Climate Effects of Anthropogenic Aerosol Forcing on Tropical Precipitation and Circulations[✉]

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ABSTRACT

Aerosols are one of the key factors influencing the hydrological cycle and radiation balance of the climate system. Although most aerosols deposit near their sources, the induced cooling effect is on a global scale and can influence the tropical atmosphere through slow processes, such as air–sea interactions. This study analyzes several simulations of fully coupled atmosphere–ocean climate models under the influence of anthropogenic aerosols, with the concentrations of greenhouse gases kept constant. In the cooling simulations, precipitation is reduced in deep convective areas but increased around the edges of convective areas, which is opposite to the “rich-get-richer” phenomenon in global warming scenarios in the first-order approximation. Tropical convection is intensified with a shallower depth, and tropical circulations are enhanced. The anomalous gross moist stability (M') mechanism and the upped-ante mechanism can be used to explain the dynamic and thermodynamic processes in the changes in tropical precipitation and convection. There is a northward cross-equatorial energy transport due to the cooler Northern Hemisphere in most of the simulations, together with the southward shift of the intertropical convergence zone (ITCZ) and the enhancement of the Hadley circulation. The enhancement of the Hadley circulation is more consistent between models than the changes of the Walker circulation. The change in the Hadley circulation is not as negligible as in the warming cases in previous studies, which supports the consistency of the ITCZ shift in cooling simulations.

1. Introduction

The impact of anthropogenic aerosols on climate usually refers to the direct and indirect effects of aerosols, which have been extensively investigated (e.g., Lohmann and Lesins 2002; Penner et al. 2004). In general, aerosols decrease the temperature of the atmosphere and Earth's surface by reflecting solar radiation, except black carbon, which absorbs solar radiation (the direct effect). This process may influence the convection strength by modifying the vertical temperature structure

(Kaufman et al. 2002). Aerosols also change cloud optical properties and lifetime through aerosol indirect effects (Ramanathan et al. 2001; Feichter et al. 2004; Liepert et al. 2004). Both direct and indirect effects generally occur in a small region near aerosol sources and on a shorter time scale than the warming effects of greenhouse gases (GHGs). Determining how GHGs and anthropogenic aerosols influence Earth's climate has been a challenge scientists facing for several decades. Several studies have emphasized that the combined effects of warming and cooling processes may involve complex nonlinear interactions (e.g., IPCC 2001; Rotstajn and Lohmann 2002; Feichter et al. 2004; McFarlane and Frierson 2017; Acosta Navarro et al. 2017; Rotstajn et al. 2015). However, the individual processes must be examined in more detail before combining the warming and cooling effects.

Aerosol-induced temperature perturbation can be distributed globally by atmospheric waves and causes cooling on a global scale. The spatial patterns of

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temperature and precipitation anomalies are similar between GHG-induced warming and aerosol-induced cooling (Chou et al. 2005; Xie et al. 2013; Wang et al. 2016) with an opposite sign. For example, anthropogenic aerosols may induce global-scale cooling that has a larger amplitude at middle to high latitudes in the Northern Hemisphere than it does in the Southern Hemisphere. Model simulations also show that the aerosol-induced temperature perturbation has significant impacts on the position of the ITCZ and the cross-equatorial energy flux (Rotstayn et al. 2015; Acosta Navarro et al. 2017), which have opposite directions to the global warming cases. Chou et al. (2005) conducted model experiments using the quasi-equilibrium tropical circulation model (QTCM1) to demonstrate how the aerosols affect tropical precipitation. They suggested that the changes in precipitation between scenarios of GHG warming and aerosol cooling can be explained by the same mechanisms—namely, the anomalous gross moist stability (M') mechanism and the upped-ante mechanism (Neelin et al. 2003; Chou and Neelin 2004). Xie et al. (2013) focused on the role of the ocean and suggested that GHGs and aerosols can both influence climate through processes involving the ocean, which also explains the similarity between the two scenarios.

Recent studies have divided precipitation responses induced by anthropogenic forcings, including GHGs and aerosols, into fast and slow responses (i.e., Andrews et al. 2010; Wang et al. 2016; Myhre et al. 2017). Andrews et al. (2010) and Myhre et al. (2017) showed that the fast response is associated with atmospheric absorption change due to GHGs and the aerosol direct effect, mostly from black carbon. The slow response is associated with surface temperature change, which appears to have contributions from GHGs and sulfate aerosols. The slow response is suggested to involve ocean processes and may be the same as the sea surface temperature (SST) mediation effect mentioned by Wang et al. (2016).

Other than spatial distribution of precipitation, global warming affects several tropical properties. For example, Vecchi and Soden (2007) showed that tropical large-scale circulations, especially the Walker circulation, are weakened in global warming scenarios. This weakening can explain the difference between the rate of precipitation change ($\sim 1\% - 2\% \text{ K}^{-1}$) and the rate of moisture change ($\sim 7\% \text{ K}^{-1}$). The depth of the troposphere increases, which enables tropical convection to reach a higher altitude. At the same time, the vertical motion is weakened because a deeper convection depth leads to larger gross moist stability (Chou and Chen 2010). The position of the ITCZ shifts

northward with increased southward energy transport across the equator (Frierson and Hwang 2012). Among these properties, the hemispheric climate shift or ITCZ shift due to anthropogenic aerosols has been investigated (Hwang et al. 2013; Chung and Soden 2017). Hwang et al. (2013) found that the ITCZ shifted southward during the latter half of the twentieth century, and they attributed this shift to aerosol indirect effects using observations and historical runs from phases 3 and 5 of the Coupled Model Intercomparison Project (CMIP3 and CMIP5). However, it is difficult to isolate the influences of aerosols from other climate forcings in their study. Chung and Soden (2017) suggested that the interhemispheric temperature change and ITCZ shift are more closely related to the aerosol-mediated cloud response than the instantaneous aerosol forcing (i.e., shortwave forcing by aerosols). Both Wang et al. (2016) and Chung and Soden (2017) suggested that the interhemispheric asymmetric response is caused by aerosol forcing and a symmetric response is caused by GHG forcing.

