Response of surface shortwave cloud radiative effect to greenhouse gases and aerosols and its impact on summer maximum temperature

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Abstract. Shortwave cloud radiative effects (SWCRE), defined as the difference of shortwave radiative flux between all-sky and clear-sky conditions at the surface, have been reported to play an important role in influencing the Earth's energy budget and temperature extremes. In this study, we employed a set of global climate models to examine the SWCRE responses to CO_2, black carbon (BC) aerosols and sulfate aerosols in boreal summer over the Northern Hemisphere. We found that CO_2 causes positive SWCRE changes over most of the NH, and BC causes similar positive responses over North America, Europe and East China but negative SWCRE over India and tropical Africa. When normalized by effective radiative forcing, the SWCRE from BC is roughly 3-5 times larger than that from CO_2. SWCRE change is mainly due to cloud cover changes resulting from the changes in relative humidity (RH) and, to a lesser extent, changes in cloud liquid water, circulation, dynamics and stability. The SWCRE response to sulfate aerosols, however, is negligible compared to that for CO_2 and BC because part of the radiation scattered by clouds under all-sky conditions will also be scattered by aerosols under clear-sky conditions. Using a multilinear regression model, it is found that mean daily maximum temperature (Tmax) increases by 0.15 K and 0.13 K per W m^{-2} increase in local SWCRE under the CO_2 and BC experiment, respectively. When domain-averaged, the contribution of SWCRE change to summer mean Tmax changes was 10-30% under CO_2 forcing and 30-50% under BC forcing, varying by regions, which can have important implications for extreme climatic events and socio-economic activities.

1 Introduction

Clouds have a pivotal role in influencing the Earth's energy budget (Ramanathan et al., 1989). By enhancing the planetary albedo, clouds exert a global mean shortwave cloud radiative effects (SWCRE) of about -50 W m^{-2} at the top-of-the-atmosphere, and by contributing to the greenhouse effect, exert a mean longwave effect (LWCRE) of approximately +30 W m^{-2} (Boucher et al., 2013). On the whole, clouds cause a net forcing of -20 W m^{-2} relative to a cloud-free Earth, which is approximately five times as large as the radiative forcing from a doubling of CO_2 concentration. Therefore, a subtle change in cloud properties has the potential to cause significant impacts on climate (Boucher et al., 2013; Zelinka et al., 2017). Recent studies contended that the cloud feedback, especially the shortwave (SW) cloud feedback, is very likely to be positive (Clement et al., 2009; Dessler, 2010; Zelinka et al., 2017). As the SW cloud feedback is positively correlated with the net climate feedback parameter (Andrews et al., 2012; Andrews et al., 2015; Zhou et al., 2016), a stronger positive SW cloud feedback will lead to higher climate sensitivity and may lead to a future warming towards the high end of current projections (Zhai et al., 2015; Andrews et al., 2018).

On seasonal scales, SWCRE is strongest in the summer months when the solar heating is strongest (Harrison et al., 1990). Because SWCRE is in effect only during daytime, it can substantially modify daily maximum temperature (Tmax). For instance, Dai et al. (1999) found that increased cloud cover can reduce Tmax, thereby decreasing diurnal temperature range. Tang and Leng (2012) reported that the damped Tmax over Eurasia could be partially explained by the cloud cover increase during 1982-2009. As a positive feedback, SWCRE at the surface has also been reported to play a role in heatwave and drought events over Europe by enhancing solar heating (Rowell & Jones, 2006; Vautard et al., 2007; Zampieri et al., 2009; Chiriaco et al., 2014; Myers et al., 2018). This has influenced the environment, ecosystems and the economy through affecting the frequency and intensity of forest fires, power cuts, transport restrictions, crop failure and loss of life (De Bono et al., 2004; Ciais et al., 2005; Robine et al., 2008). For example, Wetherald and Manabe (1995) reported that in the summer for mid-latitude continents, higher temperature enhances evaporation in the spring and then evaporation decreases in the summer due to depleted soil moisture. Combined with higher temperature, this summertime evaporation reduction leads to lower relative humidity (RH), which reduces cloud cover and thereby invigorates solar heating. Cheruy et al. (2014) revealed that the inter-model spread of summer temperature projections in Northern mid-latitudes in CMIP5 (Climate Model Inter-comparison Project Phase 5) models is greatly influenced by SWCRE.

All the above studies suggest that the SWCRE plays an important role in influencing the surface energy budget and extreme temperature. Well-mixed greenhouse gases (WMGHGs) and aerosols are currently the two largest anthropogenic forcings (Myhre et al., 2013b). A better understanding on the climate response to these individual forcing agents is increasingly needed, considering their different trends across the globe and opposite impacts on climate (Shindell & Faluvegi, 2009). Due to the difficulty of separating the forced climate signal of a single agent within observational records, these studies are generally based on model simulations, such as the widely used quadrupling of CO_2 experiments (Andrews et al., 2012). Many attempts have also been made to explore the aerosol impact on clouds and Earth's energy balance (Lohmann & Feichter, 2005; Chung & Soden, 2017), mean temperature (Ruckstuhl et al., 2008; Philipona et al., 2009), as well as extreme temperature (Sillmann et al., 2013; Xu et al., 2018). However, all these studies treated aerosols as a whole and the individual impacts from absorbing and scattering aerosols are still less understood. Though some studies investigated the impact from individual aerosol species (Williams et al., 2001; Chuang et al., 2002; Koch & Del Genio, 2010), they generally used only a single model, and the results may be subject to model biases (Flato et al., 2013). Moreover, due to the continuing increase in the likelihood of hot temperature extremes (Seneviratne et al., 2014), as well as their serious consequences (De Bono et al., 2004), it is imperative to have a better understanding on the role of SWCRE from individual forcing agents in hot extremes. However, a multi-model study on the cloud response to individual aerosol species and the impact of that response on Tmax is still lacking. Given these knowledge gaps, here we investigate the changes of SWCRE to CO_2, BC and sulfate aerosols individually and explore its potential impact on Tmax by using a set of state-of-the-art global climate models. CO_2 is the most dominant WMGHG while the latter two represent absorbing and scattering aerosols respectively. This paper will proceed as follows: data and methods are described in Section 2. Results are presented in Section 3, discussions and summary are given in Section 4.

2 Data and Methods

2.1 Data

This study employs the model output from groups participating in the Precipitation Driver and Response Model Intercomparison Project (PDRMIP), utilizing simulations examining the climate responses to individual climate drivers (Myhre et al., 2017). The nine models used in this study are CanESM2, GISS-E2R, HadGEM2, HadGEM3, MIROC, CESM-CAM4, CESM-CAM5, NorESM and IPSL-CM5A. The versions of most models used in the PDRMIP are essentially the same as their CMIP5 versions. The configurations and basic settings are listed in Table 1. In these simulations, global-scale perturbations were applied to all the models: a doubling of CO_2 concentration (CO_2×2), a tenfold increase of present-day black carbon concentration/emission (BC×10), and a fivefold increase of present-day SO_4 concentration/emission (SO_4×5). All perturbations were abrupt. Each perturbation was run in two parallel configurations, a 15-year fixed sea surface temperature (fsst) simulation and a 100-year coupled simulation. One model (CESM-CAM4) used a slab ocean setup for the coupled simulation whereas the others used a full dynamic ocean. CO_2 was applied relative to the models' baseline values. For aerosol perturbations, monthly year 2000 concentrations were derived from the AeroCom Phase II initiative (Myhre et al., 2013a) and multiplied by the stated factors in concentration-driven models. Some models were unable to perform simulations with prescribed concentrations. These models multiplied emissions by these factors instead (Table 1). The aerosol loadings in the CanESM2 model for the two aerosol perturbations are shown in Fig. 1 for illustrative purpose; the spatial patterns are similar for other models. In the BC experiment, the concentration is highest in East China (E. China), followed by India and tropical Africa. For the SO_4 simulations, the aerosols are mainly restricted to the Northern Hemisphere (NH), with the highest loading observed in E. China, followed by India and Europe. The eastern US also has moderately high concentrations. It is noted that only three of the nine models include aerosol-cloud interactions while the remaining ones only have aerosol-radiation interactions. However, this does not impact our main conclusions (see section 4). More detailed descriptions of PDRMIP and its initial findings are given in Samset et al. (2016), Myhre et al. (2017), Liu et al. (2018) and Tang et al. (2018).

2.2 Methods

In this study, we focus on the SWCRE at the surface in the low and mid-latitudes during boreal summer months (June-July-August, JJA hereafter), which is calculated as the difference in the SW radiative flux at the surface between all-sky and clear-sky conditions (Ramanathan et al., 1989). The base state of SWCRE in each model is shown in Fig. S1, with a multi-model mean (MMM) value of -57.9±1.8 W m^{-2} (MMM±1 standard error). The spatial patterns are fairly consistent across the models, with strong SWCRE in tropical regions and mid-to-high latitudes and weaker SWCRE in subtropics, regions generally with less clouds. Changes in SWCRE are obtained by subtracting the control simulations from the perturbations using the data of the last 20 years in each coupled simulation. The changes are then normalized by the effective radiative forcing (ERF) in the corresponding experiments to obtain the changes per unit global forcing for comparison. Previous studies demonstrated that climate changes linearly with climate forcing for various forcing agents, including BC (Hansen et al., 2005; Mahajan et al., 2013). The ERF values for each model are obtained from Tang et al. (2019), which diagnosed those from the data for years 6-15 of the fsst simulations of each perturbation by calculating the radiative flux changes at the top-of-the-atmosphere (Hansen et al., 2002). The MMM ERF values are 3.65±0.09 W m^{-2} (CO_2×2), 1.16±0.25 W m^{-2} (BC×10), and -3.52±0.63 W m^{-2} (SO_4×5) for indicated experiments, respectively (MMM±1 standard error). Then the MMM changes are estimated by averaging all the nine models' results, giving the same weighting factor to each model. A two-sided student t-test is used to examine whether the MMM results are significantly different from zero. The same process was also repeated to other variables analyzed (i.e., temperature and humidity).

In order to investigate the impact of circulation changes on specific humidity, following Banacos and Schultz (2005), the horizontal moisture flux convergence (MFC) is calculated as:

$$MFC = -\nabla \cdot (\mathrm{q}\mathrm{V}) = -V \cdot \nabla \mathrm{q} - q\nabla \cdot \mathrm{V} \quad (1)$$

In Eq (1), q is specific humidity in g kg^{-1}, and V is horizontal wind including both zonal and meridional components. All variables have a monthly temporal resolution. Equation (1) could be further written as:

$$MFC = -u\frac{\partial \mathrm{q}}{\partial \mathrm{x}} - v\frac{\partial \mathrm{q}}{\partial \mathrm{y}} - q\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \quad (2)$$

In which u and v are zonal and meridional wind components in m s^{-1}.

3 Results

3.1 SWCRE Change

Figure 2a-c show the SWCRE changes in response to abrupt changes in CO_2, BC and SO_4. CO_2 causes positive changes in SWCRE over most areas in the NH, indicating that more SW radiation reaches the surface. BC causes similar changes, but with enhanced (ERF-normalized) magnitude, especially in North America (N. America), Europe and East Asia (E. Asia). In some source regions of BC aerosols (tropical Africa and India), however, the SWCRE changes are negative, which means more SW was reflected. These changes are all statistically significant and are unlikely to be caused by natural variability. When it comes to individual model response (Fig. S2-S3), these patterns are also consistent across at least eight of the nine models and are not very sensitive to the model setup (emission-based or concentration-based). For SO_4, the SWCRE changes are relatively small compared with the other two forcings and few significant changes are found over low-to-mid latitude regions. When domain averaged (green boxes in Fig. 2), the MMM SWCRE from CO_2 forcing is, 1.7 W m^{-2} (N. America), 2.0 W m^{-2} (Europe) and 1.5 W m^{-2} (E. China) respectively for the indicated regions. The SWCRE of BC forcing is 7.0 W m^{-2} (N.

