Impacts of Wildfire Aerosols on Global Energy Budget and Climate: The Role of Climate Feedbacks

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ABSTRACT

Aerosols emitted from wildfires could significantly affect global climate through perturbing global radiation balance. In this study, the Community Earth System Model with prescribed daily fire aerosol emissions is used to investigate fire aerosols' impacts on global climate with emphasis on the role of climate feedbacks. The total global fire aerosol radiative effect (RE) is estimated to be $-0.78 \pm 0.29 \text{ W m}^{-2}$, which is mostly from shortwave RE due to aerosol–cloud interactions (RE_{aci}; $-0.70 \pm 0.20 \text{ W m}^{-2}$). The associated global annual-mean surface air temperature (SAT) change ΔT is $-0.64 \pm 0.16 \text{ K}$ with the largest reduction in the Arctic regions where the shortwave RE_{aci} is strong. Associated with the cooling, the Arctic sea ice is increased, which acts to reamplify the Arctic cooling through a positive ice-albedo feedback. The fast response (irrelevant to ΔT) tends to decrease surface latent heat flux into atmosphere in the tropics to balance strong atmospheric fire black carbon absorption, which reduces the precipitation in the tropical land regions (southern Africa and South America). The climate feedback processes (associated with ΔT) lead to a significant surface latent heat flux reduction over global ocean areas, which could explain most ($\sim 80\%$) of the global precipitation reduction. The precipitation significantly decreases in deep tropical regions (5°N) but increases in the Southern Hemisphere tropical ocean, which is associated with the southward shift of the intertropical convergence zone and the weakening of Southern Hemisphere Hadley cell. Such changes could partly compensate the interhemispheric temperature asymmetry induced by boreal forest fire aerosol indirect effects, through intensifying the cross-equator atmospheric heat transport.

1. Introduction

Fire is an integral component of the Earth system, and it is driven by climate, ecosystems, and human activities (Li et al. 2017). Fire imposes great impacts on global climate through effects on radiative forcing, terrestrial ecosystems, and biogeochemical cycling (Li and Lawrence 2017). The feedback mechanisms between fire and climate interactions are still a great challenge and remain fundamentally uncertain (Liu et al. 2014; Hantson et al. 2016).

The impacts of fire-emitted aerosols on climate have received more attention recently. The fire aerosols' radiative effect (RE) and radiative forcing (RF) are estimated to quantify its impacts. RE represents the instantaneous radiative impact of atmospheric particles on Earth's energy balance (Heald et al. 2014), and RF is calculated as the change of RE between two different periods (e.g., preindustrial and present-day). The fire aerosols' radiative effects/forcings could be due to aerosol–radiation interaction (ARI), aerosol–cloud interaction (ACI), and surface albedo change (SAC). All of these effects are intensively estimated recently.

The present-day fire radiative effect due to ARI (RE_{ari}) is a small and positive value (warming), ranging from 0.13 to 0.18 W m^{-2} , based on several recent

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studies with the state-of-the-art global climate models (Ward et al. 2012; Tosca et al. 2013; Jiang et al. 2016). The fire REari is mostly due to fire black carbon (BC) absorption, while fire particulate organic matter (POM) induces a small overall effect (Jiang et al. 2016). The present-day fire radiative forcing due to ARI (compared to preindustrial time) was estimated in the IPCC Fifth Assessment Report (AR5), and the estimated value is near zero (0.0 W m^{-2} ; ranging from -0.20 to 0.20 W m^{-2}).

The fire aerosols' radiative effect due to ACI (REaci) could be negative and larger than REari (Grandey et al. 2016). With the climate model only considering cloud albedo effect, the present-day fire aerosols' REaci was reported to be -1.16 W m^{-2} in Chuang et al. (2002). With different fire emissions, the REaci of biomass burning aerosols was estimated to range from -1.64 to -1.00 W m^{-2} for the year 2000 by global climate models (Ward et al. 2012). With daily fire emissions, Jiang et al. (2016) reported that the global fire aerosol shortwave REaci is $-0.70 \pm 0.05 \text{ W m}^{-2}$, resulting mainly from the fire POM effect ($-0.59 \pm 0.03 \text{ W m}^{-2}$).