Previous studies have shown the importance of ocean and aerosol–cloud interactions on influencing the global temperature distribution, tropical precipitation, and circulations. Our study focuses on understanding the dynamic and thermodynamic mechanisms through which the precipitation and circulations response to a cooler temperature. The goals of the present study are 1) to investigate atmospheric responses to the cooling forced by anthropogenic aerosols with constant GHG concentrations in CMIP5 models and 2) to examine the mechanisms responsible for precipitation changes in a cooling atmosphere. The major difference between this study and Chou et al. (2005) is that they forced QTCM1 with an aerosol-associated solar forcing produced by a general circulation model (ECHAM4; Roeckner et al. 1995) instead of imposing aerosols in the model, meaning that the aerosol–cloud interaction was not considered. In this study, several fully coupled atmosphere–ocean general circulation models were forced by anthropogenic aerosols and compared with preindustrial control runs to demonstrate the long-term effects on tropical precipitation and circulation. Section 2 briefly introduces the model simulations and the diagnostic methods used in this study. Section 3 shows the changes in the temperature and precipitation spatial patterns due to the aerosol-cooling effect, and section 4 discusses the validation of the existing mechanisms. Section 5 discusses the connection of the ITCZ shift and large-scale circulation changes, as well as the role of the ocean. Conclusions and final remarks are presented in section 6.

TABLE 1. CMIP5 simulations used in this research and their forcing agents (AA: anthropogenic aerosol without specific details; SO₄: sulfate aerosols; BC: black carbon, OC: organic carbon).

Model	Organization and country	Forcing	Driven by	Original resolutions
CanESM2	Canadian Centre for Climate Modelling and Analysis (Canada)	AA	Emission	2.812° × 2.812°, L22, 1850–2005
CCSM4	National Center for Atmospheric Research (United States)	AA	Concentration	0.9375 × 1.25, L17, 1850–2005
CESM1	National Center for Atmospheric Research (United States)	AA, consider BC on snow	Emission	2° × 2.5°, L30, 1850–2005
GFDL CM3	Geophysical Fluid Dynamics Laboratory (United States)	BC, OC, SO ₄	Emission	2° × 2.5°, L23, 1860–2005
GISS-E2-H (r1i1p107)	NASA Goddard Institute for Space Studies (United States)	BC, OC, SO ₄ (also includes nitrates)	Concentration	2° × 2.5°, L17, 1850–2005
GISS-E2-R (r1i1p107)	NASA Goddard Institute for Space Studies (United States)	BC, OC, SO ₄ (also includes nitrates)	Concentration	2° × 2.5°, L17, 1850–2005
GISS-E2-H (r1i1p310)	NASA Goddard Institute for Space Studies (United States)	BC, OC, SO ₄ (also includes nitrates), consider BC on snow	Emission	2° × 2.5°, L17, 1850–2005
GISS-E2-R (r1i1p310)	NASA Goddard Institute for Space Studies (United States)	BC, OC, SO ₄ (also includes nitrates), consider BC on snow	Emission	2° × 2.5°, L17, 1850–2005
IPSL-CM5A-LR	Institut Pierre Simon Laplace (France)	AA	Emission	1.875° × 3.75°, L17, 1850–2005
NorESM1-M	Norwegian Climate Centre (Norway)	BC, OC, SO ₄ , consider BC on snow	Emission	1.875° × 2.5°, L17, 1850–2005

2. Data and methodology

a. CMIP5 model simulations

CMIP5 provides several types of forced experiments in the “historicalMisc” category for the period 1850–2005. To demonstrate the influence of anthropogenic aerosols, we employ 10 simulations (listed in Table 1) that are forced by historical changes in anthropogenic aerosols with the GHGs fixed at the preindustrial level, and we compare them with their preindustrial control run for each model. Some models are driven by historical aerosol emissions and others by historical aerosol concentrations in the atmosphere. The anthropogenic aerosols have slightly different combinations between models: some models are forced by sulfate aerosols (SA; including direct and indirect effects), organic carbon (OC), and black carbon (BC), whereas some are simply forced by anthropogenic aerosols without specifying the details (AA). We also perform an anthropogenic aerosol-forced simulation using the Community Earth System Model (CESM1) at the Research Center of Environmental Changes in Academia Sinica, Taiwan (Table 1), to compare with other simulations. Four of the simulations considered the effects of BC on snow; however, the influences from BC are not detectable.

All model outputs are regridded to 90 × 144 grids (2° × 2.5°), and 17 vertical levels (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10 hPa) are selected. Because the atmospheric responses induced by aerosol forcing may be weak compared with model internal variabilities, we compute the monthly climatology using the last 80 years (out of the 155-yr historical simulations; i.e., 1926–2005) to remove internal variabilities. We examine the pattern correlations of long-term climatology between the preindustrial control run and the aerosol-forcing run using different lengths of simulation years, and we find that the correlation exceeded 0.7 when using the 80-yr climatology. In addition, the Student's *t* test is used to examine the significance of temperature and precipitation anomalies. For temperature anomaly, all grids (total grid number: 12 960) pass the 95% significance test as the average period extended longer than 30 years. For precipitation anomaly, the number of grids passing the 95% significance test reaches the maximum (8902 grids) when using an 80-yr period.