America), 9.0 W m^{-2} (Europe) and 9.4 W m^{-2} (E. China) respectively, which is roughly 3 to 5 times larger than that from CO_2 forcing whereas sulfate aerosols induced 1.2 W m^{-2} over E. China and near-zero impact in N. America and Europe, with even the sign of change being uncertain (Fig. 3 and Fig. S4). Such SWCRE changes could be largely explained by the changes of cloud cover (Fig. 2d-f). Low-level cloud cover decreased significantly in regions where SWCRE is positive for CO_2 and BC forcing, with a stronger decrease from the latter, indicating that the cloud response is more sensitive to BC forcing than to WMGHGs. The sulfate aerosols caused increased cloud cover over mid-latitudes (Fig. 2f). The cloud cover in other levels show similar patterns of change (Fig. S5). In order to better understand these cloud responses, we will explore a set of potential mechanisms driving such changes.

3.2 Mechanism of the Cloud Changes

Clouds form when air rises and cools to saturation, and are thus closely linked to changes in RH (Fig. 4a-c). The general pattern of RH changes corresponds well with cloud cover changes (Fig. 2d-f). That is, the cloud cover decreases in regions where the RH drops and vice versa for most areas. A larger RH reduction due to BC compared with CO_2 also aligns with a larger cloud cover decrease under BC forcing, especially over N. America and Europe. This spatial pattern is not surprising as it is easier for air masses to reach saturation in conditions with higher RH. By definition, RH depends on both specific humidity and saturation vapor pressure (which, in turn, depends on temperature). To probe which factor determines the RH changes, we further analyzed specific humidity changes (Fig. 4d-f). Specific humidity increases ubiquitously under both CO_2 and BC scenarios, as a result of increased evaporation in a warmer climate. Thus, the main driver of the RH drop is the atmospheric temperature that drives a faster increase of saturation vapor pressure. Figure 5 shows the changes of vapor pressure as a function of temperature change over Europe at 850 hPa. For example, the temperature increases by ~1.1 K under CO_2 forcing, accompanied by ~0.02 kPa vapor pressure increase. Such a vapor pressure increase, however, cannot keep pace with the rise in saturation vapor pressure, which is about 0.1 kPa. Consequently, the RH decreases in Europe and this is also the case for most other land areas. BC causes stronger temperature increases (and hence larger RH drop) in Europe and N. America, explaining the larger cloud cover reductions compared with CO_2. In the source regions of BC, such as India and tropical Africa, the RH increases because of stronger increases of specific humidity, combined with weak or no temperature changes (Fig. S6). The response of cloud liquid water in the BC experiment could further support this conclusion (Fig. 4h). Liquid water decreases (increases) in regions with decreasing (increasing) cloud cover, following the pattern of RH. As cloud water content directly impacts cloud optical thickness and albedo, such a response may further impact SWCRE (i.e., enhance reflectance in regions showing increasing liquid water and enhance transmittance in regions with decreasing liquid water). However, the liquid water responses under CO_2 and sulfate aerosols are much weaker, only significant in part of Asia and tropical Africa (Fig. 4g and i).

Changes in moisture flux, dynamics and stability may also play a role in altering specific humidity and cloud formation (Bretherton, 2015). Here we analyze the changes of MFC, vertical velocity (omega), as well as lower tropospheric stability (LTS), and find significant changes under the BC experiment again (Fig. 6). It is seen that more moisture is transported to tropical Africa and India (Fig. 6b), which could explain the abovementioned increases of specific humidity in these regions despite their lack of warming. A similar response was noted by Liu et al. (2018), which suggested that more moisture could be brought into monsoon regions due to BC forcing. Koch and Del Genio (2010) noted that BC particles could promote cloud cover in convergent regions as they enhance deep convection and low-level convergence when drawing in moisture from ocean to land regions. This is also observed in our analyses, for example over Africa, North India, Pakistan and part of North China (Fig. 6b and e), which is consistent with the dynamic cloud response mechanism noted by Myers and Norris (2013). However, these impacts may be further compounded by cloud type, circulation and the altitude of BC particles relative to the clouds (Koch & Del Genio, 2010; Samset & Myhre, 2015). The changes in moisture flux and dynamics in the CO_2 experiment are relatively weaker compared with those from BC, and most of the changes are only observed in low-latitude regions, possibly due to the shift of Intertropical Convergence Zone (ITCZ) or monsoon circulations. The sulfate aerosols, on the other hand, generally show opposite changes to those from CO_2 and BC (Fig. 4c and f), owing to sulfate's cooling effect. Another mechanism that has been reported to influence cloud cover is LTS, in which a stable boundary layer could trap more moisture, thereby permitting more low-level clouds (Wood & Bretherton, 2006; Bretherton, 2015). In order to investigate this mechanism, we further analyzed LTS, defined as the difference of potential temperature between 700 hPa and surface (Fig. 6g-i), in which positive anomalies indicate a stronger inversion or weaker lapse rate. The LTS response is again strongest in response to BC forcing (Fig. 6h), with a widespread increase in stability. A previously reported positive correlation between LTS and low-level cloud cover is, nonetheless, only observed in BC source regions (tropical Africa and India) and part of the central US (Fig. 6h). The LTS responses over land are much weaker in response to CO_2 and SO_4 forcing, with some responses in Africa and India in response to sulfate aerosols (weaker inversion and less cloud). Some other factors have also been suggested to play a role in modifying low-level clouds, such as the diurnal cycle (Caldwell & Bretherton, 2009) and radiative effects of cirrus clouds (Christensen et al., 2013). Due to the limited model output, however, we acknowledge that it is impossible to examine these factors in the current study and it is beyond the scope of our study to probe all possible factors driving the cloud changes. In summary, the above analyses illustrate that the cloud cover changes we see can be primarily explained by RH changes and, to a lesser extent, changes of liquid water content, circulation, dynamics, and stability.

3.3 Fast and Slow Responses

The above responses shown are total responses, which could be further split into fast responses (also called rapid adjustments) and slow responses (Andrews et al., 2010; Boucher et al., 2013). The fast responses generally occur within weeks to a few months with the global mean temperature unchanged, and also with the expectation of a small change over land, which could be obtained by fsst simulations. The slow response is mainly depending on global mean temperature change, which could be estimated by the difference between coupled simulations and fsst simulations, assuming the total response is a linear combination of fast response and slow response (Samset et al., 2016; Stjern et al., 2017). For the CO_2 experiment, fast responses dominated in E. US and Europe while both fast and slow responses influence Asia (Fig. 7). When it comes to BC, both fast and slow responses are important in these regions, and in some regions the fast and slow response even show opposite changes (e.g., N. Europe). This is consistent with the findings of Stjern et al. (2017) that the response of cloud amount under BC forcing typically consists of opposite rapid adjustments. Regarding sulfate aerosols, the SWCRE changes are much weaker, with both fast and slow responses influencing Asia and Africa. As discussed in Section 3.2, the slow responses in Asia is likely to be associated with circulation changes, as significant changes in MFC, omega and stability are observed in tropical regions and monsoon regions across all three experiments (Fig. 6). These circulation changes could be, but are not limited to, shifts in the monsoons or ITCZ and tropical expansion, and both greenhouse gases and aerosols have been reported to impact these circulations (Menon et al., 2002; Wang, 2007; Meehl et al., 2008; Seidel et al., 2008; Allen et al., 2012; Turner & Annamalai, 2012).

3.4 SWCRE Response to Sulfate Aerosol

Another interesting phenomenon worth noting is the relatively small change in SWCRE induced by sulfate aerosols compared with CO_2 and BC. SWCRE at the surface is obtained as the difference of SW fluxes between all-sky and clear-sky conditions (Fig. 8). However, both clouds and aerosol particles scatter solar radiation, so that at least part of the radiation scattered by clouds under all-sky conditions will also be scattered by aerosols under clear-sky conditions (no clouds). This means the SW radiation change at the surface due to scattering may not be as sensitive to cloud fraction changes, which leads to reduced changes in their difference (SWCRE), at least in the source regions (Fig. 8). The SWCRE under sulfate aerosols will not be further discussed due to its small radiative impact at the surface.

3.5 Impact on Radiation and Tmax

From the energy perspective, the net incoming radiation (Rin) at the surface is the combination of downward SW radiation and downward longwave (LW) radiation minus the reflected SW radiation (Rin = ↓SW − ↑SW + ↓LW). Rin represents the total energy available to maintain the surface temperature and to sustain the turbulent fluxes (Philipona et al., 2009). The surface responds to the imposed Rin by redistributing the altered energy content among the outgoing LW radiation and nonradiative fluxes (ground heat flux and turbulent flux) (Wild et al., 2004). Because SW radiation is in effect only during daytime while LW radiation works both day and night, Rin is directly related to Tmax. In a perturbed climate, both SW and LW radiation will change, thereby changing Rin and Tmax. The net SW radiation change is further linearly decomposed into SW changes under clear-sky conditions and SWCRE changes. The changes of Rin and its individual components, as well as Tmax are shown in Fig. 9. For the $CO_2\times2$ experiment, the SW under clear-sky conditions shows slight decreases over most of land surfaces, mainly due to the absorption of SW radiation by enhanced water vapor, except for some high-latitude regions where albedo effect is important (Fig. 9a). Combined with the changes of SWCRE and ↓LW radiation, Rin shows significant increases over all land surfaces and thus, increasing Tmax (Fig. 9g and i). The BC×10 experiment shows similar responses, with significantly negative SW radiation under clear-sky conditions due to SW absorption by BC particles (Fig. 9b) and enhanced ↓LW radiation resulted from atmospheric heating (Fig. 9f). The resulting Rin changes largely explained Tmax changes on the first order, with cooling observed in source regions (India and tropical Africa) and warming elsewhere (Fig. 9h and j). Nonetheless, some exceptions occurred (i.e., E. China), with decreased Rin but increased Tmax, possibly due to the atmospheric heat transport (Menon et al., 2002) and reduced turbulent fluxes (Wild et al., 2004).

In order to further determine the contributions in Tmax changes from each individual radiative component, a multilinear regression model is applied by regressing Tmax changes to SW clear-sky, SWCRE and ↓LW radiation changes with zero intercept, obtaining the following models:

$CO_2\times2$:

$$Tmax = 0.08\times SW_{clear\text{-}sky} + 0.15\times SWCRE + 0.14\times\downarrow LW \ (R^2 = 0.73,\ p < 0.001)$$

BC×10:

$$Tmax = 0.05\times SW_{clear\text{-}sky} + 0.13\times SWCRE + 0.15\times\downarrow LW \ (R^2 = 0.80,\ p < 0.001)$$

All values in the linear models are MMM changes in each experiment. The models could explain 73% and 80% of the Tmax change in $CO_2\times2$ and BC×10 experiment respectively. The coefficients represent the Tmax change under unit radiative flux change, in which the Tmax increases by 0.15 K (0.13 K) per unit increase in local SWCRE under the CO_2 (BC) experiment respectively. Furthermore, the coefficients demonstrate that Tmax changes are more sensitive to unit SWCRE and ↓LW changes than to unit $SW_{clear\text{-}sky}$. A comparison of the original Tmax values and the fitted values from the linear models is shown in Figure 10. The linear models predict the Tmax changes fairly well, with the values scattering along the one-to-one line. The contributions from each radiative component to Tmax changes were estimated with the linear models and the domain-averaged changes for N. America, Europe, E. China and India (purple boxes in Fig. 9a) are listed in Table 2. Physically, Tmax increases in these regions are mainly due to the increased flux from SWCRE and ↓LW, and partially offset by the reduced flux from $SW_{clear\text{-}sky}$ (Table 2 & Fig. 9). Taking N. America under $CO_2\times2$ experiment as an example, the warming in Tmax from SWCRE and ↓LW are 0.95 K and 3.24 K respectively, in which SWCRE contributed roughly by 23% to the total warming and the remaining 77% is from the ↓LW radiation change. Such warming is offset by the 0.27 K cooling from SW changes under clear-sky conditions, leading to a net increase of 3.92 K in Tmax. The contributions of SWCRE in Tmax increases are 29% (Europe), 20% (E. China) and 9% (India) for the indicated regions under the $CO_2\times2$ experiment. For the BC×10 experiment, the contributions from SWCRE are larger than those in the CO_2 experiment, i.e. 34% (N. America), 47% (Europe) and 34% (E. China) for each region. The response over India under the BC experiment is opposite, in which both SW components cause cooling in Tmax due to reduced fluxes and such cooling is slightly offset by the warming from increased ↓LW radiation. In this case, the negative SWCRE change contributed 54% to the reduction in Tmax. It is noted that the radiation change might not explain all Tmax changes, as other factors may come into play. For instance, the temperature response would be different when surface is getting drier under a warmer climate. This is because more net radiation is realized as sensible heat instead of latent heat under drier conditions, which has been suggested to play an important role in recent European heatwaves (Seneviratne et al., 2006; Fischer et al., 2007).