The fire-emitted BC could deposit on snow and change the surface albedo (Flanner et al. 2007; Jiang et al. 2017). The fire RE due to SAC (REsac) tends to induce a warming, but its global-mean effect could be very small. Bond et al. (2013) made a summary of BC-in-snow forcing/effect and found that the present-day fire BC-in-snow effect ranges from 0.006 to 0.02 W m^{-2} . Jiang et al. (2016) estimated REsac of fire emitted BC with a new climate model and found that its value is about 0.02 W m^{-2} which is generally consistent with previous studies. The global fire aerosols' radiative effect is in general negative (cooling) and mostly determined by REaci (Ward et al. 2012; Jiang et al. 2016; Landry et al. 2017; Grandey et al. 2016).

More evidence has shown that fire radiative effect has great impacts on regional cloud and precipitation. In the tropical regions, fire aerosols generally tend to reduce the warm cloud and suppress the precipitation. Based on observational data, Tosca et al. (2014) found that fire aerosols tend to reduce the cloud fraction in sub-Saharan Africa. Similarly, Koren et al. (2004) also reported a correlation between thick smoke and decreased cloud cover over the Amazon. Based on modeling studies, Zhang et al. (2009) reported that fire aerosols may warm and stabilize the lower troposphere and reinforce the dry-season rainfall pattern in southern Amazonia. In contrast, fire aerosols could invigorate the tropical convective clouds through suppressing warm rain processes in the convection, and enhance the latent heat release at higher levels (Andreae et al. 2004).

Fire aerosols' effects on cloud and precipitation in the extratropical regions are also investigated. With a regional model, Lu and Sokolik (2013) found that fire aerosols lead to an increase of cloud water path during Canadian boreal wildfires in the summer of 2007. Liu et al. (2018) found that fire aerosols have significant impacts on both liquid and ice clouds over southern Mexico and the central United States during the fire events in April 2009.

The global climate response to fire aerosols' radiative effect is rarely investigated. Tosca et al. (2013) found that fire aerosols' radiative effect reduces the global surface temperature by 0.13 K and has very small impacts on the global-mean precipitation. They also reported a significant precipitation decrease near the equator and a weakening of the Hadley cell due to fire aerosols. However, only aerosols' direct and semidirect effects are considered in their study, without inclusion of any indirect effect. This study aims at investigating the global energy budgets and climate change induced by fire aerosols with a state-of-the-art climate model considering all of the fire radiative effects. Considering that the global response to fire radiative effect could be from both rapid adjustments to the fire radiative effect (fast response) and slow climate feedback process (slow response) (Gregory and Webb 2008), this study will discuss the relative role of two kinds of responses to fire aerosols. We emphasize the role of boreal forest fire aerosol indirect effect and associated climate feedbacks, which is not revealed in previous studies. The rest of the paper is organized as follows. Section 2 describes the model and the experiments used. Section 3 introduces the methodology of diagnostic aerosol climate feedbacks. Section 4 presents the main results of this study. The final section is devoted to conclusions and discussion.

2. Model and experimental design

a. Model

The Community Earth System Model (CESM), version 1.2, developed by the National Center for Atmospheric Research (NCAR) is used for this study (Neale et al. 2012). Its atmospheric component, Community Atmosphere Model, version 5.3 (CAM5.3), has several major improvements in its physics parameterizations compared to previous CAM versions. CAM5.3 could predict both mass and number mixing ratios of cloud liquid and cloud ice, with a two-moment stratiform cloud microphysics scheme (Morrison and Gettelman 2008). A new four-mode version of the Modal Aerosol Module (MAM4) (Liu et al. 2016) developed from Liu et al. (2012) was

TABLE 1. Numerical experiments and associated fire aerosol emissions in each experiment.

