Differing Impacts of Black Carbon and Sulfate Aerosols on Global Precipitation and the ITCZ Location via Atmosphere and Ocean Energy Perturbations

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ABSTRACT

This study explores the effects of black carbon (BC) and sulfate (SO_4) on global and tropical precipitation with a climate model. Results show that BC causes a decrease in global annual mean precipitation, consisting of a large negative tendency of a fast precipitation response scaling with instantaneous atmospheric absorption and a small positive tendency of a slow precipitation response scaling with the BC-caused global warming. SO_4 also causes a decrease in global annual mean precipitation, which is dominated by a slow precipitation response corresponding to the surface cooling caused by SO_4 . BC causes a northward shift of the intertropical convergence zone (ITCZ), mainly through a fast precipitation response, whereas SO_4 causes a southward shift of the ITCZ through a slow precipitation response. The displacements of the ITCZ caused by BC and SO_4 are found to linearly correlate with the corresponding changes in cross-equatorial heat transport in the atmosphere, with a regression coefficient of about -3°PW^{-1} , implying that the ITCZ shifts occur as manifestations of the atmospheric cross-equatorial heat transport changes in response to the BC and SO_4 forcings. The atmospheric cross-equatorial heat transport anomaly caused by BC is basically driven by the BC-induced interhemispheric contrast in instantaneous atmospheric absorption, whereas the atmospheric cross-equatorial heat transport anomaly caused by SO_4 is mostly attributable to the response of evaporation. It is found that a slab-ocean model exaggerates the cross-equatorial heat transport response in the atmosphere and the ITCZ shift both for BC and SO_4 , as compared with an ocean-coupled model. This underscores the importance of using an ocean-coupled model in modeling studies of the tropical climate response to aerosols.

1. Introduction

The climatic effects of aerosols are less constrained than those of greenhouse gases because of the complex ways in which aerosols influence the climate system. For example, hygroscopic aerosols can act as cloud condensation nuclei and change the properties of clouds before changing surface temperature; absorbing aerosols can also change the properties of clouds, but mainly through heating the

atmosphere (Ming et al. 2010). These make the apparent hydrological sensitivities (the ratio between global-mean precipitation change and surface temperature change) of aerosols more diverse among different climate models than those of greenhouse gases (Samset et al. 2016).

Among the various types of aerosols, black carbon (BC) and sulfate (SO_4) have been compared more often, because of their distinctively different optical properties (e.g., Kim et al. 2014; Ocko et al. 2014; Lin et al. 2016; Samset et al. 2016; Suzuki and Takemura 2019). Although still with uncertainties in magnitude, it is commonly recognized that BC has a warming effect and SO_4 has a cooling effect on global-mean surface temperature (Boucher et al. 2013; Takemura and Suzuki 2019). The peculiarity of BC is that it has a negative hydrological

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sensitivity, resulting from a positive global-mean surface temperature response and a negative global-mean precipitation response (e.g., Yoshimori and Broccoli 2008; Ming et al. 2010; Ocko et al. 2014; Samset et al. 2016). Ming et al. (2010) demonstrated that the effect of BC on global-mean precipitation can be divided into a suppressing effect through strong shortwave atmospheric absorption and an enhancing effect through surface warming. This is analogous to the division between fast and slow precipitation responses in Bala et al. (2010). The negative hydrological sensitivity of BC results from the suppressing effect of atmospheric absorption (or the fast precipitation response), which outweighs the enhancing effect of surface warming (or the slow precipitation response). The fast precipitation response and the associated rapid adjustment processes are receiving increased attention in the science community, as they are very important in understanding the diversity of the precipitation responses to different climate drivers and in different models (Myhre et al. 2017; Stjern et al. 2017). In contrast to BC, SO₄ has a positive hydrological sensitivity, more similar to greenhouse gases, as SO₄ has little shortwave absorption in the atmosphere (Ocko et al. 2014; Samset et al. 2016). But the ways in which SO₄ and greenhouse gases influence the climate are different.

In climatology, Earth's energy budget features a northward cross-equatorial heat transport in the ocean and a southward cross-equatorial heat transport in the atmosphere (e.g., Marshall et al. 2014; Kang et al. 2015; Stephens et al. 2016). The southward atmospheric cross-equatorial heat transport is associated with a northward cross-equatorial moisture transport, as manifested in the upper and lower branches of the Hadley cell. Some studies demonstrated that the northward oceanic cross-equatorial heat transport determines the north-of-the-equator annual mean location of the intertropical convergence zone (ITCZ) (e.g., Frierson et al. 2013; Marshall et al. 2014). Accordingly, perturbations of the hemispheric energy balance can change the latitudinal location of the ITCZ and move it toward the relatively warmed hemisphere (e.g., Chiang and Bitz 2005; Broccoli et al. 2006; Kang et al. 2008, 2009; Hwang et al. 2013; Mahajan et al. 2013; Schneider et al. 2014; Voigt et al. 2014). Therefore, considering the energetic effects of aerosols that are abundant in the Northern Hemisphere (NH), some studies have explored the effects of aerosols, including BC and SO₄, on ITCZ displacement (e.g., Chung and Seinfeld 2005; Hwang et al. 2013; Ocko et al. 2014). Kim et al. (2015) examined the sensitivity of the ITCZ response to the altitude of BC concentration using an aquaplanet GCM. Some studies have demonstrated that anthropogenic aerosols were the dominant drivers of the southward shift of the ITCZ during the second half of the twentieth century; with the mitigation

of anthropogenic aerosol emissions in the twenty-first century, the ITCZ is expected to migrate northwardly (Allen et al. 2015; Allen 2015). However, a complete picture of interhemispheric energy budget perturbations, both at the top of the atmosphere (TOA) and the surface, due to BC and SO₄ aerosols, has been missing in previous studies. This has also hindered achieving a clear understanding of how initial perturbations of the interhemispheric energy budget due to instantaneous radiative forcing (RF) are linked to eventual climate responses.