All simulations and their ensemble means are analyzed using the methods described herein. The four simulations from GISS are averaged first to obtain a single-model ensemble for this model. Then, all models are straight averaged to obtain the multimodel ensemble (MME).

b. ITCZ position and cross-equatorial energy transport

We follow the definition given by Frierson and Hwang (2012) to calculate the ITCZ position, which is the centroid of the area-integrated precipitation from 15°S to 15°N. The cross-equatorial energy transport $F_A(\varphi = 0)$ is computed as

$$F_A(\varphi = 0) = \int_{-\pi/2}^0 \int_0^{2\pi} F^{\text{net}} a^2 \cos\varphi d\lambda d\varphi, \quad (1)$$

where φ is the latitude, λ the longitude, and a the radius of Earth. The net energy flux into the atmospheric column is $F^{\text{net}} = \text{LH} + \text{SH} + \text{LW} + \text{SW}$, where LH and SH represent surface latent and sensible heat fluxes, respectively, and LW and SW represent column net longwave and net shortwave radiative fluxes, respectively.

c. Convective mass flux

Changes in tropical circulations in terms of strength and depth are widely discussed in global warming scenarios (e.g., Vecchi and Soden 2007; Chou and Chen 2010). Here, we examine the same features in the cooling scenario by calculating the change rate of convective mass flux M_c

$$\frac{M'_c}{\overline{M}_c} = \frac{P'}{\overline{P}} - 0.07T', \quad (2)$$

where P and T represent precipitation and surface air temperature, respectively. The number 0.07 is the rate of saturation moisture change per kelvin, as estimated from the Clausius–Clapeyron equation. The overbar indicates the preindustrial control simulation, and the prime indicates the difference between aerosol-forcing and control simulations. The rate of change of M_c is normalized by global mean surface temperature anomaly following the method in Vecchi and Soden (2007).

d. Diagnosis of moisture budget and moist static energy budget

Among the mechanisms proposed by Chou and Neelin (2004), the M' mechanism and the upped-ante mechanism could be responsible for modifying the distribution of tropical precipitation. The processes involved in these two mechanisms can be decomposed by examining the moisture budget and moist static energy (MSE) budget. Similar derivatives can be found in Chou and Neelin (2004) and Chou et al. (2005, 2006, 2009). Considering a column atmosphere, the vertical integration of moisture and temperature equations in the energy flux form can be written as

$$\partial_t \langle q \rangle + \langle \mathbf{v} \cdot \nabla q \rangle + \langle \omega \partial_p q \rangle = \langle Q_q \rangle + \text{LH}, \quad (3)$$

$$\partial_t \langle T \rangle + \langle \mathbf{v} \cdot \nabla T \rangle + \langle \omega \partial_p s \rangle = \langle Q_c \rangle + \text{LW} + \text{SW} + \text{SH}, \quad (4)$$

where q is specific humidity with L absorbed and T the temperature with C_p absorbed; L and C_p are the latent heat of evaporation ($2.5 \times 10^6 \text{ J kg}^{-1}$) and specific heat of dry air at constant pressure ($1004 \text{ J kg}^{-1} \text{ K}^{-1}$), respectively, and $\mathbf{v}$ and ω are horizontal and vertical winds, respectively. The dry static energy s is defined as $s = T + gz$, where g is gravity and z the geopotential height. The terms $\langle Q_q \rangle$ and $\langle Q_c \rangle$ are moisture sink and convective heating, respectively. The vertical integral, denoted by angle brackets, is a mass integration through the depth of troposphere p_T from the surface pressure p_s

$$\langle X \rangle = g^{-1} \int_{p_s}^{p_s - p_T} X dp. \quad (5)$$

Assuming relative humidity is constant, the amount of moisture removed from the atmosphere $-\langle Q_q \rangle$ is equal to precipitation P ; $-\langle Q_q \rangle$ is also equal to $\langle Q_c \rangle$ if we assume that the latent heat released by condensation all contributes to increasing the atmospheric temperature. Because the climate system is in a steady state, the time-dependent term in (3) can be neglected, and the moisture budget can be shown as

$$P = -\langle \omega \partial_p q \rangle - \langle \mathbf{v} \cdot \nabla q \rangle + \text{LH}, \quad (6)$$

where $-\langle \omega \partial_p q \rangle$ is analogous to $M_q \nabla \cdot \mathbf{v}_1$; M_q is defined as the gross moisture stratification and $\mathbf{v}_1$ is the first baroclinic mode of the horizontal wind [see Chou et al. (2005) and references therein for details]. In short, this term represents the overall column moisture convergence due to humid low-level inflow and dry upper-level divergent flow. The second and third terms on the right-hand side represent moisture supply from horizontal advection and surface evaporation, respectively.

The same steps can be applied to the temperature equation, (4), and shown as follows:

$$\langle Q_c \rangle = \langle \omega \partial_p s \rangle + \langle \mathbf{v} \cdot \nabla T \rangle - \text{LW} - \text{SW} - \text{SH}. \quad (7)$$

The term $\langle \omega \partial_p s \rangle$ is also analogous to $M_s \nabla \cdot \mathbf{v}_1$, and M_s is gross dry stability. Combining (6) and (7), we obtain the MSE equation:

$$\langle \omega \partial_p h \rangle + \langle \mathbf{v} \cdot \nabla (T + q) \rangle = F^{\text{net}}, \quad (8)$$

where $h = q + s$ is MSE. Similar to the moisture equation, $\langle \omega \partial_p h \rangle$ is analogous to $M \nabla \cdot \mathbf{v}_1$; where M is gross moist stability and equals to $M_s - M_q$.