Experiment	Years (used)	Fire emission	SST and sea ice
FIRE_AMIP	5×10 (5×9)	On	Prescribed
NOFIRE_AMIP	5×10 (5×9)	Off	Prescribed
FIRE_CMIP	99 (45)	On	Online
NOFIRE_CMIP	99 (45)	Off	Online

implemented in CAM5, which significantly increases the BC and POM concentrations in the remote regions (e.g., over oceans and the Arctic).

b. Experimental design

CESM was run in a resolution of 0.9° latitude $\times$ 1.25° longitude and 30 vertical levels with the finite-volume dynamics core. Global Fire Emissions Database version 3.1 (GFED 3.1) daily emissions (Giglio et al. 2013) for BC, POM, and sulfur dioxide (SO_2) from 2003 to 2011 (9 years) are used for the fire (FIRE_AMIP and FIRE_CMIP) cases, and the vertical distribution of fire emissions is based on the AeroCom protocol (Dentener et al. 2006). Biomass burning aerosols from different type landscape areas (forest, shrubland, savanna, grassland, cropland, and other) are considered (van der Werf et al. 2017). Anthropogenic aerosol and precursor gas emissions are from the IPCC AR5 dataset (Lamarque et al. 2010). Both the fixed-SST (AMIP type) and coupled (CMIP type) simulations were run in this study to diagnose the radiative effect and climate feedback of fire aerosols. The CMIP-type simulations were run with the single-layer ocean model (SOM) (Neale et al. 2012), which allows the SST's adjustments to the aerosol forcing.

Table 1 summarizes the information of the four simulations used in this study. The FIRE_CMIP simulation was performed with 11 cycles (99 years) of the 2003–11 fire emissions described above. Thus, the FIRE_CMIP run includes observed year-to-year variability in emissions during each cycle, similar to Tosca et al. (2013). The NOFIRE_CMIP simulation was identical to the FIRE_CMIP simulation, but without inclusion of fire emissions. Only last 45 years (5 cycles) of CMIP simulations were analyzed and the first 54 years are used as a model spinup. Five-member ensembles of AMIP-type simulations (fixed SST and sea ice) were run with and without fire emissions, respectively, to represent the rapid adjustment to fire aerosols. Each AMIP simulation includes one fire cycle (9 years; 2003–11) and the first year was repeated for spinup (a total of 10 years). Thus, in total 45-yr AMIP simulations (5 cycles) are analyzed, which is comparable to CMIP-type

TABLE 2. Summary of notation.

Symbols for quantities	
N	Net radiative flux at TOA
S	Net surface energy
RE	Radiative effect
CF	Climate feedback
SE	Surface effect
SCF	Surface climate feedback
ΔT	Global annual-mean surface air temperature change
TOA	Top of the atmosphere
Subscripts denoting components	
SW	Shortwave
LW	Longwave
TOA	Top of the atmosphere
SH	Surface sensible heat flux
LH	Surface latent heat flux
AMIP	Fixed-SST simulation
CMIP	Coupled simulation
ARI	Aerosol–radiation interaction
ACI	Aerosol–cloud interaction
SAC	Surface-albedo change
CC	Clean-and-clear sky

simulations. The notation of this study is summarized in Table 2.

3. Methodology

a. Estimation of aerosol radiative effect and fast response

The fast response refers to the climate adjustment before any change in global annual-mean surface air temperature ΔT occurs. The radiative effect (forcing) means the fast adjustment of the net radiative fluxes at the top of the atmosphere (TOA). Both the fast response and radiative effect can be estimated by the response to external forcing in AMIP type (fixed-SST and sea ice) simulations, which is known as the “fixed-SST” method introduced by Hansen et al. (2002). The radiative effect estimated with such a method is known as the effective radiative effect (ERE; IPCC 2014). It is noted that, with such a method, there could be a small climate change (ΔT ; 0.03 K) resulting from change in land surface temperature (Bala et al. 2009).

b. Estimation of aerosol climate feedback (slow response)

The climate feedback or slow response is usually indicated by the change in a specific variable (e.g., radiative flux) per unit ΔT (the difference of global annual-mean surface air temperature between CMIP- and AMIP-type simulations). The climate change in the coupled system (CMIP type) represents the total climate

response, which is the sum of both fast and slow responses. Following Bala et al. (2009), the climate feedback can be calculated by subtracting the fast response from the total response in the CMIP-type simulations. Thus, the climate feedback parameter (CFP) can be calculated by

$$\text{CFP} = (\Delta N_{\text{CMIP}} - \Delta N_{\text{AMIP}})/|\Delta T|, \quad (1)$$

where the CFP is scaled by the absolute value of ΔT to keep its original sign. Thus, a positive CFP value means that the climate feedback process induces a warming, and vice versa.