In this study, we explore the following aspects of the climatic effects of BC and SO₄: 1) comparing the ways in which the climate responds to the instantaneous RFs of BC and SO₄; 2) investigating the precipitation responses caused by BC and SO₄ in the context of energy balance constraint at both global and hemispheric scales; 3) quantitatively assessing the relationship between the ITCZ displacements caused by BC and SO₄ and the corresponding atmospheric cross-equatorial heat transport changes; and 4) analyzing the different contributions to the cross-equatorial heat transport changes caused by BC and SO₄ and extracting pictures of how BC and SO₄ affect the latitudinal position of the ITCZ. Some of the first two aspects have been discussed by previous studies (e.g., a series of papers from the Precipitation Driver Response Model Intercomparison Project; <https://www.cicero.oslo.no/en/PDRMIP/PDRMIP-publications>). However, we still include them into our discussions in order to again analyze the links between the instantaneous RFs and the climate responses, which have been missing in previous studies, by quantifying relative contributions of different energy components through rapid adjustment, and the slow response to the RFs of BC and SO₄. The quantitative relationship between the atmospheric absorption response and column water vapor response to BC and SO₄ is also investigated in this study, as it largely determines the simulated hydrological sensitivity of aerosols.

Recent studies have highlighted the importance of using a fully coupled ocean model when studying the shift of the ITCZ (e.g., Ocko et al. 2014; Hawcroft et al. 2017; Kay et al. 2016; Green and Marshall 2017). For example, Green and Marshall (2017) found that for a given interhemispheric heating contrast, the ocean responses in a fully coupled ocean model can damp ITCZ shift owing to a coupling between the atmosphere's Hadley cells and the ocean's subtropical cells by the trade winds and their associated surface stresses. However, few among them directly compared fully coupled ocean and slab-ocean models within a common model framework in the context of the climate impacts of BC and SO₄. Therefore, this study pays particular attention to comparing the ITCZ displacements caused by BC and SO₄ between an atmosphere–ocean fully coupled model and atmosphere–slab-ocean coupled model, to better

understand the relative roles of the atmosphere and ocean in the aerosol-induced ITCZ shift. Section 2 briefly describes the models used in this study and the experimental setup. Results are presented in section 3, followed by conclusions and a summary in section 4.

2. Models and experimental setup

a. Models

The Model for Interdisciplinary Research on Climate, version 5.2 (MIROC5.2; Watanabe et al. 2010), is used in this work, with the standard resolution of T85 (about $1.4^\circ \times 1.4^\circ$) for the atmospheric model and $1^\circ \times 1^\circ$ for the ocean model. The Spectral Radiation Transport Model for Aerosol Species (SPRINTARS; Takemura et al. 2005, 2009) was implemented in MIROC5.2. SPRINTARS includes five major aerosol types: BC, organic carbon (OC), SO_4 , soil dust, and sea salt. The last two types of aerosols and dimethyl sulfide (a precursor of SO_4) are emitted online, whereas prescribed emission data are used for the other aerosols and/or their precursors. SPRINTARS is coupled with radiation and cloud microphysics, so the direct, indirect, and semidirect effects of aerosols are incorporated. SPRINTARS considers the internal mixing between BC and OC. As described in the following section, we fix the OC emission instead of prescribing its concentration in the atmosphere. This introduces OC responses to BC and SO_4 emission changes. The emission level of OC in the preindustrial era was already much higher than that of BC, even in 2000. Therefore, the climate effects of BC may be influenced by the OC response, especially when the emissions of BC are assumed insufficiently large. However, fixing aerosol emissions instead of their concentrations also has an advantage, because aerosol emissions are directly controlled by human activities.

A brief introduction to the treatment of BC and OC in SPRINTARS is provided here, but a more detailed description is documented in Takemura et al. (2005). The physical processes of BC and OC, including their emissions, in-cloud activation, subcloud scavenging, in-cloud scavenging, re-emission from droplet evaporation, dry deposition, and gravitational settling, are treated in the same subroutine. Half of the BC emitted from fossil fuels and aircraft and all of it emitted from biomass burning, domestic activities, and agriculture are assigned the same in-cloud coefficient as OC, and are treated simultaneously with OC during the processes of wet and dry deposition. These parts of BC are internally mixed with OC. The optical properties of BC and OC internally mixing aerosols are divided into three categories according to BC/OC concentration ratio, and the corresponding internally mixing aerosol concentrations are calculated as follows:

$$\begin{aligned} \text{QCIM}(1) &= \text{QC}, \quad f \geq 0.3, \\ \left\{ \begin{aligned} \text{QCIM}(1) &= \frac{f - 0.15}{0.15} \text{QC} \\ \text{QCIM}(2) &= \left(1 - \frac{f - 0.15}{0.15}\right) \text{QC} \end{aligned} \right. , \quad 0.15 \leq f < 0.3, \\ \left\{ \begin{aligned} \text{QCIM}(2) &= \frac{f}{0.15} \text{QC} \\ \text{QCIM}(3) &= \left(1 - \frac{f}{0.15}\right) \text{QC} \end{aligned} \right. , \quad 0 \leq f < 0.15, \end{aligned} \quad (1)$$

where f is the ratio between BC and OC concentrations, QC represents the total concentration of BC and OC, and QCIM represents the mixed aerosol concentration, dependent on the BC/OC ratio. The remaining half of the BC emitted from fossil fuels and aircraft is treated as externally mixed with OC.

b. Experimental setup

In this work, we concentrate on the difference between the BC- and SO_4 -induced global and tropical precipitation responses. The prescribed aerosol emission data between 1850 and 2000 from the Intergovernmental Panel on Climate Change (IPCC) representative concentration pathways (<http://www.iiasa.ac.at/web-apps/tnt/RcpDb>) are used.

Three groups of experiments with different model configurations are conducted to investigate the equilibrium climate responses to the BC- and SO_4 -induced forcings (Table 1). The first group of experiments uses the atmosphere component model of MIROC5.2 with prescribed sea surface temperature (SST) and sea ice, hereafter referred to as the AGCM simulations. The second and third groups replace the prescribed SST and sea ice boundary conditions with MIROC5.2's ocean component model, COCO (CCSR Ocean Component Model; Hasumi 2015) and a mixed layer slab-ocean model, respectively. The second and third groups of experiments are hereafter referred to as the CGCM and slab-ocean model (SOM) simulations, respectively. The slab-ocean model employs a constant mixed layer depth of 50 m, and uses a prescribed Q flux to represent the ocean heat transport (Yoshimori et al. 2017). The AGCM simulations are run for 15 years, with the last 10 years of results used for analysis. The CGCM simulations are run for 100 years, and the SOM simulations are run for 50 years, since a balance is generally achieved after about 20 years (see Fig. S1 in the online supplemental material). The results of the last 30 years for the

TABLE 1. Simulation setup.