The contribution of precipitation anomaly P' can be decomposed using the linearized form of the moisture budget:

$$P' = -\langle \omega' \partial_p \bar{q} \rangle - \langle \bar{\omega} \partial_p q' \rangle - \langle \bar{\mathbf{v}} \cdot \nabla q' \rangle - \langle \mathbf{v}' \cdot \nabla \bar{q} \rangle + \text{LH}' + \text{Res}_q, \quad (9)$$

where the overbar indicates the preindustrial control simulation, and the prime indicates the difference between aerosol-forcing and control simulations. As shown in section 4, the most dominant term for precipitation anomaly is $-\langle \omega' \partial_p \bar{q} \rangle$, which represents a dynamic feedback process associated with anomalous vertical motion ω' . Later in the MSE budget, we show that a couple of factors contribute to the vertical motion anomaly, creating a feedback process that produces a large precipitation anomaly. The second term on the right-hand side represents a direct moisture effect due to global temperature change, which is named “the M'_q effect” (Chou and Neelin 2004). The third and fourth terms represent contributions from anomalous horizontal moisture distribution (i.e., the upped-ante mechanism) and from the anomalous wind mechanism (Neelin and Su 2005), respectively. The final two are the surface latent heat anomaly and residual terms, with the latter including nonlinear components as well as errors from using the finite-difference method.

Because the dynamic feedback process is the most dominant term in the moisture budget (Chou et al. 2005), we use the MSE equation to discuss processes that may contribute to anomalous vertical motion ω' . The linearized form of the MSE equation can be shown as follows:

$$\langle \omega' \partial_p \bar{h} \rangle = -\langle \bar{\omega} \partial_p h' \rangle - \langle \bar{\mathbf{v}} \cdot \nabla (T + q)' \rangle - \langle \mathbf{v}' \cdot \nabla (\bar{T} + \bar{q}) \rangle + F^{\text{net}'} + \text{Res}_h. \quad (10)$$

As aforementioned, $\langle \omega \partial_p h \rangle$ is analogous to $M \nabla \cdot \mathbf{v}_1$, and $\langle \omega' \partial_p \bar{h} \rangle$ on the left-hand side is analogous to $M \nabla \cdot \mathbf{v}'_1$. Similarly, $-\omega' \partial_p \bar{q}$ in (9) is analogous to $M_q \nabla \cdot \mathbf{v}'_1$. We can substitute the dynamic terms in (9) and (10) with their analogous forms and then combine them to eliminate $\nabla \cdot \mathbf{v}'_1$. Finally, the precipitation anomaly can be estimated as follows:

$$P' = -\frac{\bar{M}_q}{\bar{M}} \langle \bar{\omega} \partial_p h' \rangle - \frac{\bar{M}_q}{\bar{M}} \{ \langle \bar{\mathbf{v}} \cdot \nabla (T + q)' \rangle + \langle \mathbf{v}' \cdot \nabla (\bar{T} + \bar{q}) \rangle \} + \frac{\bar{M}_q}{\bar{M}} F^{\text{net}'} - \langle \bar{\omega} \partial_p q' \rangle - \langle \bar{\mathbf{v}} \cdot \nabla q' \rangle - \langle \mathbf{v}' \cdot \nabla \bar{q} \rangle + \text{LH}' + \text{Res}. \quad (11)$$

The first term on the right-hand side is referred to as the M' mechanism, with a multiplier $\bar{M}_q/\bar{M}$ that enhances the contribution of the M' mechanism by a factor of

approximately 4 (Yu et al. 1998). The second term is a horizontal advection term with the upped-ante mechanism involved, and it is also enhanced by the multiplier. Together with the $F^{\text{net}'}$ term, these three terms all contribute to the anomalous vertical motion in (10), and the vertical motion feeds back to these processes to further enhance the precipitation anomaly, as shown in (11). The final term (Res) is a residual term.

Briefly, the M' mechanism dominates in the center of the convective region where low-level moisture increases due to GHG warming based on the quasi-equilibrium mediation (Neelin and Su 2005). Convection is enhanced at where gross moisture stability M decreases, and produces more precipitation. This is the “rich-get-richer” mechanism (Chou and Neelin 2004). However, at the margin of convective regions, the warmer air requires more moisture to become saturated and unstable (i.e., to increase the convective ante), but the low-level inflow from the climatic subsidence area is dryer than before. Thus, the convection at the margin is reduced (Chou and Neelin 2004; Chou et al. 2006, 2009).

3. Tropospheric temperature and precipitation anomaly

Figure 1 shows the MME of aerosol loads, tropospheric temperature, and precipitation anomalies. The aerosol loads are ensembles using six simulations because some of the simulations are no longer available from the CMIP5 data port; however, the aerosol spatial distributions mainly depend on historical data and can be expected to be similar in all simulations. Figure 1a shows that the values of aerosol loads are larger over Northern Hemisphere continents, such as the Eurasian continent, Africa, and South and East Asia, and the loads are much lower over tropical oceans. The temperature anomaly (Fig. 1b) is the vertical average weighted by pressure thickness through the whole troposphere (from 1000 to 100 hPa). The ensemble of temperature anomaly shows that the cooling phenomenon is on a global scale with greater amplitude in the Northern Hemisphere than in the Southern Hemisphere, likely due to the locations of aerosol sources. The aerosol-induced temperature perturbation is propagated by atmospheric waves around the globe while aerosols are removed from the atmosphere quickly by dry and wet deposition.