c. Estimation of aerosol surface effect and surface feedback

The fire aerosols also affect the surface energy budget, in which both radiative (longwave and shortwave) and nonradiative (sensible and latent heat) fluxes are included. Understanding the linkage between radiative and nonradiative fluxes are important, since the fire aerosols induced surface solar flux reduction should be balanced by the nonradiative fluxes. Thus, following Andrews et al. (2009), the surface effect (SE) and surface feedback (SF) is defined as the fast and slow adjustment of surface energy S to aerosols, respectively.

$$\Delta S = \Delta S_{\text{sw}} + \Delta S_{\text{lw}} + \Delta S_{\text{sh}} + \Delta S_{\text{lh}}. \quad (2)$$

d. Decomposition of aerosol radiative effects

The method proposed by Ghan (2013) is used to separate the RE into different terms induced by different aerosol effects (e.g., ARI, ACI, and SAC). A summary of the method can be found in Jiang et al. (2016). Here is a brief introduction. As shown in Eq. (3), N_{clean} is the radiative flux at TOA calculated from a diagnostic radiation call in the same control simulations, but neglecting the scattering and absorption of solar radiation by aerosols; $N_{\text{clean,clear}}$ is the clear-sky radiative flux at TOA calculated from the same diagnostic radiation call, but neglecting scattering and absorption by both clouds and aerosols. Thus, the TOA radiative flux change could be written to three terms, which are due to ARI, ACI, and SAC, respectively. Accordingly, the radiative effect could be broken into three terms [Eq. (4), i.e., ARI, ACI, and SAC]:

$$\begin{aligned} \Delta N = & \underbrace{\Delta(N - N_{\text{clean}})}_{\Delta N^{\text{ARI}}} + \underbrace{\Delta(N_{\text{clean}} - N_{\text{clean,clear}})}_{\Delta N^{\text{ACI}}} \\ & + \underbrace{\Delta N_{\text{clean,clear}}}_{\Delta N^{\text{CC}}}, \end{aligned} \quad (3)$$

$$\text{RE} = \text{RE}^{\text{ARI}} + \text{RE}^{\text{ACI}} + \text{RE}^{\text{SAC}}. \quad (4)$$

e. Decomposition of aerosol climate feedbacks

The decomposing methods proposed by Ghan (2013) also could be applied to the coupled simulations, and the radiative flux change in the coupled simulations ΔN_{CMIP} could be separated into three items (i.e., $\Delta N_{\text{CMIP}}^{\text{ARI}}$, $\Delta N_{\text{CMIP}}^{\text{ACI}}$, and $\Delta N_{\text{CMIP}}^{\text{CC}}$), similarly. In each item, both the effects of the rapid adjustment (from ARI, ACI, and SAC) and the impacts of aerosol climate feedbacks on these adjustments are included. Thus, the difference with the corresponding radiative effect component represents the different climate feedback process, respectively [Eqs. (5) and (6)]. The first (CFP^{ARI}) and second ($\text{CFP}^{\text{Cloud}}$) items of Eq. (6) represent the impacts of climate feedback on aerosol–radiation interaction (i.e., aerosol feedback) and aerosol–cloud interaction (i.e., cloud feedback), respectively. The third term (CFP^{CC}) is irrelevant to both impacts of aerosols and cloud. Its SW component is due to planetary albedo change (albedo feedback), and its LW component represents the longwave feedback (e.g., blackbody feedback and lapse rate feedback) associated with tropospheric warming (cooling).

$$\begin{aligned} \text{CFP} = & (\Delta N_{\text{CMIP}}^{\text{ARI}} - \Delta N_{\text{AMIP}}^{\text{ARI}})/|\Delta T| \\ & + (\Delta N_{\text{CMIP}}^{\text{ACI}} - \Delta N_{\text{AMIP}}^{\text{ACI}})/|\Delta T| \\ & + (\Delta N_{\text{CMIP}}^{\text{CC}} - \Delta N_{\text{AMIP}}^{\text{CC}})/|\Delta T|, \end{aligned} \quad (5)$$

$$\text{CFP} = \text{CFP}^{\text{ARI}} + \text{CFP}^{\text{Cloud}} + \text{CFP}^{\text{CC}}. \quad (6)$$