Group name (running years)	Expt	Aerosol emissions
AGCM (15)	A _{REF}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₁₈₅₀
	A _{1BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀
	A _{5BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀ × 5
	A _{10BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀ × 10
	A _{1SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀
	A _{5SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀ × 5
CGCM (100)	A _{10SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀ × 10
	C _{REF}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₁₈₅₀
	C _{1BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀
	C _{5BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀ × 5
	C _{10BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀ × 10
	C _{1SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀
SOM (50)	C _{5SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀ × 5
	C _{10SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀ × 10
	S _{REF}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₁₈₅₀
	S _{10BC}	OC ₁₈₅₀ ; SO ₄ ₁₈₅₀ ; BC ₂₀₀₀ × 10
	S _{5SO4}	OC ₁₈₅₀ ; BC ₁₈₅₀ ; SO ₄ ₂₀₀₀ × 5

CGCM and SOM simulations are used for analysis. It should be noted that to obtain a real equilibrium state requires a longer period than 100 years, to wait for the ocean to fully adjust. But, here, our focus is on comparing BC and SO₄. Thus, a quasi-equilibrium state is used, as we found that besides the TOA net radiation (Fig. S1), the global-mean surface temperature change and the ITCZ shift caused by aerosols do not change very much after 30 years in CGCM simulations (figures not shown).

The simulations are performed with the emissions of BC and SO₄ perturbed globally. For this purpose, three levels of emission magnitudes are assumed for BC and SO₄ in the AGCM and CGCM simulations: 1, 5, and 10 times the 2000 emission magnitudes (denoted as 1BC and 1SO₄, 5BC and 5SO₄, and 10BC and 10SO₄, respectively). The emissions of OC are fixed as per those in 1850, and the emissions of soil dust and sea salt aerosols are turned off. The CGCM and SOM simulations are compared with each other to examine the ocean's role in modulating the precipitation responses to BC and SO₄, particularly over the tropical region. For this comparison, only the 10BC and 5SO₄ cases are conducted in the SOM experiment group to analyze the pronounced responses of precipitation. In each group, the experiment with all prescribed aerosol emissions in 1850 is used as reference; except for the emissions of BC and SO₄, all model settings among the different experiments are identical.

c. Analytical method

By analyzing the results of the first two groups of experiments (i.e., AGCM and CGCM), we can separately

quantify the instantaneous radiative forcing (RF), rapid adjustment, and climate response for BC and SO₄, as well as the fast and slow precipitation responses induced by BC and SO₄ following the approach of Samset et al. (2016). In the rest of this section, we provide a brief introduction to how RF, rapid adjustment, and response are calculated in this study. A flowchart is also given as an auxiliary illustration (Fig. 1).

RF means the instantaneous net radiation change at the TOA when introducing a forcing agent, while keeping the states of the atmosphere and the surface unchanged. In this work, RF includes aerosols' direct effect and the Twomey effect (Twomey 1977). The model we use can diagnose the direct effect of aerosols by invoking the radiation module twice at each step in a single AGCM simulation—once with aerosols and then without aerosols. The difference in TOA net radiation between the two radiative calculations is considered as the aerosols' direct effect. The Twomey effect can also be diagnosed, using a method analogous to that for the aerosols' direct effect, except the two invocations of the radiation module are replaced by one with a prescribed cloud droplet radius and the other with a prognostic cloud droplet radius (based on the aerosol number concentration). Finally, RF is calculated as the direct and Twomey effects of aerosols in simulations with aerosol emissions perturbed minus those in the reference simulation.

Rapid adjustment caused by a forcing agent consists of the TOA net radiation changes from the adjustments in the atmosphere and on the land, but the ocean is not allowed to change. For different forcing agents, rapid adjustments can offset or amplify RF to varying degrees (Smith et al. 2018). Before calculating the rapid adjustment caused by aerosols, we first need to calculate the aerosols' effective RF—a concept recommended by the IPCC AR5 (Boucher et al. 2013). In this work, the effective RF caused by aerosols is calculated as the TOA net radiation difference between two AGCM simulations—one using present-day aerosol emissions and the other using preindustrial aerosol emissions. Then, the rapid adjustment caused by aerosols can be obtained by subtracting their RF from the corresponding effective RF. Rapid adjustment can be further divided into a cloud component and a noncloud component. Noncloud rapid adjustment is calculated with the same approach as rapid adjustment, but under clear-sky conditions (if not specified, the radiative calculation is under all-sky condition in this work). Then, cloud rapid adjustment can be calculated as the residual of rapid adjustment and its noncloud component.

The response to aerosols results from the aerosol-induced change in global-mean surface temperature.

FIG. 1. Flowchart showing how the RF, rapid adjustment, and response are calculated in this study. Note that ΔF_{net} and $\Delta F_{\text{net,clr}}$ mean the TOA net radiation changes caused by a forcing agent under all-sky and clear-sky conditions calculated from two CGCM simulations; ERF and ERF_{clr} mean the effective RFs caused by a forcing agent under all-sky and clear-sky conditions calculated from two AGCM simulations; and RF and RF_{clr} can be first diagnosed using a double-call method in one AGCM simulation. After that, RF and RF_{clr} are calculated as the difference between perturbed and reference AGCM simulations. The double-call method means one invocation of radiation scheme with and the other one without aerosols when calculating aerosol direct effect, or one with prognostic and the other one with prescribed cloud droplet radius when calculating aerosol Twomey effect. Brackets indicate the difference between two variables.

Before calculating the response to aerosols, we first need to calculate the total TOA net radiation change caused by aerosols using CGCM simulations (Fig. 1), analogous to the method for calculating the aerosols' effective RF. Then, the response to aerosols can be obtained by subtracting the aerosols' effective RF from the corresponding total TOA net radiation change. The response can also be divided into a cloud component and a non-cloud component, analogous to rapid adjustment.