Precipitation decreases mostly in the tropics, especially in the deep convective areas (Fig. 1c, grids with insignificant precipitation change are masked in white), such as the ITCZ, South Pacific convergence zone (SPCZ), north Indian Ocean, and rain forests.

FIG. 1. MME means for (a) aerosol load anomaly (mg m^{-2}), (b) tropospheric temperature anomaly (K) with the global mean value labeled in the upper-left corner, and (c) precipitation anomaly (mm day^{-1}) overlaid by the climatology contour of 4 mm day^{-1} (red contour). Anomalies that do not pass the 95% significance test are masked in white.

Precipitation increases in the tropical southeastern Pacific, which is partially due to the double-ITCZ bias in the CESM1 (Wang et al. 2015). The double-ITCZ bias is also shown in CanESM2 and GFDL-CM3 (Figs. S1a and S4a in the online supplemental material) with weaker amplitudes than the one in the CESM1 (Fig. S3a). Precipitation also increases in the south Indian Ocean, South America, and the tropical

FIG. 2. Meridional distribution of zonally averaged (a) tropospheric temperature anomaly and (b) precipitation anomaly for each model (colors and line patterns) and for MME (black). Four GISS simulations are averaged.

South Atlantic Ocean, right on the edges of convective areas in the control run (indicated by the red contour). This presents a reversed change in the precipitation spatial pattern in a global cooling scenario (decreasing in the tropics but increasing elsewhere), compared with global warming. Some positive anomalies exist over subtropical regions and near the south of Greenland, but the amplitudes are weaker than those in the tropics. The ensemble precipitation anomaly exhibits a clear tendency to southward shift of the ITCZ.

The zonal means of tropospheric temperature and precipitation anomalies for individual models and MME are plotted versus the latitude (Fig. 2). Most of the models reveal larger temperature decreases in the Northern Hemisphere than those in the Southern Hemisphere, and we did observe intermodel discrepancies (Fig. 2a). Some models result in stronger north-south temperature contrasts, such as GFDL CM3 and

FIG. 3. ITCZ shifts ($^{\circ}$) and the cross-equatorial energy transport (PW) for each model and for MME. Positive values indicate northward shift (y axis) and northward energy transport (x axis) and vice versa.

CESM1, whereas others result in relatively smaller contrasts, such as CCSM4. The zonal-mean precipitation anomaly (Fig. 2b) shows significant dips near the equator in all models, whereas the positive precipitation anomaly exists in the tropics in the Southern Hemisphere. Notably, CESM1 shows a spike at 10°S (also shown in Fig. S3a in the online supplemental material), representing a strong double-ITCZ bias in this model (Wang et al. 2015).

4. Validation of existing mechanisms

Global temperature and precipitation spatial patterns in aerosol-cooling simulations change in an opposite direction to those in global warming simulations. This suggests that the mechanisms used to explain the responses of the tropical atmosphere in global warming scenarios can be applied in a cooling situation. This section examines tropical atmosphere in several aspects, including the interhemispheric energetic framework for the ITCZ shift, the strength of large-scale circulations, and the M' and upped-ante mechanisms for tropical precipitation distribution.

a. The ITCZ shift and cross-equatorial energy transport

The interhemispheric energetic framework for the ITCZ shift states that the Northern Hemisphere–South Hemisphere temperature contrast drives meridional energy transport across the equator through the shift of Hadley cell. The upper-level divergent flow of Hadley cell transports potential energy toward the cooler hemisphere and the lower-level convergent flow carries moisture to the warmer hemisphere, which shifts the precipitation band (or the ITCZ). Therefore, the ITCZ

FIG. 4. Vertical profile of vertical motion anomaly over the tropical convective area [indicated by upward motion at 500 hPa (ω_{500}^+) between 30°S and 30°N] for each model (colors and line patterns) and for MME (black). The y axis indicates the vertical coordinate (hPa), and the x axis is for vertical motion anomaly (Pa s^{-1}). Negative values indicate enhanced upward motion, and positive values indicate downward motion.

shift can be indicated by the centroid of the area-integrated precipitation. The cross-equatorial energy transport is computed for all models and MME and compared with the ITCZ shift (Fig. 3), which can be compared to Fig. 3a in Frierson and Hwang (2012). Although the shift of ITCZ is less than 0.5° in latitude, this amount of shift passes 99% significance test. In Fig. 3, most models are diagonally aligned and have the ITCZ shift southward with the energy transported northward, except for the GISS model. The models that result in the most significant shifts are CESM1 and GFDL-CM3, which also have a stronger temperature contrast between the Northern and Southern Hemispheres (Fig. 2a) and larger precipitation dips (Fig. 2b).

b. Tropical convection and circulation changes

Convection depth is influenced by global mean temperature (Chou and Chen 2010). In a warmer Earth, the higher tropopause in the tropics leads to larger M due to the larger depth of vertical integration in (5), and therefore the troposphere is more stable and convection weakens (Chou et al. 2013). We compute the averaged ω' over the convective area, which is determined by the upward motion at 500 hPa (ω_{500}^+) between 30°S and 30°N (Fig. 4). All models have similar profiles of ω' where the vertical motion becomes stronger below 200–300 hPa and weaker above, resulting in a decrease in the

FIG. 5. Change rate of convective mass flux M_c in MME, normalized by the global-mean surface temperature anomaly ($\% K^{-1}$). Positive values indicate subsidence, and negative values indicate enhanced convection in cooling simulations. The red contour indicates 4 mm day^{-1} climatology precipitation of MME.

convection depth. This result is consistent with that shown in Fig. 3 in [Chou and Chen \(2010\)](#), with an opposite tendency, although larger intermodel differences exist in our study.