4 Discussion and Summary

Our study shows that cloud cover in the summer is reduced in a warming climate over most mid-latitude land regions. The reduction of clouds, at the same time, may also reduce the warming effect by reducing downwelling LW radiation (LWCRE, Fig. S7). Specifically, the LWCRE changes per unit CO_2 forcing, in MMM, are -1.1 W m^{-2} (N. America), -0.8 W m^{-2} (Europe) and -1.0 W m^{-2} (E. China) respectively, resulting in net CRE (SWCRE+LWCRE) changes of 0.6 W m^{-2} (N. America), 1.2 W m^{-2} (Europe) and 0.5 W m^{-2} (E. China) at the surface. The LWCRE changes per unit BC forcing are -1.7 W m^{-2} (N. America), -2.1 W m^{-2} (Europe) and -1.5 W m^{-2} (E. China) respectively, leading to net CRE changes of 5.3 W m^{-2} (N. America), 6.9 W m^{-2} (Europe) and 7.9 W m^{-2} (E. China). The net CRE changes are positive under both forcings and work as a positive feedback in these areas. As SWCRE is only active during daytime, the CRE changes have an even more pronounced amplifying effect on summer extreme temperature in these populated regions.

Recent European heatwave events have been linked to the shift of mean temperature (Schär et al., 2004; Barriopedro et al., 2011). Thus, the enhanced increase in summer mean Tmax may significantly increase the number of hot days and the probability of heatwave events. Our model simulations show that both N. America and Europe show faster increases in Tmax than in Tmin (daily minimum temperature) under both CO_2 and BC experiments (figure not shown), indicating an increase in diurnal temperature range, which has also been reported by Wang and Dillon (2014). These changes can have substantial socio-economic impacts (De Bono et al., 2004; Ciais et al., 2005), influencing human health (Robine et al., 2008), labor productivity (Kjellstrom et al., 2018), and disease transmission (Paaijmans et al., 2010), as well as environmental and other ecological functions (Vasseur David et al., 2014; Wang & Dillon, 2014).

Some limitations also exist in the current study. Firstly, aerosol-cloud interactions cannot be realistically represented, as more than half of the PDRMIP simulations were run with fixed concentrations, where changes in cloud lifetime cannot affect aerosols. For the BC simulations, three models include aerosol indirect effects (MIROC, NorESM and IPSL) while the remaining ones have only aerosol-radiation interactions included (instantaneous and rapid adjustments). The responses of SWCRE for the two categories are shown in Figure 11. For the regions of interest in the current study, the positive SWCRE over N. America, Europe and E. China and negative SWCRE over India are still observed in the models including indirect effects, but with reduced magnitude. Thus, our main conclusions hold in both sets of models, since the responses do not qualitatively vary between those with indirect effects and models without those effects. Such effects are not likely to be a large source of uncertainty but merit future study. Secondly, the aerosol perturbations are idealized time-invariant 10× and 5× present-day aerosol concentrations. Such simulations provide valuable physical insights into the effects of different forcings on a variety of aspects of the climate system. Aerosol concentrations, however, changed inhomogeneously during the historical period and in recent decades, both spatially and temporally. For example, aerosol concentrations have been decreasing in Europe and N. America since the 1980s and have been increasing in Asia since the 1950s (Smith et al., 2011). Future simulations may use aerosol forcing with realistic spatio-temporal changes.

In conclusion, our study shows that both CO_2 and BC could cause positive SWCRE changes over most regions in the NH, with a stronger response caused by BC, except over some key source regions of BC aerosols (e.g., India, tropical Africa) which show opposite changes. The SWCRE changes under sulfate aerosol forcing are, however, relatively small compared with the other two forcers. The SWCRE changes are mainly a consequence of RH changes and, to a lesser extent, liquid water, circulation, dynamics and stability changes. The SWCRE changes may have contributed 10~50% of summer mean Tmax increases, depending on forcing agent and region, and contributed substantially to Tmax decreases in the source regions of India and Africa, which has important implications for extreme climatic events and socio-economic activities.

Data code availability

The PDRMIP model output used in this study are available to public through the Norwegian FEIDE data storage facility. For more information, please see http://cicero.uio.no/en/PDRMIP. This study is performed by using Matlab R 2019a. The Matlab code is available upon reasonable request.

Competing interests

The authors declare no competing interests.

Author contributions

T.T. and D.S. designed this study. T.T. performed data analysis and wrote the initial manuscript. All authors contributed to scientific discussion, results framing and manuscript polishing.

Acknowledgement

We thank the two reviewers for their helpful and constructive comments. We acknowledge the NASA High-End Computing Program through the NASA Center for Climate Simulation at Goddard Space Flight Center for computational resources to run the GISS-E2R model and support from NASA GISS. PDRMIP is partly funded through the Norwegian Research Council project NAPEX (project number 229778). A.Voulgarakis is supported by NERC under grant NE/K500872/1. Computing resources for CESM1-CAM5 (ark:/85065/d7wd3xhc) simulations were provided by the Climate Simulation Laboratory at NCAR Computational and Information System Laboratory, sponsored by the National Science Foundation and other agencies.

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Table 1. Descriptions of the nine PDRMIP models used in this study, adapted from Tang et al. (2019).

Model name	Version	Resolution	Ocean setup	Aerosol setup	references
CanESM	2010	2.8×2.8 35 levels	Coupled	Emission	Arora et al. (2011)
GISS-E2	E2-R	2×2.5 40 levels	Coupled	Fixed concentration	Schmidt et al. (2014)
HadGEM2-ES	6.6.3	1.875×1.25 38 levels	Coupled	Emissions	Collins et al. (2011)
HadGEM3	GA 4.0	1.875×1.25 85 levels	Coupled	Fixed concentration	Bellouin et al. (2011) Walters et al. (2014)
MIROC- SPRINTARS	5.9.0	T85 40 levels	Coupled	HTAP2 emissions	Takemura et al. (2009) Takemura et al. (2005) Watanabe et al. (2010)
CESM-CAM4	1.0.3	2.5×1.9 26 levels	Slab	Fixed concentration	Neale et al. (2010) Gent et al. (2011)
CESM-CAM5	1.1.2	2.5×1.9 30 levels	Coupled	Emissions	Hurrell et al. (2013) Kay et al. (2015) Otto-Bliesner et al. (2016)
NorESM	1-M	2.5×1.9 26 levels	Coupled	Fixed concentration	Bentsen et al. (2013) Iversen et al. (2013) Kirkevåg et al. (2013)
IPSL-CM	5A	3.75×1.9 19 levels	Coupled	Fixed concentration	Dufresne et al. (2013)

Note: GA = Global Atmosphere. HTAP2 = Hemispheric Transport Air Pollution, Phase 2.

Table 2. Domain-averaged Tmax changes from each radiative component estimated from the linear models (unit: K).

CO_2×2				
Region	$SW_{clear\text{-}sky}$	SWCRE	↓LW	Total
N. America	-0.27 ± 0.01	0.95 ± 0.02	3.24 ± 0.03	3.92 ± 0.06
Europe	-0.24 ± 0.01	1.14 ± 0.03	2.79 ± 0.02	3.69 ± 0.06
E. China	-0.23 ± 0.01	0.71 ± 0.02	2.82 ± 0.02	3.30 ± 0.05
India	-0.29 ± 0.01	0.26 ± 0.01	2.59 ± 0.02	2.56 ± 0.04
BC×10				
Region	$SW_{clear\text{-}sky}$	SWCRE	↓LW	Total
N. America	-0.56 ± 0.03	1.00 ± 0.02	1.94 ± 0.04	2.38 ± 0.10
Europe	-0.73 ± 0.04	1.15 ± 0.03	1.32 ± 0.03	1.74 ± 0.10
E. China	-1.40 ± 0.08	0.98 ± 0.02	1.92 ± 0.04	1.50 ± 0.15
India	-0.89 ± 0.05	-1.05 ± 0.02	1.10 ± 0.02	-0.84 ± 0.05

Note: uncertainty range was estimated from the 95% confidence interval of each coefficient.

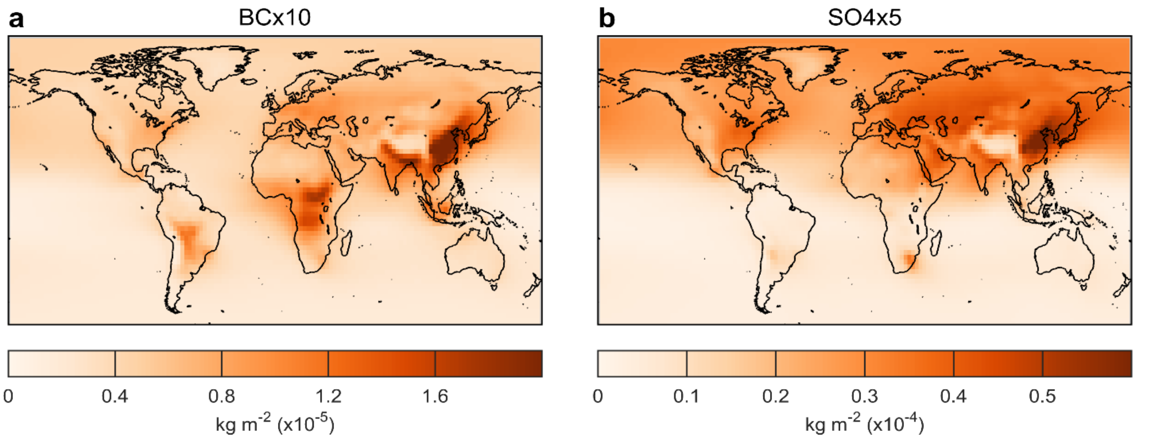

Figure 1: Aerosol loadings for the two aerosol experiments in CanESM2 model (as an illustrative example).

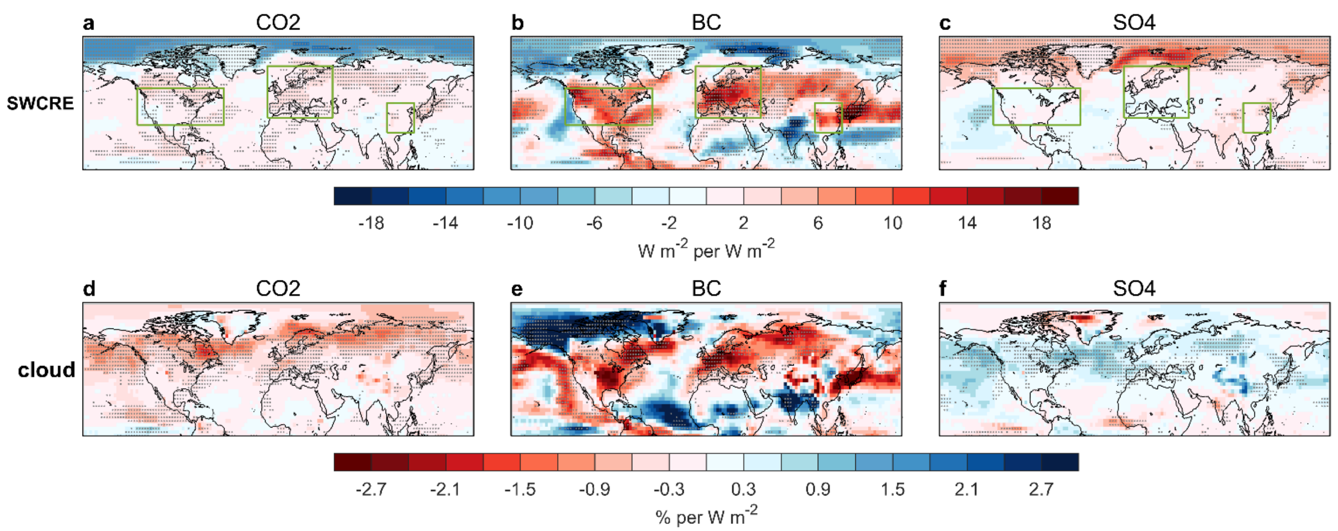

Figure 2: SWCRE changes (a-c) and cloud cover changes per unit forcing at 850 hPa (d-f) in JJA, results for SO_4 are changes per negative forcing. Grey dots indicate changes are significant at 0.05 level. Positive anomalies in a-c indicate more radiation reaching the surface.