3. Results

a. RF, rapid adjustment, and response at the TOA

The three components estimated above (RF, rapid adjustment, and response) correspond to different time scales of the net radiation change at the TOA in response to a given forcing agent. When the climate regains its balance, the sum of the three terms approaches a near-zero global-mean value (though often not exactly zero because of the change in oceanic heat storage), which means rapid adjustment plus response is approximately $-\text{RF}$. Therefore, the relative magnitudes of rapid adjustment and response indicate how the climate system responds to the RF due to a given forcing agent over two different time scales.

The zonal-mean RF, rapid adjustment, and response for BC and SO_4 are shown in Fig. 2. The zonal-mean values (here and hereafter) are weighted by $\cos\phi$ (where ϕ is latitude) so that the area encircled by each curve and

the x axis generally equal the corresponding global mean. Owing to its absorption of shortwave radiation, BC causes positive RF at the TOA, particularly over the tropics of both hemispheres and the midlatitude NH (Fig. 2a). In response to the BC-induced RF, rapid adjustment takes the primary role, especially over the equatorial region, and cloud rapid adjustment dominates the total rapid adjustment of BC (Fig. 2c). In global annual mean terms, the normalized magnitudes of cloud and noncloud rapid adjustments relative to RF for BC are -0.42 [-0.51 to -0.33 (minimum to maximum among different aerosol emission levels; the same hereafter)] and -0.29 (-0.33 to -0.21), respectively. The large negative cloud rapid adjustment for BC is attributable to the increase in low clouds and decreases in middle and high clouds, which exert a cooling effect on TOA radiation. The response to BC is relatively small, but with large uncertainties, and the cloud and noncloud components of the BC-induced response are opposite in sign over most latitudes (Fig. 2e).

SO_4 causes negative RF at the TOA, especially over the tropical and midlatitude NH (Fig. 2b). It is found that the rapid adjustment and response for SO_4 are generally opposite in sign, and the magnitude of the former is smaller than the latter. In other words, a slow response is the main contributor to the climate response to the SO_4 -induced RF, especially over the Southern Hemisphere (SH) and NH high latitudes (Fig. 2b). The global annual mean rapid adjustment for SO_4 has a

FIG. 2. (top) Zonal-mean (scaled by the cosine of latitude) TOA RF (black lines), rapid adjustment (blue lines), and response (orange lines) caused by (a) BC and (b) SO₄. (middle) The rapid adjustments caused by (c) BC and (d) SO₄ further decomposed into cloud (pink lines) and noncloud (gray lines) components; (bottom) the responses caused by (e) BC and (f) SO₄ also decomposed into cloud (pink lines) and noncloud (gray lines) components. All variables are normalized by the absolute values of the corresponding global-mean RF; thus, their units are all 1 and the results from different experiments (e.g., 1BC, 5BC, and 10BC) can be averaged.

magnitude of 0.35 (0.18 to 0.49) relative to RF, which is primarily attributable to the cloud component, especially over the NH low and midlatitudes (Fig. 2d), where middle and low cloud covers increase significantly (not

shown). The cloud component of the rapid adjustment caused by SO₄ is a manifestation of the SO₄ cloud lifetime effect. However, the response to SO₄ is largely attributable to the noncloud component over most

FIG. 3. (a) Fast precipitation response (%) vs instantaneous atmospheric absorption (W m^{-2}) and (b) slow precipitation response (%) vs surface temperature change (K).

latitudes (Fig. 2f). The global annual mean cloud and noncloud response have normalized magnitudes relative to the SO_4 -induced RF of -0.30 (-0.34 to -0.28) and -0.69 (-0.87 to -0.46), respectively. There is a residual (about 0.35 RF) left for SO_4 , which is offset by the change in oceanic heat storage. It is also apparent from Fig. 2f that the cloud response to SO_4 is opposite in sign on the two sides of the equator, with a positive sign on the south side and a negative sign on the north side, which amplifies the original interhemispheric energy contrast perturbation induced by SO_4 , consistent with the ECHAM6-TTT model in Voigt et al. (2014). The contrast in the cloud response to SO_4 on the two sides of the equator is consistent with the changes in middle and high clouds, but opposite to the change in low cloud (not shown). The noncloud response to SO_4 can change the energy contrast in the vicinity of the equator (Fig. 2f), but overall it does not amplify the original interhemispheric energy contrast perturbation induced by SO_4 . This is not consistent with Clark et al. (2018) as they found that water vapor feedback can obviously amplify the original interhemispheric energy contrast perturbation induced by altering incoming solar radiation at NH low and high latitudes. But it should be noted that in Clark et al. (2018), cloud radiative effects were not considered.

The analysis described above suggests that BC and SO_4 interact with the climate in quite different ways. For BC, rapid adjustment is of great importance in the climate response to its RF, leaving a small energy imbalance for the climate to respond; and the rapid adjustment caused by BC can be largely attributed to the adjustment of clouds. These findings imply that the rapid adjustments to BC-induced forcing are sensitive to how BC and clouds interact with each other. Therefore, estimates of the climate responses to BC forcing depend heavily on model representations of the hygroscopicity

of BC, the internal mixing between BC and other hygroscopic aerosols, and the performance of cloud schemes. For SO_4 , however, a slow response takes the primary role in the response to the RF, particularly in the form of the noncloud response largely due to column water vapor change, and the rapid adjustment stemming from the SO_4 cloud lifetime effect amplifies the RF.

b. Global-mean precipitation

Previous studies have demonstrated that the total precipitation response to a given forcing can be decomposed into fast and slow components. We first estimate the fast and total precipitation responses caused by a specific type of aerosol from AGCM and CGCM simulations, respectively; then, the slow precipitation response to the aerosol can be diagnosed by subtracting the fast precipitation response from the corresponding total precipitation response (Samset et al. 2016).

Figure 3a shows that the fast precipitation responses (both to BC and SO_4) have a good linear relationship with the corresponding instantaneous atmospheric absorptions, derived as the RF difference between the TOA and the surface, with a regression coefficient of about -0.5% (W m^{-2}) $^{-1}$. This indicates that a certain fraction of the aerosol-induced instantaneous absorption is balanced by the fast response of precipitation (or evaporation). On the other hand, Fig. 3b shows that the slow responses of precipitation to BC and SO_4 are linearly related with the corresponding global-mean surface temperature changes, with a regression coefficient of about 3.0% K^{-1} , close to the multimodel ensemble-mean hydrological sensitivity of SO_4 reported in Samset et al. (2016) and of many other forcing drivers when the effects of rapid adjustments are removed (Samset et al. 2018).