The large-scale circulation is examined by analyzing the convective mass flux M_c (Fig. 5), which can be compared with Fig. 8c in [Vecchi and Soden \(2007\)](#). The spatial pattern of the M_c change rate in our study is similar to that in [Vecchi and Soden \(2007\)](#), with an opposite tendency, because the rate of change of M_c is normalized by the global mean surface temperature anomaly, which is negative in our study. As shown in Fig. 5, enhanced convection (negative values) occurs around the tropics in the Southern Hemisphere, except the SPCZ. By contrast, subsidence (positive values) tends to occur in the Northern Hemisphere tropics, such as the central Pacific and the double-ITCZ, SPCZ, and Indo-China regions, as well as tropical islands over maritime continents and parts of the north Indian Ocean and central Africa. The significant negative values over Northern Hemisphere subtropical regions, such as North Africa, the Middle East, and North America, are due to weaker subsidence. Our result reveals more spatial details because the model resolutions in CMIP5 are generally higher than those in CMIP3.

We then separate the large-scale circulation change into the zonal component (i.e., Hadley circulation) and the stationary eddy component (i.e., Walker circulation) as Fig. 6 in [Vecchi and Soden \(2007\)](#). The total strength of large-scale circulation can be defined as the spatial variance of the upward motion at 500 hPa (ω_{500}^+) over the tropics (30°S – 30°N) ([Held and Soden 2006](#)). The zonal component is calculated using the zonal mean of ω_{500}^+ ,

FIG. 6. Change in the spatial variance of ω_{500}^+ (upward motion) over 30°S – 30°N for all models and MME. The x axis indicates large-scale circulation change in total variance, and the y axis indicates changes in variance due to different circulation components. Color symbols indicate variance change due to zonal means of ω_{500}^+ (i.e., Hadley circulation). Black symbols indicate variance change due to stationary eddies of ω_{500}^+ (i.e., Walker circulation). The magnitude is $1 \times 10^{-5} (\text{Pa s}^{-1})^2$.

and the stationary eddy component is calculated using the deviations of ω_{500}^+ from its zonal mean, only considering the changes in the pre-existing convective area. Unlike [Vecchi and Soden \(2007\)](#), who compared two 10-yr period simulations, we used 80 years of climatology monthly mean data to perform this calculation and then subtracted the preindustrial control run from the aerosol-forcing run. The long-term average may have weakened the amplitude of variance, but the result still exhibits clear signals of the circulation change.

Figure 6 shows that five out of seven models have a positive change in total variance (referring to the x axis), indicating enhanced large-scale circulations over the tropics. The change in circulation is decomposed into the Walker circulation (black symbols) and Hadley circulation (colored symbols) (referring to the y axis for the component values). Most of the models exhibit a similar change in the Hadley circulation (enhanced circulations), whereas large intermodel differences exist for the Walker circulation, which dominates the total circulation change in each model. The magnitude of the Hadley circulation change is comparable to the Walker circulation change. Computation using MME (asterisks in Fig. 6) shows similar results. Overall, the tropical large-scale circulations are enhanced in the cooling simulations, and the enhancement of the Hadley circulation is as significant as that of the Walker circulation.

The consistency of Hadley circulation enhancement and ITCZ shift between models reflects the results of

MME moisture budget

FIG. 7. Moisture budget for MME. The (a) precipitation anomaly (converted to W m^{-2} for better comparison) has contributions from the (b) vertical dynamic feedback $-\langle \omega' \partial_p q \rangle$, (c) direct moisture effect $-\langle \bar{\omega} \partial_p q' \rangle$, (d) horizontal advection $-\langle \bar{\mathbf{v}} \cdot \nabla q' \rangle - \langle \mathbf{v}' \cdot \nabla \bar{q} \rangle$, (e) surface latent heat flux, and (f) the residual term. Panels are overlaid with a climatology contour of 4 mm day^{-1} (red contour).

some previous studies regarding the asymmetry response of the atmosphere to the anthropogenic aerosols. Further discussion is provided in section 5.

c. Moisture budget analysis

The moisture and MSE budget analyses presented here and in the next subsection use MME variables. Figures for individual simulations are shown in the online supplemental material (Figs. S1–S20). Practically, the precipitation anomaly shown in Fig. 1c can be approximated by (9). The contribution of each term is shown in Fig. 7, superimposed with the contour of 4 mm day^{-1} climatology precipitation of the control run (red contour). Here, the unit of precipitation has been converted into energy flux for better comparison with other terms. The spatial pattern of the vertical dynamic term $-\langle \omega' \partial_p q \rangle$ (Fig. 7b) closely resembles the precipitation anomaly, suggesting that changes in vertical motion ω' are the dominant process for the precipitation change.

The M'_q effect $-\langle \bar{\omega} \partial_p q' \rangle$ (Fig. 7c) is associated with changes in moisture vertical distribution $\partial_p q'$. When the surface evaporation (Fig. 7e) decreases in response to a cooling atmosphere, the vertical gradient of moisture decreases too, and this causes a negative change in the

vertical gradient of moisture $\partial_p q'$ globally. Therefore, the sign of the direct moisture effect term is determined by the climatological pressure velocity $\bar{\omega}$, which is similar with the climatological precipitation of 4 mm day^{-1} (Chou et al. 2009). This term causes an overall precipitation reduction in the convective region by approximately 30%, which is consistent with the amplitude of precipitation change found in Chou and Neelin (2004) and a precipitation increase outside the convective region.