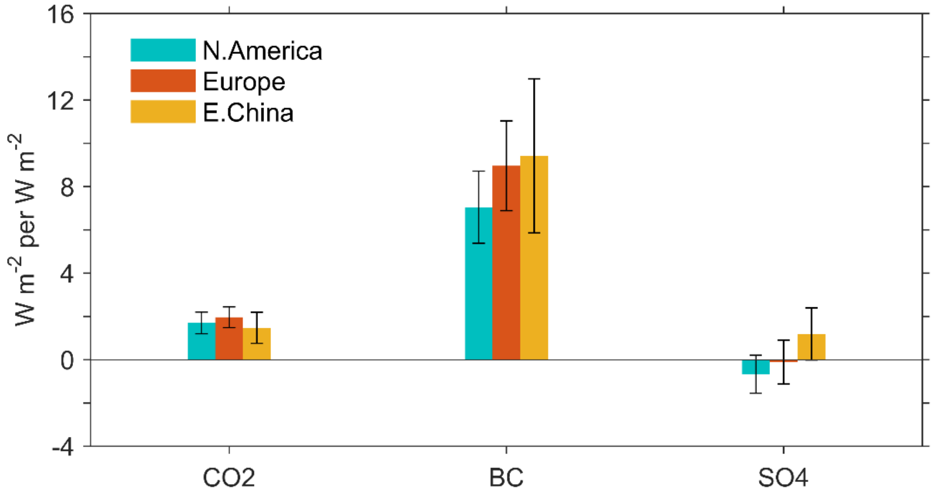

Figure 3: Domain-averaged SWCRE changes for three regions (green boxes in Fig. 2). Bars represent MMM results and errorbars indicate one standard error across the models.

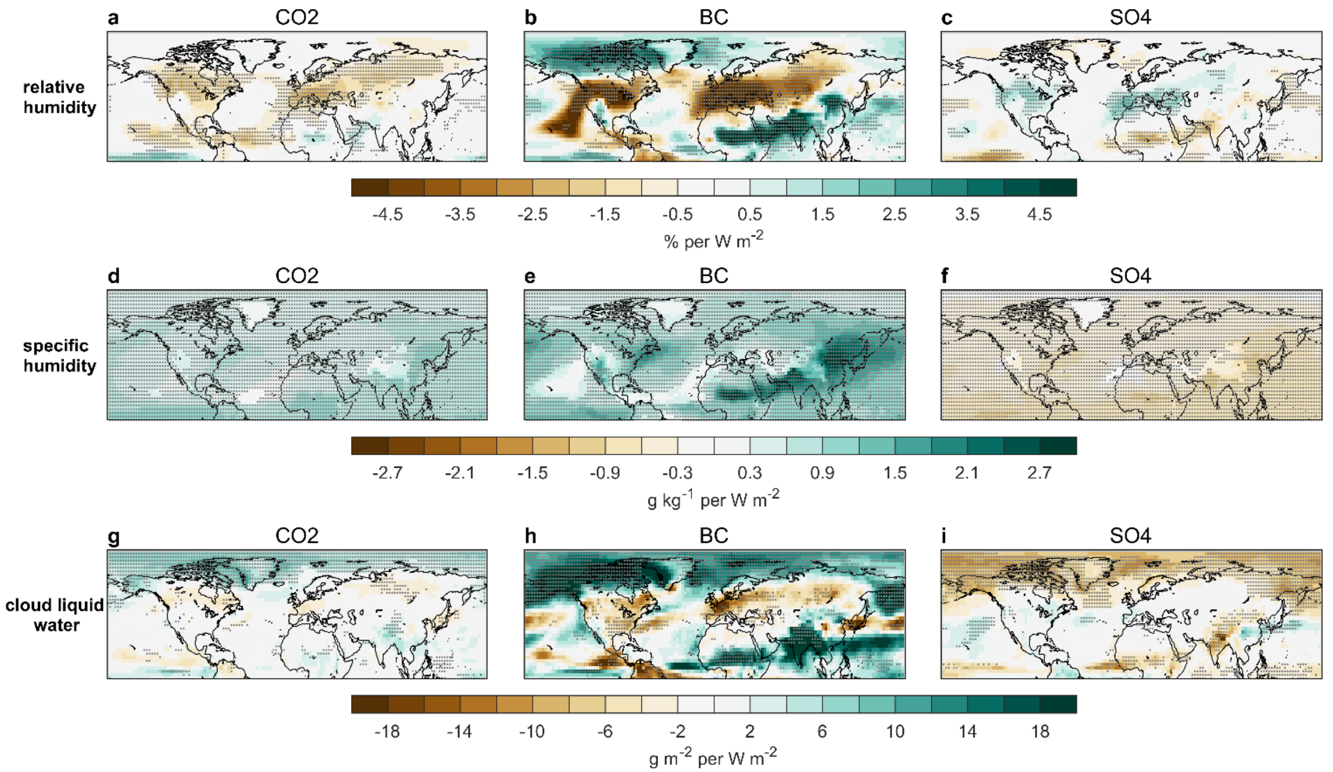

Figure 4: Same as Figure 2, but for humidity at 850 hPa.

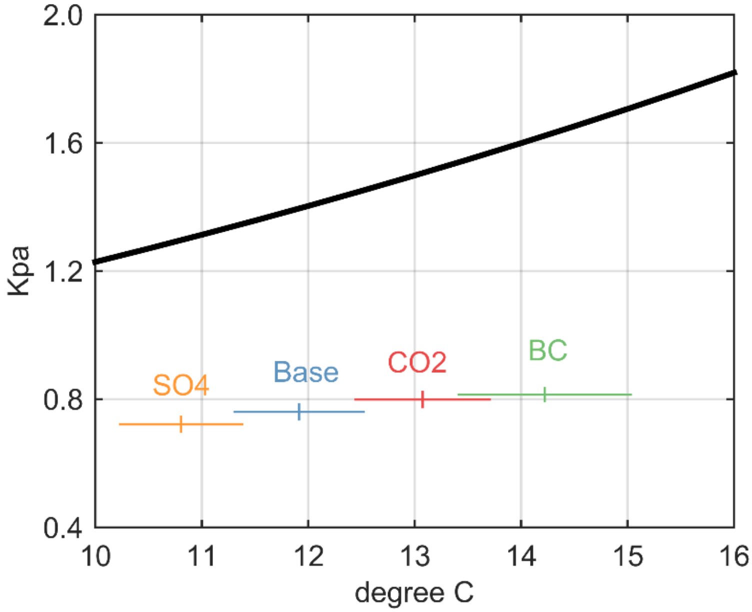

Figure 5: Domain-averaged vapor pressure changes per unit forcing as a function of temperature at 850 hPa for Europe. Errorbars indicate one standard error across the models. The thick black line represents saturation vapor pressure.

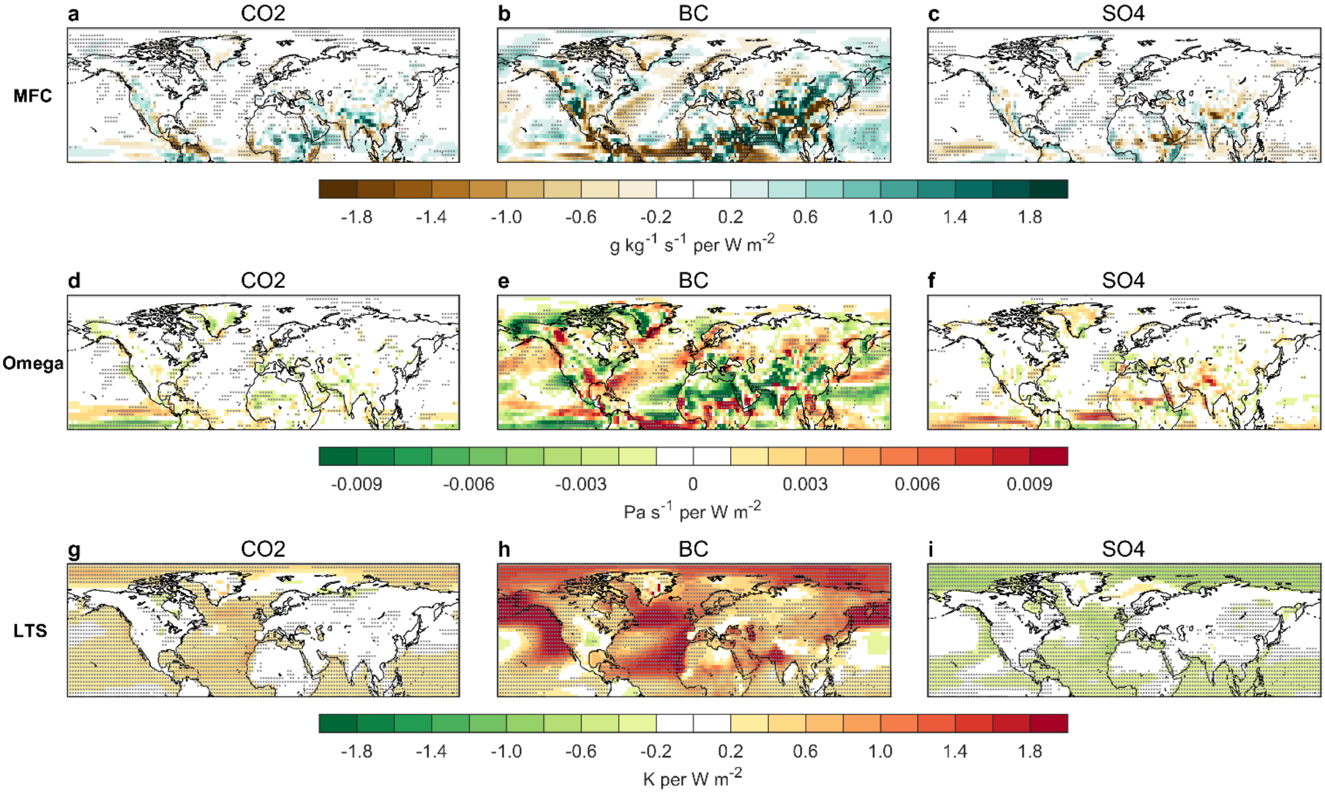

Figure 6: Same as Fig. 2, but for changes of moisture flux convergence (MFC, a-c), vertical velocity (omega, d-f) and lower tropospheric stability (LTS, g-i) per unit forcing. For vertical velocity (omega), positive anomalies indicate the air is less convective. LTS is calculated as the difference of potential temperature between 700 hPa and the surface. Positive LTS anomalies in g-i indicate stronger inversion or weaker lapse rate.

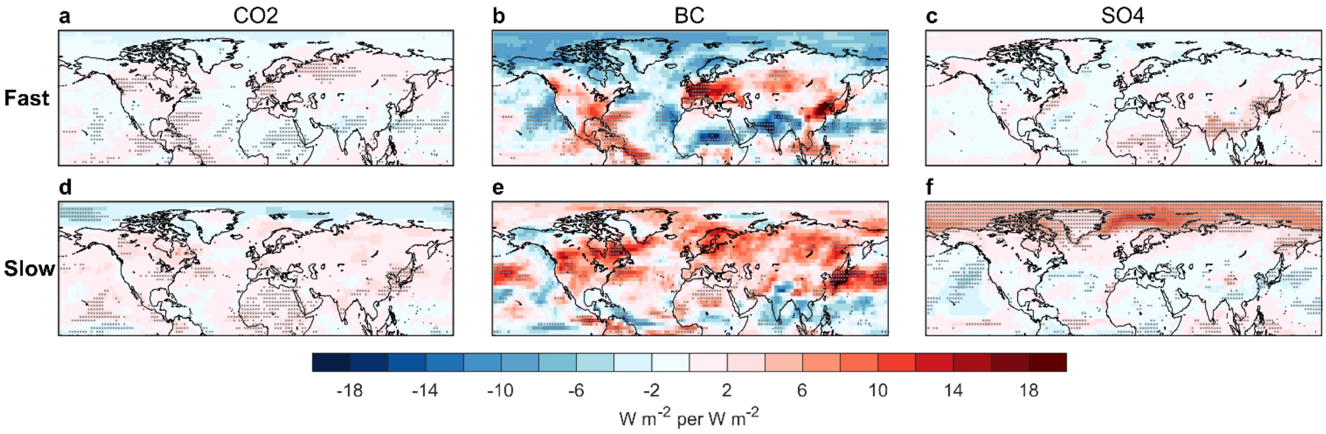

Figure 7: Same as Figure 2 (d-f), but for fast (a-c) and slow responses (d-f) of SWCRE changes per unit forcing.