4. Results

a. Fire aerosol distribution

Figure 1 shows the latitudinal and longitudinal distributions of aerosol optical depth (AOD) change induced by fire aerosols. Both the AOD changes in both AMIP- (Fig. 1a) and CMIP-type simulations (Fig. 1b) are quite similar to each other, which implies that the SST impact on global fire aerosol distribution is small. The AOD increase is mostly from the particulate organic matter (POM) and black carbon (BC) emitted by the fires. The AOD change is distributed in two latitudinal zones: one in the tropics (30°S–15°N; southern Africa, South America, and Southeast Asia) and the other in the middle to high latitudes (north of 45°N) of the Northern Hemisphere (Northeast Asia, Alaska, and Canada). The largest AOD change is found in southern Africa (~0.15) and South America (~0.1), which is mostly from deforestation, savanna, and grassland fires. The AOD increase in the middle to high latitudes is

FIG. 1. Spatial distribution of the change in AOD at 550 nm induced by fire aerosols in (a) AMIP- and (b) CMIP-type simulations.

relatively small (~ 0.03) and mostly from forest fires. As a result, the fire BC to organic carbon (OC) ratio (BC/OC) in the tropical regions is about 3 times higher than that of high latitudes, since fire emissions in low latitudes come mainly from the flaming phase of burning (Jiang et al. 2016). A high fire BC ratio leads to strong atmospheric absorption in southern Africa and South America (figure not shown).

Figure 2 compares the simulated AOD in both AMIP and CMIP simulations with observations from the Aerosol Robotic Network (AERONET; <http://aeronet.gsfc.nasa.gov>) at sites significantly affected by biomass burning activity in southern Africa, South America, and the Arctic regions. The AERONET AOD data are averaged for the years from 2003 to 2011 (with missing values for some

years) to match the fire emissions period. The simulated AOD are quite similar in both AMIP- and CMIP-type simulations, which is consistent with Fig. 1. In both southern Africa and South America, the modeled monthly AOD is generally consistent with observations within a factor of 2. The underestimation is the most obvious in the fire season (large AOD values), which may be attributed to the underestimation of fire emissions in these two regions (Jiang et al. 2016; Tosca et al. 2013). The model significantly underestimates the AOD in the Arctic for all months. The underestimation of AOD in the Arctic could be primarily due to the excessive scavenging of aerosols during their transport from the midlatitude industrial regions by liquid-phase clouds (Wang et al. 2013). Meanwhile, the underestimation

FIG. 2. Comparison of modeled monthly mean AOD from (a) Fire_AMIP and (b) Fire_CMIP simulations with observations (2003–11) from the AERONET sites of wildfire regions (southern Africa, South America, and the Arctic). Dashed lines are 1:2 and 2:1 for AOD.

FIG. 3. Annual-mean (a) total (SW + LW) radiative effect, (b) SW radiative effect due to aerosol-radiation interactions, (c) SW radiative effect due to aerosol-cloud interactions, and (d) LW radiative effect (all W m^{-2}) due to fire aerosols. The plus signs denote the regions where the radiative effect is statistically significant at the 0.05 level.

of fire emissions in the NH high latitudes (e.g., Stohl et al. 2013) and/or fossil fuel emissions in Asia (e.g., Cohen and Wang 2014) could also contribute to the underestimation.

b. Radiative effect and total response

The global annual-mean radiative effects (RE) due to fire aerosols are shown in Fig. 3 and Table 3. The total (SW + LW) fire radiative effect is $-0.78 \pm 0.29 \text{ W m}^{-2}$ (cooling) and the RE is the largest in the Arctic (north of 60°N) and tropical regions (Fig. 3a). The fire radiative effect could be from ARI, ACI, and SAC. The SW RE due to ARI (SW REari) is $0.16 \pm 0.04 \text{ W m}^{-2}$ (warming), and the LW REari is small and insignificant ($0.01 \pm 0.01 \text{ W m}^{-2}$). The largest SW REari (Fig. 3b) is located in ocean areas west of southern Africa ($\sim 5.0 \text{ W m}^{-2}$) and South America ($\sim 2 \text{ W m}^{-2}$). It is because biomass burning aerosols are transported above the low-level stratocumulus clouds and its absorption is amplified through the reflection of solar radiation by underlying clouds (Abel et al. 2005; Zhang et al. 2016). Both SW and LW radiative effects due to ACI are negative ($-0.70 \pm 0.20 \text{ W m}^{-2}$ and $-0.25 \pm 0.14 \text{ W m}^{-2}$, cooling, respectively). In the tropical regions, strong negative