The horizontal advection of moisture (Fig. 7d) includes the anomalous wind component $-\langle \mathbf{v}' \cdot \nabla \bar{q} \rangle$ and the anomalous moisture component (or the upped-ante mechanism) $-\langle \bar{\mathbf{v}} \cdot \nabla q' \rangle$. The two components have stronger magnitudes in the extratropics than in the deep tropics. However, their values are generally comparable with opposite signs (figures not shown), and thus their combined contribution is minor compared with other terms. Nevertheless, we can still see positive values along the edge of convective areas (red contour), such as over the south Indian Ocean, the southern and northern margins of the SPCZ, and the tropical Atlantic Ocean. This pattern supports that part of the precipitation increase at the margin of convective areas is through the anomalous wind mechanism and the upped-ante mechanism.

MME MSE budget

FIG. 8. MSE budget for MME. The (a) vertical dynamic term $\langle \omega' \partial_p \bar{h} \rangle$ has contributions from the (b) M' mechanism $-\langle \bar{\omega} \partial_p \bar{h}' \rangle$, (c) advection term $-\langle \bar{\mathbf{v}} \cdot \nabla(T + q) \rangle - \langle \mathbf{v}' \cdot \nabla(\bar{T} + \bar{q}) \rangle$, and (d) $F^{\text{net}'}$; also shown is (e) the residual term. Panels are overlaid with a climatology contour of 4 mm day⁻¹ (red contour).

The LH anomaly (Fig. 7e) is mostly negative globally as expected in a cooling world, except over Southern Africa, Australia, South America, and parts of the subtropical lands in the Northern Hemisphere. The residual is negligible in the warm pool but has larger values over areas with complex topography, such as East Africa, south of the Tibetan Plateau, the Rocky Mountains, and the Andes. This is likely due to the central finite difference involved in the calculation of horizontal gradients in the advection terms, which often produces noisy signals over complex topography.

d. MSE budget analysis

Because the most dominant term in the moisture budget analysis is the vertical dynamic term $-\langle \omega' \partial_p \bar{q} \rangle$, the MSE Eq. (10) can be used to diagnose the processes that contribute to ω . These processes then feedback to precipitation, as shown in (11). Figure 8 presents the spatial patterns of each term in (10). In the MSE equation, the magnitude of the vertical dynamic term (the term on the left-hand side) is greatly weakened because the vertical gradient of dry static energy $\partial_p \bar{s}$ is the opposite of that of moisture $\partial_p \bar{q}$, and the summation of these two terms $\langle \omega' \partial_p \bar{h} \rangle = \langle \omega' \partial_p \bar{s} \rangle + \langle \omega' \partial_p \bar{q} \rangle$ is partially canceled in the tropics. This cancelation is not significant

in the high latitudes where the atmosphere is baroclinic. The spatial pattern of $\langle \omega' \partial_p \bar{h} \rangle$ closely resembles that of the precipitation anomaly (Fig. 7a), supporting that the M' mechanism, horizontal advection term, and $F^{\text{net}'}$ can influence precipitation through vertical motion. The M' mechanism $-\langle \bar{\omega} \partial_p \bar{h}' \rangle$ (Fig. 8b) shows negative values (reducing precipitation) over most of the tropics, except over the northeastern Pacific ITCZ, cold tongue area, and North Atlantic Ocean. The M' mechanism may seem small in Fig. 8b, but its contribution to precipitation is enhanced by the gross moist stability multiplier described in (11).

The advection term $-\langle \mathbf{v}' \cdot \nabla(\bar{T} + \bar{q}) \rangle - \langle \bar{\mathbf{v}} \cdot \nabla(T + q) \rangle$ (Fig. 8c) includes contributions from temperature and moisture advection. Temperature advection is responsible for most high-latitude variation, whereas moisture advection is dominant in the tropics (figures not shown). Large anomalies are found at the margin of convective regions, such as over the southern Indian Ocean, the southern boundary of the SPCZ, and the northern boundary of the Pacific ITCZ (i.e., along the red contour), which supports the importance of the upped-ante mechanism in these regions. The $F^{\text{net}'}$ term (Fig. 8d) exhibits positive contributions over the eastern Pacific double-ITCZ regions and over the subtropics.

Notably, all these terms are further enhanced by the gross moist stability multiplier in (11) and can contribute to significant precipitation changes. The residual term (Fig. 8e) is generally large in the extratropics because the quasi-equilibrium assumption is no longer applicable.

The simulations reveal large intermodel differences in terms of spatial pattern and magnitude, but the moisture and MSE budget analysis suggests common results where the precipitation anomaly in the tropics is mainly produced through the process associated with the M' mechanism in the center of the convective areas. The horizontal advection terms (including the upped-ante mechanism) are important along the edge of the convective area where the horizontal motion converges energy into the atmospheric column and enhances convection, producing more precipitation. Note that F^{net} plays secondary roles over different places and may be important for the double-ITCZ phenomenon over the southeastern Pacific.