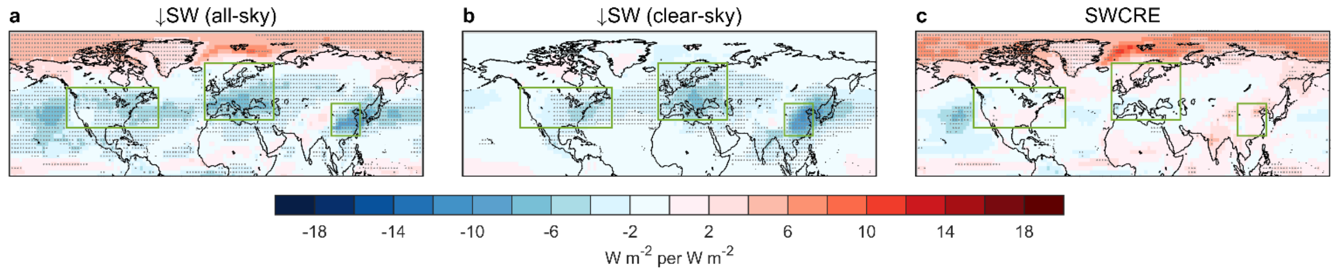

Figure 8: Changes of SW flux per unit negative forcing under all-sky (a), clear-sky (b) conditions and their difference (c) for the SO_4 experiment.

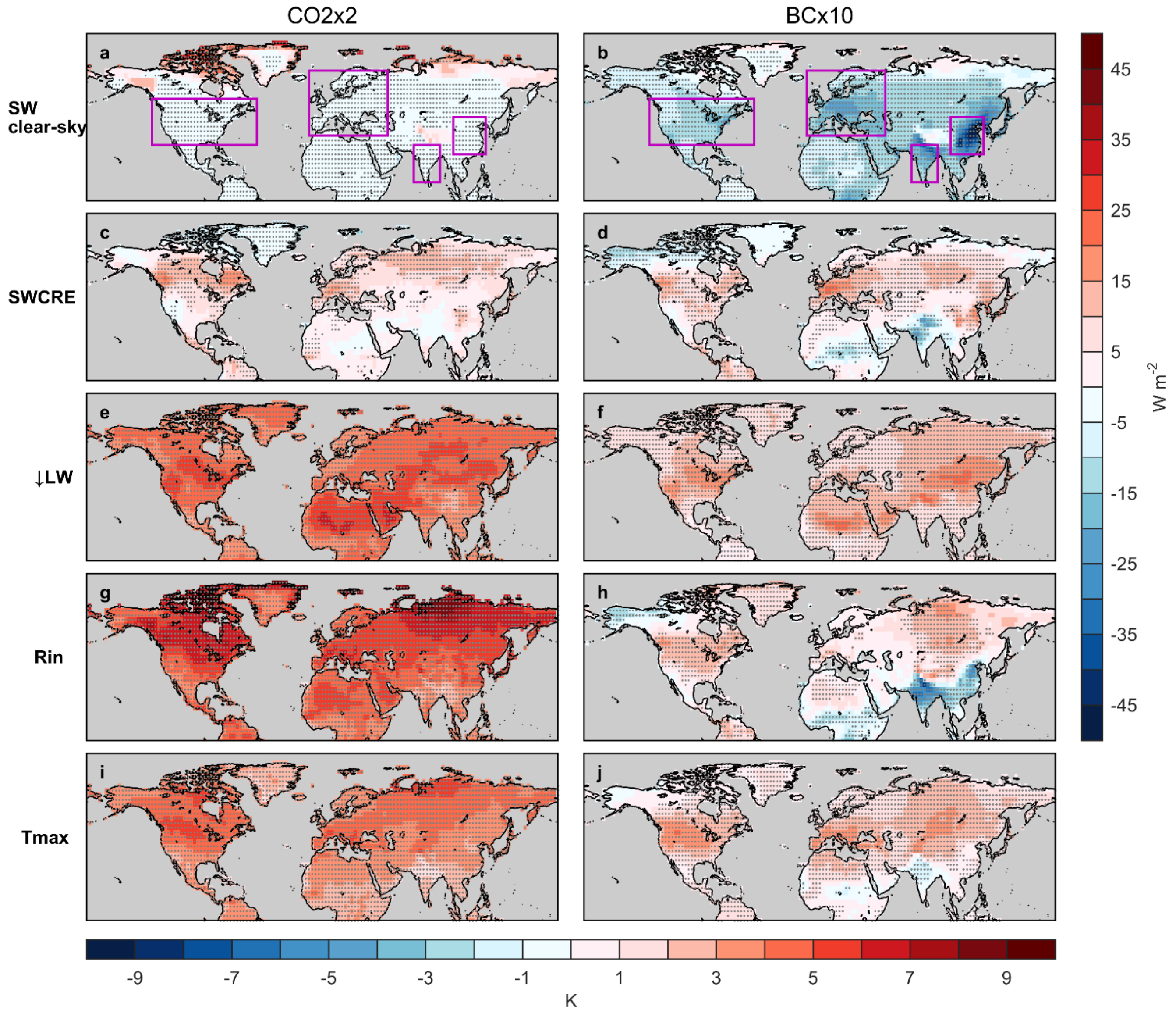

Figure 9: Changes of Rin and its components (a-h) as well as changes of Tmax (i-j) for the CO_2×2 (left) and BC×10 (right) experiments (original output, no normalization applied).

705

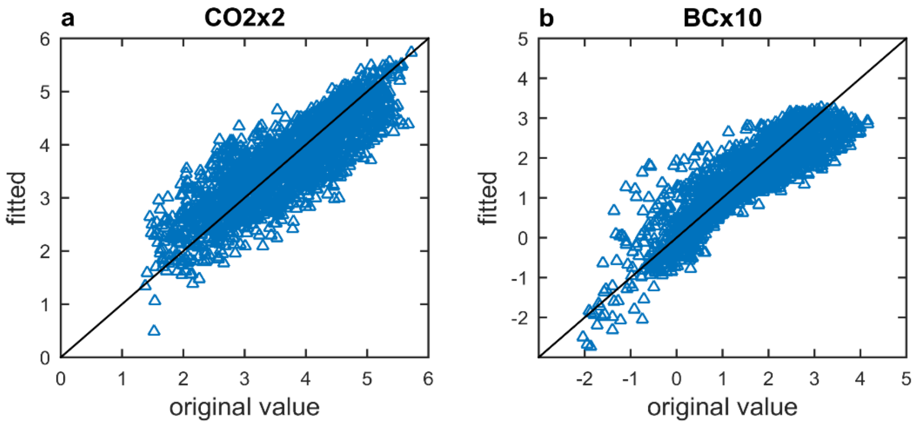

Figure 10: Comparison of fitted Tmax from the linear models vs original Tmax values. Blue triangles are values for all grid boxes over NH and black solid line represents one-one line.

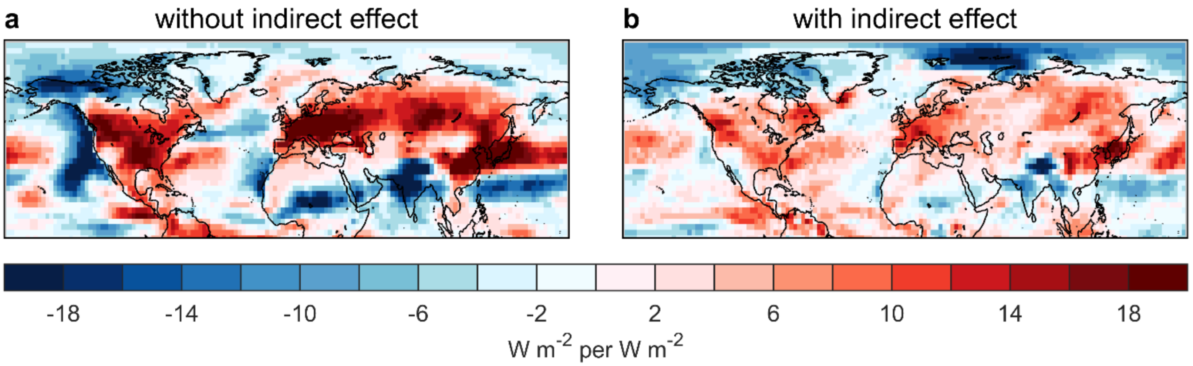

Figure 11: SWCRE changes for the BC experiment, (a) for models without aerosol indirect effects and (b) for models with indirect effects.

Response to comments #1

Response: Thanks for your helpful and constructive comments. We have made several modifications and implemented the suggestions as described below. We describe a few major changes first, followed by our response to individual comments.

i) Response of cloud liquid water is added.

ii) Response of lower tropospheric stability is added.

iii) SWCRE response for individual models is added to the supporting material.

iv) Replace Fig. 7 with SWCRE response.

This paper investigates the response of shortwave cloud radiative effect and daily maximum temperature to greenhouse gases and aerosols (BC and sulfate). It is found that BC results in a stronger positive SWCRE change than CO2 when normalized by effective radiative forcing, but sulfate does not have much effect on SWCRE. It is also shown that the increase in SWCRE resulting from CO2 and BC leads to an increase in daily maximum temperature during the summer. The results are interesting and have some important implications, however a number of things need to be addressed before recommendation for publication.

Major

1. Most of the results are normalized by effective radiative forcing. What are the surface temperature responses to CO2 and BC, respectively? Could the difference in SWCRE be partly due to the difference in the temperature change (i.e., the efficacy of BC)?
Response: the multi-model mean temperature changes for CO_2 and BC experiments are 2.5K and 0.7K respectively. The ratio of 2.5/0.7=3.6 is slightly larger than the ERF ratio (3.65/1.16=3.15), which means if SWCRE changes were normalized by dT, the difference of between CO_2 and BC would be slightly larger. As the results would not change much, however, the efficacy of BC will not significantly influence our results in the bar plot (Fig. 3).

2. The SWCRE change is attributed to the change in cloud cover. I would be interested to see some discussion in the change in cloud liquid water content or liquid water path, which also plays an important role in determining SWCRE.

Response: added in Fig. 4 and section 3.2. We added the following discussion after line 172:

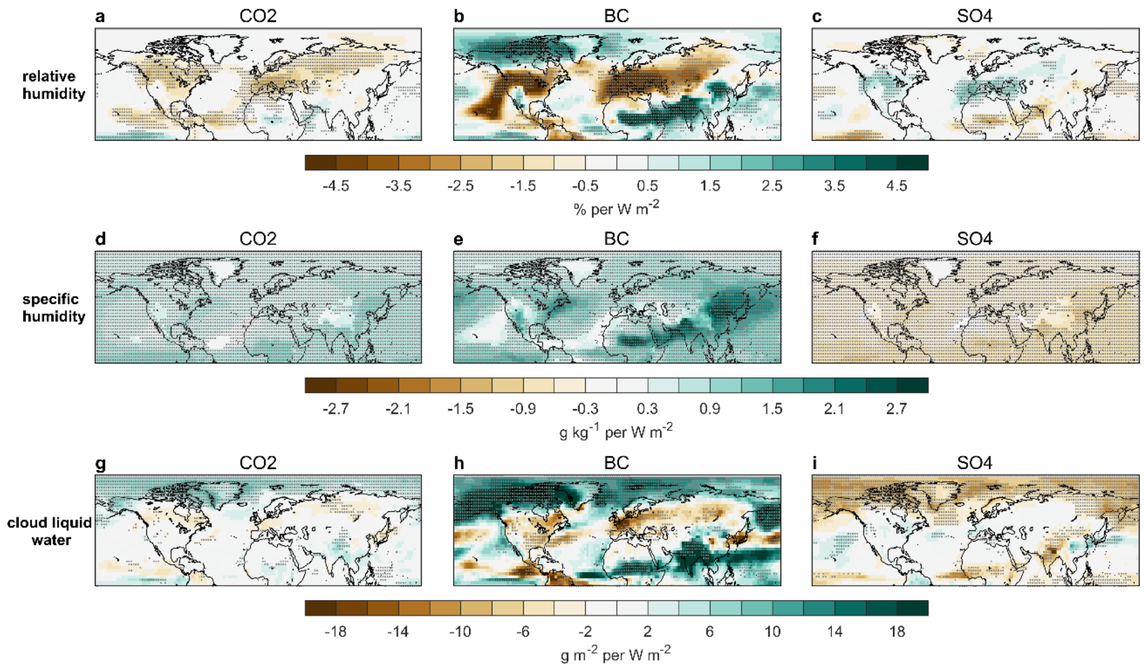

Figure 1: Same as Figure 2, but for humidity at 850 hPa.