Figure 3b also contains the slow precipitation responses and global-mean surface temperature changes

caused by 10BC and 5SO₄ obtained from the SOM simulations. It can be seen that the two points from the SOM simulations are generally along the regression line fitted to the CGCM simulations, suggesting that the linear relationship between the slow precipitation response and global-mean surface temperature change is not affected by changing ocean models. However, the absolute values of the slow precipitation responses and global-mean surface temperature changes caused by 10BC and 5SO₄ in the SOM are larger than their counterpart values in the CGCM. It is found that the oceanic heat storage changes are smaller in the SOM than in the CGCM, with a given forcing (see section 3d; Figs. 7c–f). This means that the ocean heat uptake is smaller in the SOM than in the CGCM, and thus the atmospheric response (manifesting as the global-mean surface temperature changes and the slow precipitation responses to 10BC and 5SO₄ here) to a given instantaneous forcing in the SOM tends to be larger than that in the CGCM.

By comparing Figs. 3a and 3b, it is found that the total precipitation response caused by BC is dominated by its fast component, whereas the total precipitation response caused by SO₄ is dominated by its slow component. In global annual mean terms, the total change in the atmospheric radiation budget (the net radiation flux change at the TOA minus that at the surface) is mainly balanced by the total precipitation (or evaporation) response. Therefore, the components of the total atmospheric radiation budget changes caused by BC and SO₄ (see section 3d; Figs. 7c and 7d) are the key factors to understanding the differing impacts of these two types of aerosols on precipitation. For BC, instantaneous atmospheric absorption, which is balanced by the fast precipitation response, dominates the total atmospheric radiation budget change. Whereas, for SO₄, the instantaneous atmospheric absorption is negligible, and the response in atmospheric absorption (response at the TOA minus that at the surface), which is balanced by the slow precipitation response, dominates the total atmospheric radiation budget change.

According to Suzuki et al. (2017), the response in atmospheric absorption is linearly related with the global-mean surface temperature change, and thus can be regarded as a quantity that connects the slow precipitation response and global-mean surface temperature change. This relationship can be written in a way that is analogous to Eq. (7) in Suzuki et al. (2017), to be tested with our model results, as follows:

$$-L\Delta P_{\text{slow}} \sim \text{RPS}_{\text{atm}} \sim \text{RPS}_{\text{atm,clr}} \approx \frac{\alpha\Delta W}{W} \approx \alpha\kappa\Delta T_s. \quad (2)$$

In Eq. (2), L is specific latent heat; ΔP_{slow} is the slow precipitation response; RPS_{atm} and $\text{RPS}_{\text{atm,clr}}$ represent

the all-sky and clear-sky responses in atmospheric absorption, respectively; $\Delta W/W$ is the response in atmospheric column water vapor; ΔT_s is global-mean surface temperature change; and α and κ are empirical parameters that associate $\Delta W/W$ with the response in atmospheric absorption and ΔT_s , respectively. Unlike Eq. (7) in Suzuki et al. (2017), a cloud term is not included in Eq. (2), as the contribution from clouds to the response in atmospheric absorption is very small in our model.

Suzuki et al. (2017) found that global precipitation increased with increasing surface temperature at a rate of $\sim 2\% \text{ K}^{-1}$, based on the twentieth-century historical climate change experiments of Geophysical Fluid Dynamics Laboratory Coupled Model, version 3 (GFDL3). It should be noted that Suzuki et al. (2017) did not separate precipitation change into fast and slow components, and considered both the changes in greenhouse gases and aerosols. Nonetheless, testing Eq. (2) through comparison with Suzuki et al. (2017) can still help us to understand the regression coefficient between the slow precipitation response and global-mean surface temperature change for aerosols ($3.0\% \text{ K}^{-1}$) simulated in this study.

The parameter κ is obtained from the regression between aerosol-caused $\Delta W/W$ and ΔT_s (Fig. 4a) and has a value of about $6\% \text{ K}^{-1}$, consistent with that in Suzuki et al. (2017) and that derived from the Clausius–Clapeyron relationship (Boer 1993; Pall et al. 2007). The term $\alpha\kappa$ can be calculated from the regression between the responses in atmospheric absorption and the global-mean surface temperature changes caused by BC and SO₄ (Fig. 4a) and is about $-2.34 \text{ W m}^{-2} \text{ K}^{-1}$, which is more negative than the value of $-1.56 \text{ W m}^{-2} \text{ K}^{-1}$ reported in Suzuki et al. (2017). As the parameter κ is the same as that in Suzuki et al. (2017), α (about $-0.4 \text{ W m}^{-2} \%^{-1}$) is more negative than the value of $-0.26 \text{ W m}^{-2} \%^{-1}$ reported in Suzuki et al. (2017). The term α is the ratio between the response in atmospheric absorption and the response in atmospheric column water vapor (Stephens and Ellis 2008). To understand why the α simulated here is more negative, we further separate the aerosol-induced response in atmospheric absorption into shortwave and longwave components (Fig. 4b). The regression coefficient between the aerosol-induced shortwave response in atmospheric absorption and the response in atmospheric column water vapor is about $0.15 \text{ W m}^{-2} \%^{-1}$, which is the same as that in Suzuki et al. (2017). However, the regression coefficient between the aerosol-induced longwave response in atmospheric absorption and the response in atmospheric column water vapor (about $-0.54 \text{ W m}^{-2} \%^{-1}$) is more negative than the value of $-0.42 \text{ W m}^{-2} \%^{-1}$ reported in Suzuki et al. (2017).

