the cooled (warmed) region, possibly via (de)stabilization of the lower troposphere, leading to more cooling (warming) through the feedback of cloud forcing on TOA energy fluxes. Extratropical cloud responses that amplify the effective strength of the extratropical thermal forcing cause the compensation $C$ and the ITCZ response to be significantly larger in AM2 than in GRaM. In particular, the imposed thermal forcing is overcompensated by the AET when $\theta_0 > 20$. (Fig. 3).

In contrast to the extratropical cloud responses, the tropical cloud responses associated with the ITCZ shift acts to counteract the imposed thermal forcing in AM2: the changes in CRF in all cases in Fig. 5a exhibit cooling (warming) in the NH (SH) tropics. This is because, in AM2, the shortwave forcing following the ITCZ shift is larger than the longwave forcing associated with high cloud amount changes (Kang et al. 2013). Hence, when the thermal forcing is located in the tropics (Fig. 4c), cloud effects act to counteract the imposed forcing. Because of these opposing effects of cloud responses on altering the effective strength of the imposed forcing in the tropical and extratropical forcing cases, the compensation $C$ in the tropics in AM2 becomes larger as the forcing is located farther away from the equator. If only these cloud effects were to cause the difference between the tropical and extratropical forcing cases, the difference of $C$ between the two cases should have been much greater than that
shown in Fig. 3. For instance, the cloud effects of the case with $\theta_0=80$ amplify the effective
strength of imposed forcing by 2.7PW more compared with $\theta_0=10$ case, which would result
in 150% larger $C$, but in fact $C$ is larger by 50%. This is because larger changes in the clear-
sky radiative fluxes have an opposing effect on compensation in the extratropical forcing
cases (Fig. 4d), which we investigate next.

The lower panels in Fig. 2 show that the sea surface temperature (SST) response becomes
larger as the external thermal forcing is imposed farther away from the tropics. In particular,
the SST response in the forced region is 30 (10) times larger when $\theta_0 = 80$ than when
$\theta_0 = 10$ in AM2 (GRaM). Also, in both models, the SST response is approximately twice as
large in the cooled region than in the warmed region. The stronger SST response in the higher
latitude forcing cases is in part, but not entirely, due to the stronger local forcing applied in
order to keep the area-weighted heating the same. Differences in lapse rate feedback also
contribute. In the tropics, a small increase in SST induces a large increase in upper
tropospheric temperature and therefore OLR because moist convection in the tropics keeps
the atmospheric temperature close to a moist adiabat. Hence, tropical thermal forcing can be
effectively balanced by small changes in SST. However, in the extratropics, where the
temperature response is concentrated in the lower troposphere, thermal forcing requires a
large increase in SST in order to have sufficient increase in OLR. Moreover, in high latitude
regions with higher static stability, the temperature response to cooling is even more surface
trapped, so that the SST response to cooling is larger than to warming. This dependence of
temperature response on the meridional position of external thermal forcing was recently
discussed in Kang and Xie (2013). Large temperature changes in the extratropical forcing
case cause the clear-sky radiative fluxes to act as a limiter on the ITCZ response, diminishing
the positive feedback from cloud responses (Fig. 4d). This effect is especially significant when the thermal forcing is imposed in the polar regions \((\theta_0 > 60)\), so that the displacement of the ITCZ in AM2 slightly decreases when \(\theta_0 > 60\).

4. Conclusion

In this study, we examine the efficiency of surface thermal forcing in different latitudinal location on shifting the ITCZ using two aquaplanet GCMs, the comprehensive GFDL AM2 and an idealized moist GCM, GRaM. In GRaM, the displacement of the ITCZ is larger when the thermal forcing is located closer to the tropics, because the impact of the thermal forcing outside the tropics diminishes on its way equatorward by quasi-diffusive transport of energy. In contrast, in AM2, extratropical thermal forcing can shift the ITCZ even more than tropical thermal forcing. In AM2, low cloud responses substantially amplify the effective strength of extratropical thermal forcing. However, extratropical thermal forcing causes local temperature to change much more compared with tropical thermal forcing, thereby enhancing the response of clear-sky radiative fluxes. In AM2, the positive feedback from the cloud effects overwhelm the negative feedback from clear-sky radiative fluxes, hence, extratropical forcing is more effective at shifting the ITCZ than tropical forcing.

The results from AM2 are the applicable ones to Earth’s climate; the simulations with GRaM were run to understand the role of dynamical processes versus radiative feedbacks. Thus our study suggests that extratropical factors can be even more important than local tropical processes in displacing the ITCZ. The strong influence of the extratropics on tropical precipitation can be seen in CMIP3 simulations of global warming (Frierson and Hwang
The importance of cloud feedbacks, which may be model-dependent, means that our results should be tested in other GCMs. Also, it is important to note that despite the importance of extratropical forcing in shifting the zonal mean ITCZ, local tropical forcing is much more effective at causing zonal asymmetries in tropical precipitation (Kang et al 2013).

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