"The response of cloud liquid water in the BC experiment could further support this conclusion (Fig. 4h). Liquid water decreases (increases) in regions with decreasing (increasing) cloud cover, following the pattern of RH. As cloud water content directly impacts cloud optical thickness and albedo, such a response may further impact SWCRE (i.e., enhance reflectance in regions showing increasing liquid water and enhance transmittance in regions with decreasing liquid water). However, the liquid water responses under CO2 and sulfate aerosols are much weaker, only significant in part of Asia and tropical Africa (Fig. 4g and i)."

3. The change in cloud cover is explained by the change in RH. However, there are a lot of other factors affecting clouds (radiation, dynamics, thermodynamics, etc., see Bretherton (2015) and references therein), and I think a more detailed discussion would be helpful. The authors look at vertical velocity and suggest that the change in stability plays less of a role, but it is not clear to me how the conclusion is reached. The estimated inversion strength or lower troposphere stability may be a better predictor for stability.

Response: accepted. We added lower troposphere stability in Fig. 6 and also kept the vertical velocity, as this is reported by some previous studies saying that subsidence could impact cloud cover (e.g., Myers and Norris, 2013). Thus, for the cloud cover changes, on top of humidity, we also discussed liquid water, moisture flux, dynamics and stability. Some discussions are also included and we also acknowledged that it is impossible to examine all the factors in the current study due to limited output.

We added the following discussion after line 192:
"Another mechanism that has been reported to influence cloud cover is LTS, in which a stable boundary layer could trap more moisture, thereby permitting more low-level clouds (Wood & Bretherton, 2006; Bretherton, 2015). In order to investigate this mechanism, we further analyzed LTS, defined as the difference of potential temperature between 700 hPa and surface (Fig. 6g-i), in which positive anomalies indicate a stronger inversion or weaker lapse rate. The LTS response is again strongest in response to BC forcing (Fig. 6h), with a widespread increase in stability. A previously reported positive correlation between LTS and low-level cloud cover is, nonetheless, only observed in BC source regions (tropical Africa and India) and part of the central US (Fig. 6h). The LTS responses over land are much weaker in response to CO2 and SO4 forcing, with some responses in Africa and India in response to sulfate aerosols (weaker inversion and less cloud). Some other factors have also been suggested to play a role in modifying low-level clouds, such as the diurnal cycle (Caldwell & Bretherton, 2009) and radiative effects of cirrus clouds (Christensen et al., 2013). Due to the limited model output, however, we acknowledge that it is impossible to examine these factors in the current study and it is beyond the scope of our study to probe all possible factors driving the cloud changes. In summary, the above analyses illustrate that the cloud cover changes we see can be primarily explained by RH changes and, to a lesser extent, changes of liquid water content, circulation, dynamics, and stability."

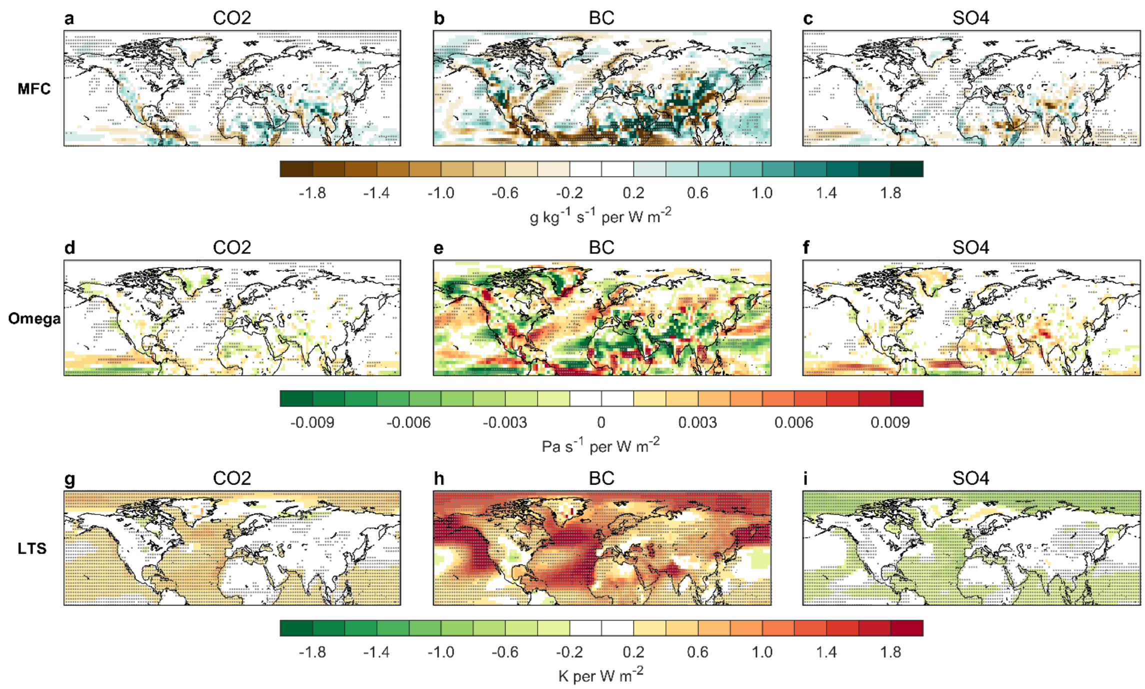

Figure 2: Same as Fig. 2, but for changes of moisture flux convergence (MFC, a-c), vertical velocity (omega, d-f) and lower tropospheric stability (LTS, g-i) per unit forcing. For vertical velocity (omega), positive anomalies indicate the air is less convective. LTS is calculated as the difference of potential temperature between 700 hPa and the surface. Positive LTS anomalies in g-i indicate stronger inversion or weaker lapse rate.

4. I have some conservation about including downward LW in the multilinear regression model. It is possible that downward LW change is a result rather than a cause of Tmax change (Tmax change results in changes in boundary layer temperature and moisture, and thus downward LW). In fact, consider the approximation LW~σT^4, dLW~$4\sigma T^3$dT, with T=300 K, dT/dLW~1/($4\sigma T^3$)~0.16, which is very close to the coefficients derived from the regression.

Response: thanks for your demonstration. For the multilinear linear regression, we still prefer to keep the LW component, as dT is directly related with incoming radiation. The aim of the linear regression is to attribute the contribution of those radiation components to Tmax changes and whether the radiative component is forcing or feedback is not important. In fact, the SWCRE we discussed in this study is mainly a feedback process, which is also included in the regression model. Your demonstration here further lends confidence to our regression results.

5. In PDRMIP BC and sulfate are increased by a factor of 10 and 5, respectively. It may be helpful to comment on whether the response is linear for such a large change. The small SWCRE response to sulfate is interesting and somewhat surprising. Given that aerosol direct effect is probably more linear than aerosol

indirect effect, would the authors expect different SWCRE response to historical change in sulfate?

Response: Previous studies show that most aspects of the climate change linearly with climate forcing, including BC. Thus, the large perturbation is unlikely to substantially impact the results and conclusions. We added a sentence in section 2.2 after line 116.

"Previous studies demonstrated that climate changes linearly with climate forcing for various forcing agents, including BC (Hansen et al., 2005; Mahajan et al., 2013)."

For SWCRE response to sulfate change, we do not have a definitive answer to the question of why these are so small or how linear they might be. Based on Fig. 8, the SWCRE change at the surface is not sensitive to cloud cover changes under sulfate aerosol forcing, as both aerosols and clouds scatter solar radiation. Thus, we cannot rule out the possibility that the historical changes are similar to our results.

Minor:

1. Please clarify that the paper analyzes SWCRE at the surface in the abstract, the main text, and the figures. It is somewhat confusing because I think SWCRE is more commonly referred to as TOA radiative forcing, and the first paragraph in the introduction describes SWCRE at the TOA.

Response: accepted and clarified where necessary.

2. Eq.(1): What is the time frequency of q and V for calculating the moisture flux?

Response: The time frequency of q and V is monthly, and we added this in the methods section, line 131:

"In Eq (1), q is specific humidity in g kg^{-1}, and V is horizontal wind including both zonal and meridional components. All variables have a monthly temporal resolution."

3. Figure 7: Maybe show the fast and slow responses of SWCRE instead of cloud cover, as the paper focuses on SWCRE.

Response: accepted and changed. Here is the new Fig. 7.

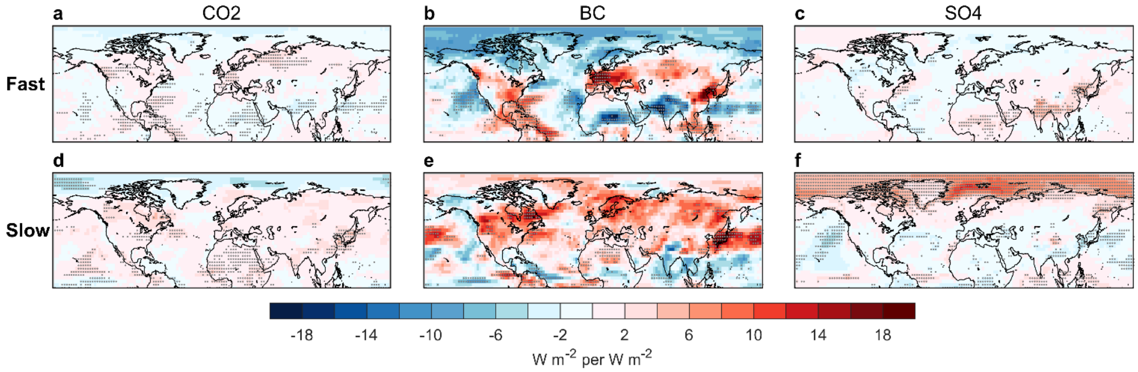

Figure 3: Same as Figure 2 (d-f), but for fast (a-c) and slow responses (d-f) of SWCRE changes per unit forcing.

Response to comments #2

Response: Thanks for your helpful comments. We have made several modifications and implemented the suggestions as described below. We describe a few major changes first, followed by our response to individual comments.

i) Response of cloud liquid water is added.

ii) Response of lower tropospheric stability is added.

iii) SWCRE response for individual models is added to the supporting material.

iv) Replace Fig. 7 with SWCRE response.