We simulate a larger negative regression coefficient between the longwave atmospheric absorption response

FIG. 4. (a) The responses of atmospheric column water vapor ($\Delta W/W$; blue points; %) and the responses of atmospheric absorption (brown points; W m^{-2}) vs the global-mean surface temperature changes ΔT_s (K) caused by aerosols. (b) The shortwave (blue points; W m^{-2}) and longwave (brown points; W m^{-2}) atmospheric absorption responses vs $\Delta W/W$ (%) caused by aerosols (diamond points for SO_4 , triangle points for BC).

and column water vapor response than that in Suzuki et al. (2017). The reason is that the longwave atmospheric absorption response in Suzuki et al. (2017) includes the absorption of greenhouse gases, which offsets some of the longwave atmospheric cooling caused by increased column water vapor. Besides BC, the hydrological sensitivities of greenhouse gases increase to different degrees when subtracting the fast precipitation response (Samset et al. 2018). For greenhouse gases, subtracting the fast precipitation response means minimizing the offsetting effect of their absorptions on the longwave atmospheric cooling caused by increased column water vapor. For BC, the offsetting effect is from both its longwave and shortwave absorptions, which explains why the hydrological sensitivity of BC shifts from negative to positive when removing the fast precipitation response. The relationship between the atmospheric absorption response and column water vapor response provides a united viewpoint to understanding the connection between the fast precipitation response (scaling with instantaneous absorption) and the slow precipitation response (scaling with surface temperature change) to a specific forcing agent. Besides the optical properties of forcing agents, the atmospheric temperature profile changes induced by different forcing agents (e.g., Hansen et al. 2002; Sherwood and Nishant 2015; Stjern et al. 2017) can also impose some influences on the relationship between the atmospheric absorption response and column water vapor response.

In summary, the total precipitation responses caused by BC and SO_4 are dominated by different components: the fast component for BC and the slow component for SO_4 . This is constrained by the different ways in which the two types of aerosols change the

atmospheric radiation budget. The linear relationship between the slow precipitation responses and global-mean surface temperature changes caused by BC and SO_4 indeed reflects the constraint of the response in atmospheric absorption, which is mainly associated with atmospheric column water vapor change, on the precipitation response.

c. ITCZ

Figure 5 shows the zonal-mean fast and slow precipitation responses caused by BC and SO_4 . The zonal-mean fast precipitation response caused by BC indicates a small northward shift of the ITCZ (Fig. 5a), whereas the zonal-mean slow precipitation response caused by BC shows very little northward shift of the ITCZ (Fig. 5b). It is worth noting that the 1BC case even causes a southward shift of the ITCZ over the Pacific (not shown). As introduced in section 2, we did not prescribe the OC concentration in our experiments, but fixed its emission level as that in 1850. The total emission of OC in 1850 is about 3 times that of BC in 2000. Therefore, for the 1BC case, the results are easily influenced by the response of the OC concentration in the atmosphere. From Figs. 5c and 5d, the zonal-mean fast precipitation response caused by SO_4 increases slightly near the equator, but with no substantial ITCZ shift, whereas the zonal-mean slow precipitation response caused by SO_4 indicates an obvious southward shift of the ITCZ. The ITCZ shifts caused by BC and SO_4 are to some extent consistent with the fast and slow responses of the East Asian summer monsoon to these two types of aerosols in Wang et al. (2017).

The meridional shift of the ITCZ has been shown to bear a close relationship with the change in atmospheric

FIG. 5. (left) The zonal-mean fast precipitation responses (mm day^{-1}) caused by (a) BC and (c) SO_4 . (right) The zonal-mean slow precipitation responses (mm day^{-1}) caused by (b) BC and (d) SO_4 .

cross-equatorial heat transport (e.g., Donohoe et al. 2013; Haywood et al. 2016; Hawcroft et al. 2017; Kay et al. 2016; Voigt et al. 2017). Figure 6a shows a scatterplot of ITCZ displacement versus the change in atmospheric cross-equatorial heat transport. The latitudinal position of the ITCZ is calculated as the precipitation centroid between 15°S and 15°N, and the atmospheric cross-equatorial heat transport is obtained by subtracting the cross-equatorial heat transport in the ocean from the total cross-equatorial heat transport. The computing methods of the total and oceanic meridional heat transport were introduced in many previous studies (e.g., Hawcroft et al. 2017). Figure 6a illustrates that the meridional displacements of the ITCZ caused by BC and SO_4 have a good linear relationship with the corresponding changes in atmospheric cross-equatorial heat transport, with a regression coefficient of about -3°PW^{-1} . This is close to the value of -3.2° to $-2.4^\circ \text{PW}^{-1}$ in Donohoe et al. (2013) and -4°PW^{-1} in Mahajan et al. (2013), but much less negative than the value of -7.5° to -6°PW^{-1} reported in Voigt et al. (2017).

The ITCZ displacements caused by BC and SO_4 are compared between the CGCM and SOM using the 10BC and 5 SO_4 cases, respectively (Fig. 6b). It is found that the respective northward and southward shifts of the ITCZ caused by 10BC and 5 SO_4 are greater in the SOM than in the CGCM. The meridional displacements of the ITCZ versus the changes in atmospheric cross-equatorial heat transport caused by 10BC and 5 SO_4 in the SOM are also plotted in Fig. 6a (the two quadrature points). It is found that the meridional displacement of the ITCZ and the change in atmospheric cross-equatorial heat transport caused by 5 SO_4 in the SOM are even greater than those by 10 SO_4 in the CGCM. It is obvious that, in the SOM, the ITCZ displacements and the changes in atmospheric cross-equatorial heat transport caused by 10BC and 5 SO_4 are more pronounced compared with those in the CGCM. Nevertheless, the ITCZ displacements versus the changes in atmospheric cross-equatorial heat transport simulated in the SOM and CGCM are generally along the same regression line (Fig. 6a), further demonstrating that the ITCZ shifts caused by BC and

FIG. 6. (a) The ITCZ shifts ($^{\circ}$) vs the atmospheric cross-equatorial heat transport changes (PW) caused by aerosols (orange crosses and blue stars are for BC and SO_4 in the CGCM, respectively; brown and deep blue quadrangle points are for 10BC and 5SO_4 in the SOM, respectively). (b) The zonal-mean total precipitation response (mm day^{-1}) caused by 10BC ($dP_{\text{total_10BC}}$, orange lines) and 5SO_4 ($dP_{\text{total_5SO4}}$, blue lines) in the CGCM (solid lines) and SOM (dashed lines).