5. Discussion

We examine changes in the tropical atmosphere induced only by anthropogenic aerosols and compared the results to simulations under GHG-warming scenarios from previous studies. These changes can be explained by the same mechanisms employed in global warming scenarios, which are the energy balance between two hemispheres, the M' mechanism, and upped-ante mechanism. However, not all changes are exactly opposite or symmetric. The spatial distribution of tropical precipitation shows reversed tendency as to the global warming scenario in the first-order approximation. This means the large-scale precipitation pattern change can be explained by the same mechanisms, but the actual precipitation is still influenced by processes of various scales. Therefore, there is tendency of decreasing (increasing) precipitation within (outside) convective areas, and there are exceptions in smaller regions.

The ITCZ shift in the warming scenarios (Frierson and Hwang 2012) is not consistently northward ($\sim 50\%$ of the simulations shifted northward and $\sim 50\%$ shifted southward), although the connection between the ITCZ shift and cross-equatorial energy transport is tight and clear. It is worth mentioning that the simulations performed in Frierson and Hwang (2012) were coupled with slab-ocean models, which may underestimate the energy transported by the ocean currents and cause the inconsistency in the atmospheric component between different models. A recent study by McFarlane and Frierson (2017) showed that in CMIP5 RCP8.5 simulations different feedback processes may have opposite

effects and counter each other, leading to an ambiguous direction of ITCZ shift. In this study, the more consistent shift of the ITCZ is observed in the single forcing cooling simulations. This phenomenon may be related to the enhancement of the Hadley circulation, which is also more significant in our study than it is in global warming cases. A possible reason is GHGs influence precipitation through both fast and slow processes and anthropogenic aerosols through mainly slow processes (Myhre et al. 2017). Slow processes involve surface temperature, which reflects the large land–sea contrast in the Northern Hemisphere and tends to produce stronger interhemispheric asymmetry. In general, our results agree with this interhemispheric asymmetry forced by anthropogenic aerosols (i.e., Wang et al. 2016; Chung and Soden 2017).

The role of the ocean is particularly interesting, and the result from a recent study may provide an explanation to the Hadley circulation changes in this study. Hill et al. (2015) compared a set of GHG-warming simulations and aerosol-cooling simulations using the GFDL AGCM (AM2.1) coupled with a slab ocean, showing that the majority of MSE meridional transportation in the two scenarios is carried by different components of atmospheric circulations. In the warming case, the extratropical eddies dominate the energy transport, whereas the mean meridional circulation (MMC) is the most important component in the cooling case. Furthermore, the meridional energy transport carried by MMC has a significant seasonal cycle in their warming simulation, which might lead to an ambiguous sign after annual averaging. However, their aerosol-cooling simulation revealed a strong year-round northward energy transport by MMC (Fig. 3 in Hill et al. 2015). This might explain the more consistent ITCZ shift and the stronger enhancement of the Hadley circulation in our study. More importantly, their result suggested that atmospheric circulations respond to global warming and cooling differently in terms of energy transport. However, Hill et al. (2015) examined only one model without the dynamic ocean. An idealized study by Kang et al. (2018) showed that Ekman transport may play an important role in cross-equatorial energy transport, which then reduces the importance of atmospheric circulation in the meridional energy transport and mutes the ITCZ shift. In extratropical oceans, Armour et al. (2016) has suggested that the Southern Ocean meridional overturning circulation plays an important role in storing heat into oceans. Hwang et al. (2017) prescribed the Southern Ocean heat uptake in a slab-ocean model to mimic this ocean dynamic process and showed that the heat uptake does influence the ITCZ shift significantly. Therefore, if the ocean dynamics is not included, the

importance of tropical circulations might be overestimated to some degrees. Considering the complexity of ocean dynamics and the thermal capacity, further analysis using different models is required for a conclusive result.

6. Conclusions

The concentrations of anthropogenic aerosols have increased since industrial revolution, but are projected to decline in the future (Rotstayn et al. 2015; Acosta Navarro et al. 2017). It is important to understand the effects of both GHGs and aerosol forcings individually before discussing the combined effects. The impacts of anthropogenic aerosols mainly go through slow processes involving surface temperature changes. This study focuses on understanding the processes in the tropical atmosphere in response to global temperature changes and complements to previous studies that emphasized more on ocean and clouds. In this study, we analyze 10 CMIP5 fully coupled simulations from seven models forced by anthropogenic aerosols only. In all simulations, the aerosol direct and indirect effects are included. The configurations are different across simulations: some are driven by the historical aerosol emissions and some by concentration. However, the simulations between these two settings cannot provide enough evidence to show that the aerosol configuration makes significant impacts to the results. There is significant intermodel discrepancy; for example, the double-ITCZ bias is much more significant in some of the models. Therefore, a set of ensemble simulations with enough members from multiple models is crucial for such a study.

We examine the spatial distribution of tropical precipitation, positional shift of the ITCZ versus cross-equatorial energy transport, vertical profile of tropical convection, and strength of large-scale circulations using individual simulations and MME. Our results suggest that the tropical precipitation and convection depth respond to a global cooling scenario through the M' and upped-ante mechanisms proposed in the global warming scenario. We examine the two mechanisms by analyzing the moisture and MSE budget. Both mechanisms can be used to explain the tropical precipitation anomaly and the associated vertical motion change successfully. However, the meridional energy transport associated with large-scale circulation (i.e., the Hadley circulation) is not symmetric to the GHG-warming case. We note a more consistent southward ITCZ shift and markedly enhanced Hadley circulation in the cooling simulations. Our results also support those of a recent study (Hill et al. 2015) that found that meridional

energy transport may be carried by different components of the atmospheric circulations in warming and cooling cases. This asymmetric circulation response may be due to air–sea interactions associated with different portions of sea surface temperature anomaly (Xie et al. 2013; Hill et al. 2015), but this requires further studies to confirm.

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