This paper is interested in GCM-produced summertime changes in the maximum land temperatures of the NH under perturbed conditions, namely doubled CO2, 10 times more black carbon aerosols and 5 times more sulphate aerosols (subject to model interpretation). Results come from a somewhat outdated database (CMIP5 era models) and the focus is on the SW effects of clouds at the surface (at least initially, later when a prediction model is built LW is added too). I'm not clear what we learn from the analysis. The general consensus since AR5 has been that low clouds provide a positive feedback under CO2 doubling (or quadrupling for that matter), so SWCRE at the surface is expected to be weaker (less cooling at the surface). So, this part is not so new, although I guess one can focus on the effect of this reduced radiative cooling on Tmax. Then there is the aerosol: aerosol changes can change the environment, the circulation, etc, so they can change cloudiness. But they can impact the clouds "faster" through alteration in microphysics (lifetime, optical thickness changes) and this part is not discussed until the conclusions. In any case, the effect of aerosol on (low?) clouds and therefore on land Tmax is not clear-cut since there is also the direct radiative dimming or brightening part that works in conjunction or competition with the cloud effect. So, it's kind of interesting to see results about this, although I imagine people have previously looked at that too. I guess the most intriguing result is that Tmax changes can largely predicted by LOCAL RADIATIVE changes; it was somewhat unexpected to me that this works as well as it does since temperature is also affected by turbulent fluxes and advection (non-local effects). I suggest the authors make a bigger deal of this finding.
Response: We really appreciate the reviewer for speaking so highly for our research. To our best knowledge, CMIP6 models do not have such multi-model inter-comparison project that investigates the climate response to individual forcing

agents yet. Thus, PDRMIP is still the only multi-model project for understanding climate response to individual climate forcings. Our main focus is SW, and LW is included later is because it also impacts surface Tmax. The linear regression aims to quantify the contribution of each radiative component to Tmax. Compared with previous cloud feedback studies, our study has contributions in the following perspectives: 1) better understanding of cloud feedback to individual forcing agents (e.g., stronger response to BC than to GHGs); 2) better understanding on surface SWCRE instead of TOA (e.g., surface SWCRE under sulfate aerosol is much weaker than TOA); 3) quantify their contributions to Tmax, as many previous studies reported this cloud impact on heatwave and drought events, but none of them quantified such impact. We added cloud liquid water analysis in the revised version. For aerosol indirect effect, it is a limitation in PDRMIP study, as the concentrations needs to be fixed. A sentence is added in data section to inform the readers that most of the model have direct effect only in line 104:

"It is noted that only three of the nine models include aerosol-cloud interactions while the remaining ones only have aerosol-radiation interactions. However, this does not impact our main conclusions (see section 4)."

The radiative dimming/brightening effect is already included in $SW_{clear\text{-}sky}$ component, whose cooling effect is outweighed by warming effect (Table 2). Fig. 10 and R value indicate that the linear fit works fairly well. However, it is not expected to explain 100% of Tmax changes. Other factors may also play a role, which has been acknowledged in the manuscript, such as line 277:

"It is noted that the radiation change might not explain all Tmax changes, as other factors may come into play. For instance, the temperature response would be different when surface is getting drier under a warmer climate. This is because more net radiation is realized as sensible heat instead of latent heat under drier conditions, which has been suggested to play an important role in recent European heatwaves (Seneviratne et al., 2006; Fischer et al., 2007)."

Here are my main issues with this paper:

1. LW and indirect cloud effects are not discussed until the concluding section.

Response: the main focus of our study is SWCRE. LW impact is small and is not our focus. Thus, we put LW in the discussion section. For aerosol-cloud interactions, it is one of the limitations in the PDRMIP models, so we acknowledge this together with other limitations in the end. However, we added a sentence in the section 2.1

to inform the readers that most of the models only have aerosol-radiation interactions in line 104:

"It is noted that only three of the nine models include aerosol-cloud interactions while the remaining ones only have aerosol-radiation interactions. However, this does not impact our main conclusions (see section 4)."

2. Since the SWCRE effects are mostly attributed to CF changes, the LW downwelling to surface changes should also be broken to clear and LWCRE effects; I mean basically the LW should be treated as the SW and not lumped into a single term in the regression.

Response: the main focus of our study is SWCRE, because SW dominates in the cloud radiative effects. Many published studies (e.g., those in the introduction section) reported that SWCRE plays an important role in amplifying heatwave and drought events world-wide, as cloud cover reduction directly enhance solar heating, thereby raising Tmax. LW effect in these processes is very limited. Our study follows these studies and extends the investigation into the contribution of SWCRE to Tmax. As LWCRE effect is small, we prefer to keep the current regression results.

3. Why are only CF changes considered and not changes in other cloud properties? Optical thickness changes can have impact in SWCRE.

Response: We focus on CF changes because CF could largely explain the SWCRE changes. PDRMIP does not provide output on cloud optical thickness. However, we added cloud liquid water analysis in the revised version (Fig. 4), which is directly related with optical thickness and impact SWCRE (please see section 3.2 in the revised version) in line 173:

"The response of cloud liquid water in the BC experiment could further support this conclusion (Fig. 4h). Liquid water decreases (increases) in regions with decreasing (increasing) cloud cover, following the pattern of RH. As cloud water content directly impacts cloud optical thickness and albedo, such a response may further impact SWCRE (i.e., enhance reflectance in regions showing increasing liquid water and enhance transmittance in regions with decreasing liquid water). However, the liquid water responses under CO_2 and sulfate aerosols are much weaker, only significant in part of Asia and tropical Africa (Fig. 4g and i)."

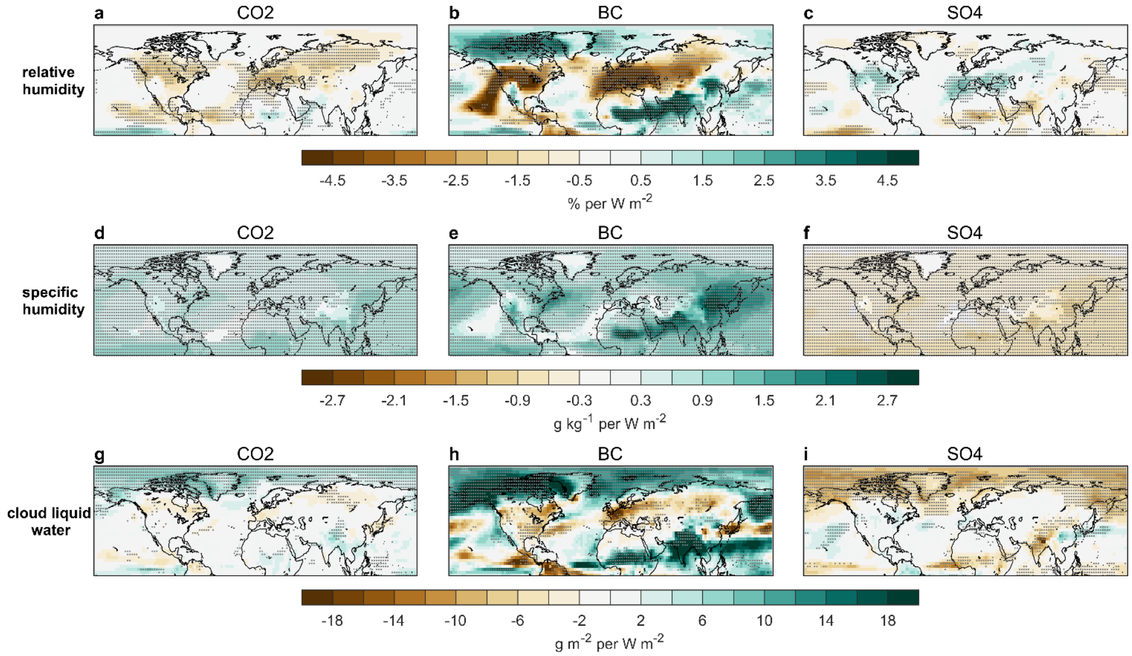

Figure 1: Same as Figure 2, but for relative humidity (a-c), specific humidity (d-f), and cloud liquid water (g-i) at 850 hPa.

4. Only CF changes for low clouds (and the corresponding RH) are considered (if I understand correctly), but for SW cloud at any altitude in the in the atmospheric column can have strong SWCRE effects

Response: We consider low-level clouds because low-level clouds dominate SW changes. We also analyzed the cloud cover changes in 500 hPa and 300 hPa, which show similar changes to those at 850 hPa. We mentioned this in section 3.1 and the figure is included in supporting material (Fig. S5).

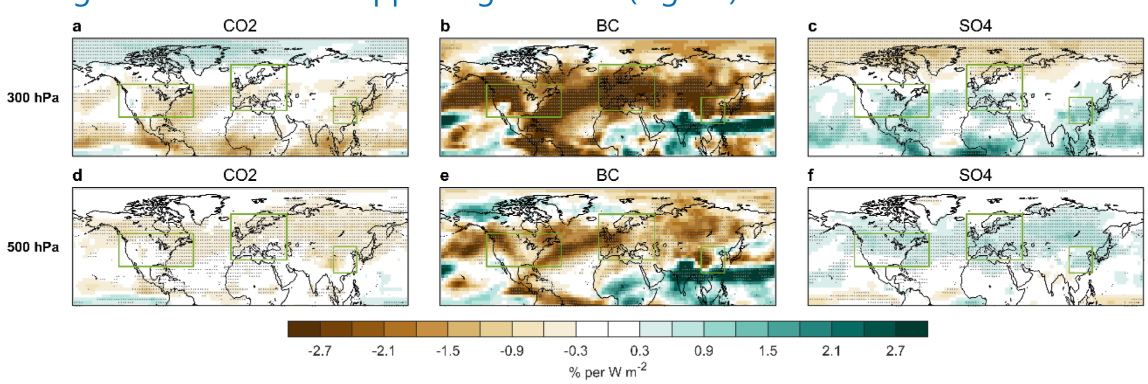

Figure S2: same as Fig. 2(d-f) in the main text, but for cloud cover changes at 300 hPa and 500 hPa.

Some minor issues:

1. Clarify from the start that SWCRE refers to surface.

Response: accepted and clarified where necessary.

2. Define changes in SWCRE more formally. For this SWCRE itself has to be defined more formally, i.e., difference between net all-sky and clear-sky fluxes where net = down-up flux. Then you have to take a difference between baseline and perturbed conditions. Just saying that a positive SWCRE change means less cooling is unsatisfying.

Response: accepted and clarified in section 2.2.

"In this study, we focus on the SWCRE at the surface in the low and mid-latitudes during boreal summer months (June-July-August, JJA hereafter), which is calculated as the difference in the SW radiative flux at the surface between all-sky and clear-sky conditions (Ramanathan et al., 1989)……. Changes in SWCRE are obtained by subtracting the control simulations from the perturbations using the data of the last 20 years in each coupled simulation."

3. I find the discussion between fast and slow feedbacks a bit superficial. Land responds to fast feedbacks, but for slow feedbacks the SST responds as well and that's what will drive circulation changes. For slow feedbacks it makes more sense to look at TOA quantities. When it comes to direct radiative effect of aerosol, TOA and SFC changes are distinct for absorbing (BC) vs non-absorbing aerosols (sulphate).

Response: we agree that the slow response is controlled by global mean temperature change (including SST). However, the aim of this part is to give a qualitative picture that the cloud response is mainly due to fast response, slow response or both. What specific process that drives these slow responses is not our focus. Following another reviewer's comment, we replaced the Fig. 7 of fast and slow cloud cover response with SWCRE changes at the surface.

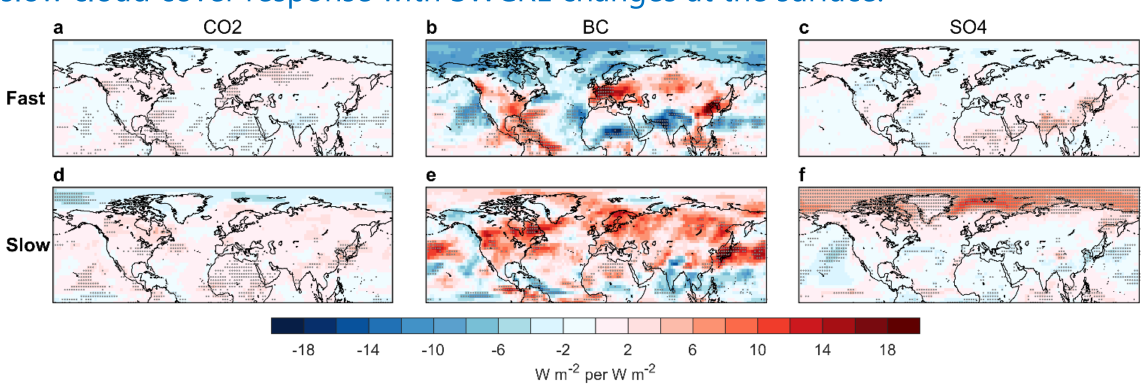

Figure 3: Same as Figure 2 (d-f), but for fast (a-c) and slow responses (d-f) of SWCRE changes per unit forcing.