SO_4 are manifestations of the changes in atmospheric cross-equatorial heat transport.

d. Hemispheric energy balance and transport

To understand the different impacts of 10BC (5SO_4) on atmospheric cross-equatorial heat transport, and thereby on the latitudinal position of the ITCZ under different ocean conditions, the changes in the hemispheric energy balance and cross-equatorial heat transport—both in the atmosphere and the ocean—caused by 10BC (5SO_4) are compared among the AGCM, CGCM, and SOM (Fig. 7). Cross-equatorial heat transport occurs as a response to interhemispheric energy contrast, and in equilibrium the ratio between cross-equatorial heat transport and interhemispheric energy contrast should be around 0.5. This suggests that all components influencing the interhemispheric contrast of the atmospheric energy budget also contribute to the change in

atmospheric cross-equatorial heat transport. The contributions from different components to the atmospheric cross-equatorial heat transport changes are estimated and compared for the 10BC and 5SO_4 cases (Table 2).

In the AGCM, the SST is fixed, and thus the imbalances of energy caused by 10BC and 5SO_4 at the surface do not change the surface temperature. This phenomenon can possibly be interpreted by introducing a hypothetical ocean concept (represented by the dashed boxes in Figs. 7a and 7b). That is, the fixed-SST ocean behaves as if the energy imbalances caused by 10BC and 5SO_4 at the surface are “absorbed” or “counteracted” by the hypothetical ocean rapidly. The interhemispheric energy imbalances caused by 10BC and 5SO_4 are also counteracted by the hypothetical ocean rapidly, as if the hypothetical ocean transports heat in the meridional direction rapidly. For 10BC, the interhemispheric energy imbalance in the atmosphere (more instantaneous atmospheric absorption in the NH than in the SH) occurs to induce the atmospheric cross-equatorial heat transport anomaly besides the quick hypothetical ocean response to the surface interhemispheric energy imbalance. The 10BC case thus causes southward anomalous cross-equatorial heat transports in both the atmosphere and the hypothetical ocean in the AGCM (Fig. 7a). For 5SO_4 , however, a cross-equatorial heat transport anomaly occurs almost solely in the hypothetical ocean and is negligible in the atmosphere in the AGCM (Fig. 7b), since SO_4 has little instantaneous shortwave absorption. This also explains why the fast precipitation response caused by SO_4 shows no ITCZ shift (Fig. 5c).

In the CGCM, both 10BC and 5SO_4 result in larger cross-equatorial heat transport anomalies in the atmosphere and smaller cross-equatorial heat transport anomalies in the ocean than they do in the AGCM (Figs. 7c,d). The reason is that, compared with the fixed SST (the quick hypothetical ocean) in the AGCM, the fully coupled ocean in the CGCM reacts more slowly in counteracting the interhemispheric energy imbalances caused by 10BC and 5SO_4 at the surface, sparing time for the atmosphere to respond. For 10BC, the southward atmospheric cross-equatorial heat transport anomaly in the AGCM is basically driven by the interhemispheric contrast in instantaneous atmospheric absorption, the majority of which is offset by the adjustment of evaporation (second column of Table 2). In the CGCM, the response of radiation (evaporation) to 10BC weakens (enhances) the 10BC-induced southward atmospheric cross-equatorial heat transport anomaly, resulting in a small net increase in the 10BC-induced southward atmospheric cross-equatorial heat transport anomaly, as compared with that in the AGCM (third column of

FIG. 7. The changes in hemisphere-mean energy balance and interhemispheric heat transport (W m^{-2}) caused by (a),(c),(e) 10BC and (b),(d),(f) 5SO₄ in the (a),(b) AGCM, (c),(d) CGCM, and (e),(f) SOM simulations; F , LH , and SH represent net radiation flux, latent heat, and sensible heat, respectively. Boxed values in the ocean represent the changes in ocean heat storage.

Table 2). For 5SO₄, the northward atmospheric cross-equatorial heat transport anomaly in the AGCM is very small, and the increase in the northward atmospheric cross-equatorial heat transport anomaly in the CGCM is mainly attributable to the response of evaporation (sixth column of Table 2).

In the SOM, even more cross-equatorial heat transport anomalies caused by 10BC and 5SO₄ are undertaken by the atmosphere than that in the CGCM

(Figs. 7e,f). In the SOM, the Q flux is fixed, which means that the heat transport in the ocean is constant. Therefore, the atmosphere has to take the primary role in responding to the interhemispheric energy imbalances caused by 10BC and 5SO₄. It is found that, for 10BC, the weakening effect of the radiation response to the southward atmospheric cross-equatorial heat transport anomaly in the SOM is smaller than that in the CGCM (fourth column of Table 2), which is the main reason

TABLE 2. Different contributions (W m^{-2}) to the change in atmospheric cross-equatorial heat transport, where northward is positive. In the columns for the CGCM and SOM, “AGCM + x ” means the value in the AGCM plus the contribution from the response. “Abs.” is short for instantaneous absorption in the atmosphere; “Adj.-atm” and “Adj.-cld” represent the noncloud and cloud components of rapid adjustment diagnosed in the atmosphere, respectively.

	10BC			5SO ₄		
	AGCM	CGCM	SOM	AGCM	CGCM	SOM
Radiation	Abs.	−1.98	AGCM + 0.15	Abs.	+0.15	AGCM + 0.53
	Adj.-atm	+0.24	AGCM + 0.02	Adj.-atm	−0.325	AGCM + 1.43
	Adj.-cld	+0.02		Adj.-cld	+0.08	
Evaporation	+1.04	AGCM − 0.26	AGCM − 0.24	−0.275	AGCM + 1.3	AGCM + 3.5
Sensible heat	+0.48	AGCM + 0.07	AGCM + 0.04	+0.39	AGCM − 0.28	AGCM − 1.03
Total	−0.2	AGCM − 0.04	AGCM − 0.18	+0.02	AGCM + 1.55	AGCM + 3.9

why the southward atmospheric cross-equatorial heat transport anomaly caused by 10BC in the SOM is larger than that in the CGCM. For 5SO₄, the larger northward atmospheric cross-equatorial heat transport anomaly in the SOM than that in the CGCM is mostly attributable to the response of evaporation and to a lesser extent to the response of radiation (seventh column of Table 2). It is expected that the anomalies of oceanic cross-equatorial heat transport and ocean heat storage caused by 10BC and 5SO₄ in the SOM should be zero if the running time is long enough and a real equilibrium climate state is achieved, as the Q flux is fixed in SOM simulations. Even though, it is seen that the oceanic cross-equatorial heat transport anomalies caused by 10BC and 5SO₄ are negligible compared with their atmospheric counterparts in the SOM (Figs. 7e,f).

The comparisons above demonstrate that, for a specific forcing agent, the ways in which the climate system responds to the interhemispheric energy imbalance can be very different depending on the oceanic conditions. The change in atmospheric cross-equatorial heat transport caused by aerosols, which determines the ITCZ shift, is underestimated in the AGCM and exaggerated in the SOM, as compared with that in the CGCM. This underscores the importance of employing a fully coupled ocean in modeling studies of tropical precipitation responses.

The different ways in which BC and SO₄ affect the position of the ITCZ are summarized in two schematics (Fig. 8), based on the AGCM and CGCM simulations. To simplify the story, only the largest contributors to the atmospheric cross-equatorial heat transport anomalies caused by BC and SO₄ (Table 2) are given in the schematics. Both BC and SO₄ suppress evaporation more greatly in the NH than they do in the SH (Figs. 7c,d), which means that the two types of aerosols increase the evaporation in the SH relative to the NH. For BC, its instantaneous atmospheric absorption in the NH basically drives the southward atmospheric cross-equatorial

heat transport anomaly, which forces a northward cross-equatorial moisture transport anomaly (Fig. S2a) and causes the ITCZ to shift northwardly. However, for SO₄, the largest contributor to the northward cross-equatorial heat transport anomaly is the response of evaporation (i.e., relatively more evaporation in the SH). Consequently, the northward cross-equatorial heat transport anomaly caused by SO₄ forces southward cross-equatorial moisture transport (Fig. S2b), even though there is already more evaporation in the SH. In summary, BC causes the ITCZ to shift northward from its more instantaneous atmospheric absorption in the NH, whereas SO₄ causes the ITCZ to shift southward, mainly through its excessive suppression of evaporation in the NH. The two processes happen over two different time scales.

4. Conclusions

This study explored the differing impacts of BC and SO₄ aerosols on global and tropical precipitation, using the climate model MIROC5.2. BC and SO₄ have quite different optical properties, resulting in positive and negative RF at the TOA, respectively. Results show that rapid adjustment, particularly cloud rapid adjustment, plays the primary role in the climate response to the RF of BC. For SO₄, in contrast, the noncloud response plays the largest role in the response to the RF at the TOA; the rapid adjustment, which results mainly from the cloud lifetime effect of SO₄, amplifies the SO₄ RF by about 35% in our model.

The total precipitation responses caused by BC and SO₄ are dominated by different components. For BC, the fast precipitation response, which scales with instantaneous atmospheric absorption by approximately $-0.5\% (\text{W m}^{-2})^{-1}$, dominates the total precipitation response. However, for SO₄, a slow precipitation response scaling with global-mean surface temperature change according to $\sim 3.0\% \text{ K}^{-1}$ dominates the total

FIG. 8. Schematic representations of the mechanisms by which (a) BC and (b) SO_4 impact the position of the ITCZ. Blue arrows: moisture transport, including evaporation (Evap.) and cross-equatorial moisture transport (Moist.); orange arrows: dry energy transport; evaporation and moisture next to orange arrows stand for the energy transport equivalent to them; Abs. represents the instantaneous absorption caused by BC in the atmosphere; Δ means interhemispheric contrast; pink circles show the anomalous circulations caused by BC and SO_4 with the anomalous cross-equatorial energy and moisture transports as their upper and lower branches.

precipitation response. The difference between BC and SO_4 in terms of the proportion of the precipitation response is attributable to the different ways in which the two types of aerosols change the atmospheric radiation budget. For BC, instantaneous atmospheric absorption dominates the total change in the atmospheric radiation budget. However, for SO_4 , the noncloud response—mainly associated with column water vapor change—dominates the total change in the atmospheric radiation budget.

By changing the atmospheric cross-equatorial heat transport, BC and SO_4 cause northward and southward shifts of the ITCZ, respectively. The ITCZ displacements and the atmospheric cross-equatorial heat transport changes caused by the two types of aerosols have a good linear relationship, with a regression coefficient of about -3°PW^{-1} . The BC-induced northward shift of the ITCZ takes place mainly through the fast precipitation response, as the basic driving force for the BC-induced southward atmospheric cross-equatorial heat transport anomaly is the interhemispheric contrast in the instantaneous atmospheric absorption caused by BC. In contrast, the SO_4 -induced southward shift of the ITCZ takes place through the slow precipitation response, and the largest contributor to the SO_4 -induced northward atmospheric cross-equatorial heat transport anomaly is the evaporation response (i.e., evaporation is suppressed more greatly in the NH than in the SH).

By comparing the precipitation responses caused by BC and SO_4 simulated in the CGCM and SOM, it is found that the SOM exaggerates both the global mean precipitation response and the ITCZ shift, and the latter is exaggerated much more obviously than the former. The SOM exaggerates the ITCZ shifts caused by BC and SO_4 by suppressing the role of the ocean in the meridional heat transport response. This underscores the importance of employing a fully coupled ocean model when studying the tropical climate response to aerosol forcing.

In general, this work demonstrates that the different precipitation responses caused by BC and SO_4 , both on global and tropical scales, are determined by their different ways in changing the atmospheric energy budget and its hemispheric asymmetry. However, uncertainties still exist and future work is needed. For example, we obtained a large negative rapid adjustment (about -0.7RF) for BC in section 3a, but previous studies have also suggested various magnitudes of rapid adjustment for BC, including those much smaller than in this study (Bond et al. 2013). On the other hand, clouds were shown to play a primary role in determining the rapid adjustment of BC in this study; plus, Kang et al. (2009) and Voigt et al. (2014) suggested that changes to the treatment of clouds could exert a considerable impact on the atmospheric energy transport over the tropics, and thereby on the ITCZ shift, implying that it is

important to study the precipitation responses caused by aerosols with different cloud schemes.

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