4. Why not use the same colorbar in Fig. 2 for normalized forcing change and cloud fraction change to make comparison easier (of course range of values can be different)?

Response: changed to same colorbar.

5. By showing only MMM results and nothing about model spread we have no idea how much the models diverge in predictions. Not sure there is an easy way to convey that.

Response: As these are spatial maps, we could not figure out a way of showing inter-model spread at this moment. So we just follow the traditional way by showing MMM results. In fact, the uncertainty bars in the bar plot (Fig. 3) could shed some light on the inter-model spread of the results. For CO2 and BC, the results are quite consistent across the models and for SO4, even the sign of change is uncertain and thus, a larger range is seen. These results are further illustrated by the individual model response, which has been included in the supporting material (Fig. S2-S4).

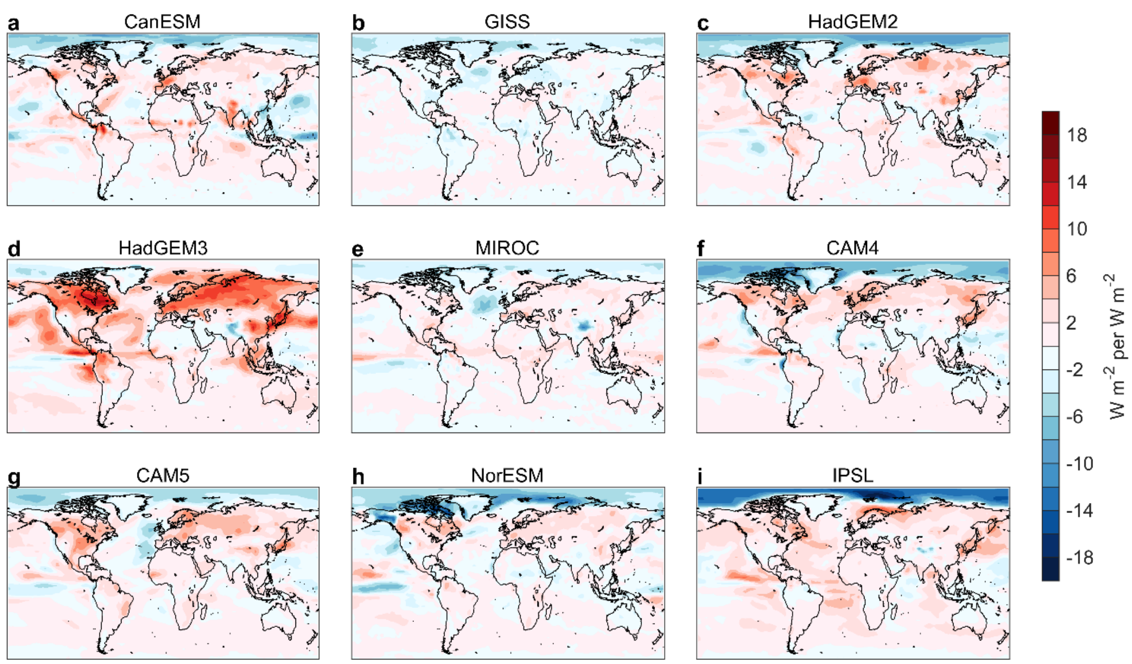

Figure S4: SWCRE changes per unit forcing by individual models for the CO_2 experiment.

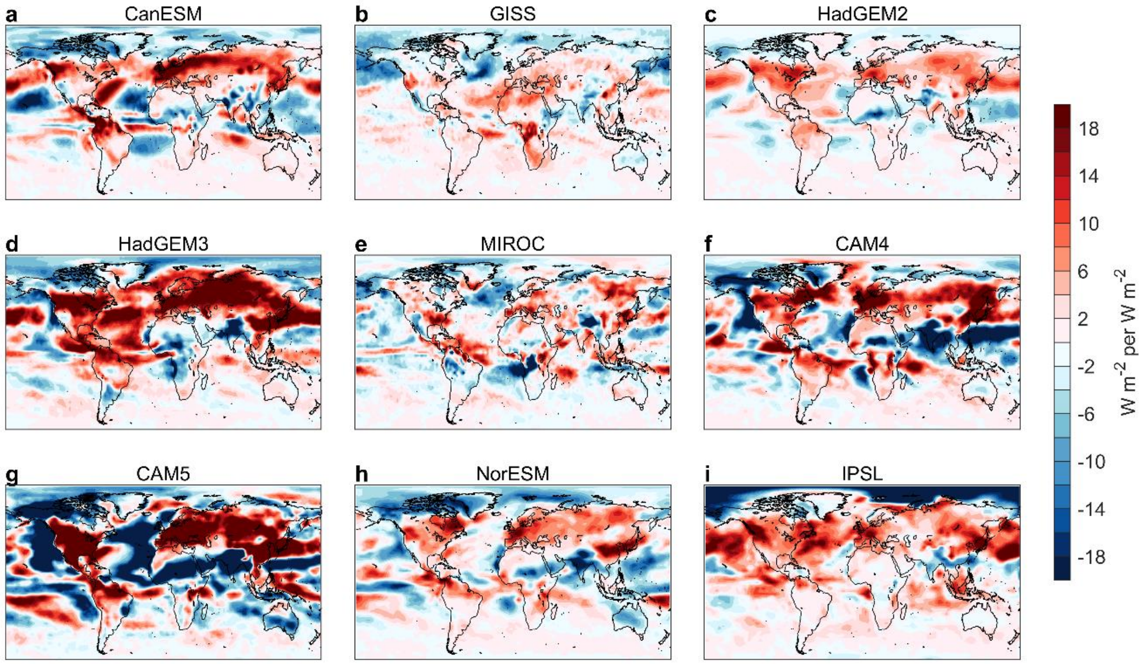

Figure S5: SWCRE changes per unit forcing by individual models for the BC experiment.

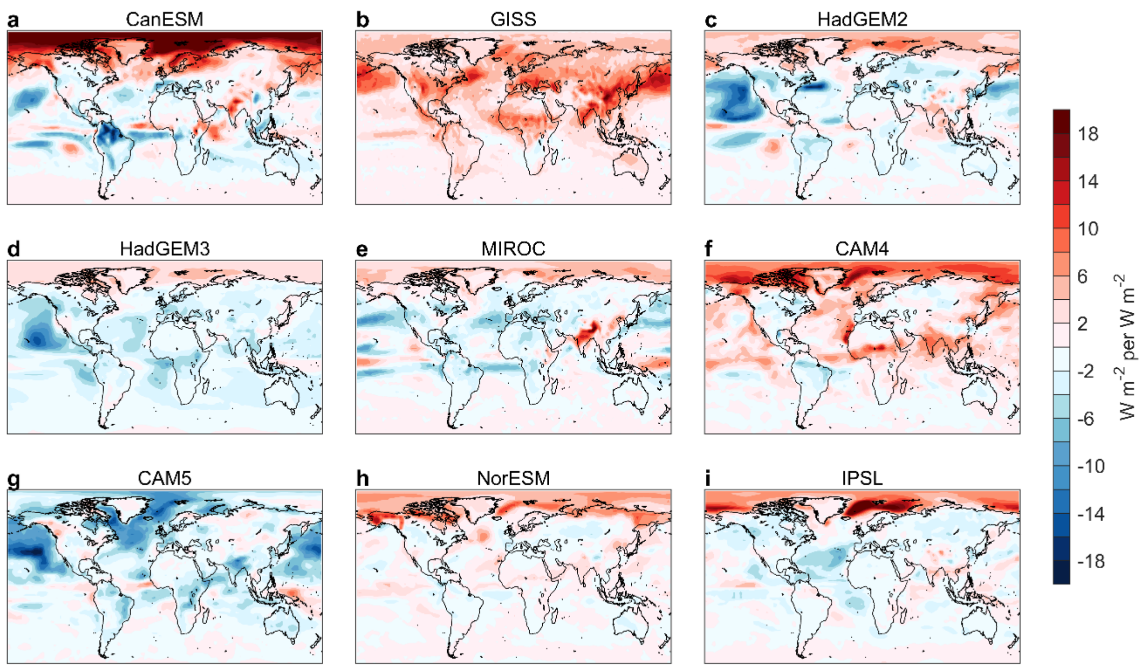

Figure S6: SWCRE changes per negative forcing for the sulfate aerosol experiment.

6. I imagine the radiative treatment of aerosol differs widely among models. Not discussed. When you change emissions instead of concentrations directly, divergence is introduced too.

Response: the readers could refer to the literature documenting each model in Table 1 for detailed radiative treatment of aerosols, as it is nearly impossible to discuss them one by one. We added the SWCRE changes for individual models into supporting material (Fig. S2-S4; see the response above). The main features are consistent across models and not sensitive to model setup (e.g., emission, concentration or radiative treatment), indicating that our results are fairly robust. We added these in section 3.1 line 144:

"When it comes to individual model response (Fig. S2-S3), these patterns are also consistent across at least eight of the nine models and are not very sensitive to the model setup (emission-based or concentration-based)."

7. I also imagine that the base state of the models is quite different too. Care to comment?

Response: The multi-model mean value of SWCRE in the base run is -57.9±1.8 W m^{-2} (MMM±1 standard error). The spatial patterns are fairly consistent across the models, with strong SWCRE in tropical regions and mid-to-high latitudes and weaker SWCRE in subtropics, regions generally with less clouds. We added this figure in the supporting material (Fig. S1) and also mentioned this in section 2.2 in line 111:

"The base state of SWCRE in each model is shown in Fig. S1, with a multi-model mean (MMM) value of -57.9±1.8 W m^{-2} (MMM±1 standard error). The spatial patterns are fairly consistent across the models, with strong SWCRE in tropical regions and mid-to-high latitudes and weaker SWCRE in subtropics, regions generally with less clouds."

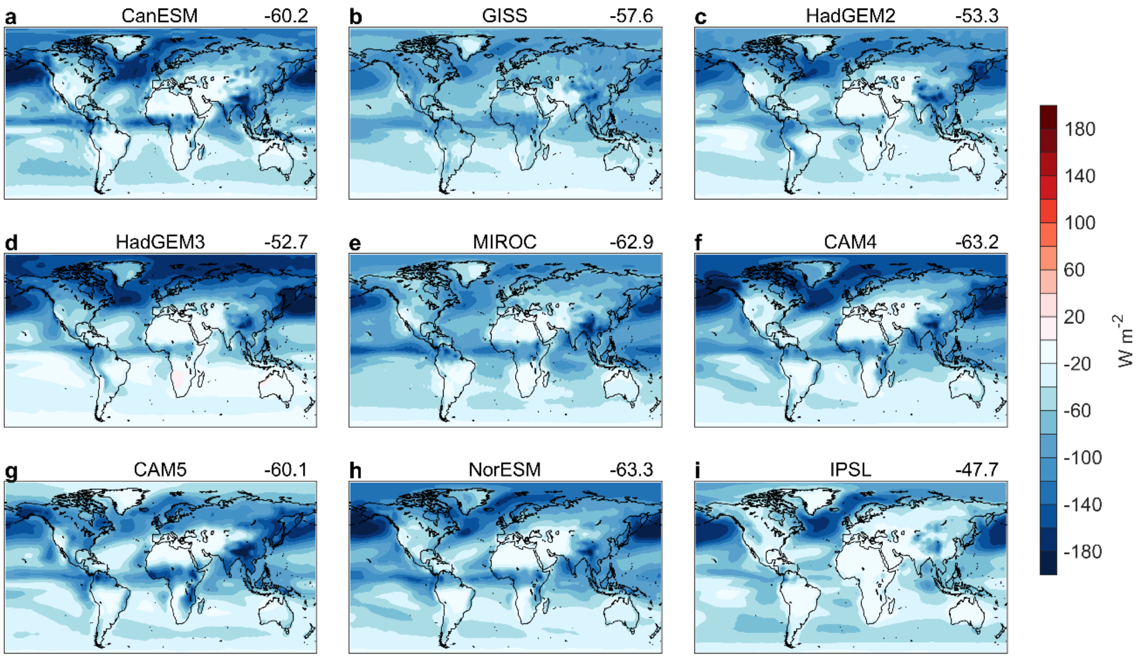

Figure S7: SWCRE in the base climate for each model. The global mean values are shown in the upper-right corner.