SW REaci is located in the adjacent ocean areas of southern Africa and South America (Fig. 3c), which is due to the brightening effects of fire aerosols on stratocumulus clouds (Lu et al. 2018). Comparable SW REaci change is also found in the Arctic regions (Fig. 3c) and the effect is the most significant in boreal summer (-15 W m^{-2}), which is in general consistent with observation (Zhao and Garrett 2015). Both the large cloud liquid water path over land areas (Jiang et al. 2015) and the low solar zenith angle (Jiang et al. 2016) favor the large aerosol indirect effects of Arctic summer. The LW REaci (Fig. 3d) is the largest over the tropical land regions (Africa and South America, cooling), which is due to the suppression of convection in these regions (discussed in section 4c). The suppression of convection tends to trap less outgoing longwave radiation and induce a cooling, which could explain the negative LW REaci over the tropical Indian Ocean. Both the SW and LW radiative effects due to SAC are small and insignificant ($-0.02 \pm 0.09 \text{ W m}^{-2}$ and $0.04 \pm 0.13 \text{ W m}^{-2}$, respectively).

Associated with negative net radiative effect at TOA, the global annual-mean surface air temperature (SAT) is reduced by $-0.64 \pm 0.16 \text{ K}$ in coupled simulation

TABLE 3. Global annual-mean fire aerosol radiative effects and surface effects (both in W m^{-2}). The boldface values are statistically significant at the 0.05 level.

Radiative effect	
Total (SW + LW) radiative effect	-0.78 ± 0.29
Total SW radiative effect	-0.57 ± 0.20
Total LW radiative effect	-0.20 ± 0.24
SW radiative effect due to ARI	0.16 ± 0.04
SW radiative effect due to ACI	-0.70 ± 0.20
SW radiative effect due to SAC	-0.02 ± 0.09
LW radiative effect due to ARI	0.01 ± 0.01
LW radiative effect due to ACI	-0.25 ± 0.14
LW clean-and-clear sky radiative effect	0.04 ± 0.13
Surface effect	
Net surface effect	-0.76 ± 0.29
Surface effect due to SW radiative flux change	-1.41 ± 0.28
Surface effect due to LW radiative flux change	0.15 ± 0.14
Surface effect due to sensible heat flux change	0.21 ± 0.08
Surface effect due to latent heat flux change	0.28 ± 0.24

(total response). The cooling is the largest in the Arctic regions and the maximum cooling is near the North Pole ($\sim 2\text{ K}$; Fig. 4a). The SAT change in the tropical fire regions ($\sim 0.5\text{ K}$) is much smaller than in the Arctic, although the radiative effect in both of regions are comparable (Fig. 3a); this is because the stable lapse rate at high latitude favors confining the cooling to low levels (Hansen et al. 1997; Forster et al. 2000). It is clear that the SAT change in the land regions is much larger than that in the ocean regions. The significant land–sea contrast in the SAT change is due to the relatively small heat capacity of land surface. The global annual-mean precipitation is reduced by $-0.06 \pm 0.01\text{ mm day}^{-1}$. The precipitation reduction is the largest in deep tropical ocean regions (equatorial Atlantic Ocean, east Pacific Ocean, and Maritime Continent), and an increase of

precipitation is found in the tropical ocean regions of Southern Hemisphere (15°S ; Fig. 4b). The precipitation is also decreased in the middle to high latitudes of the Northern Hemisphere (e.g., North Pacific Ocean, western United States). Overall, the precipitation is decreased in most of regions of Northern Hemisphere, but increases in Southern Hemisphere tropical ocean regions. Such precipitation redistributions could be directly from the fire radiative effect (fast response) and indirectly from slow climate feedback (associated with the ΔT), which will be discussed below, respectively.

c. Fast response

The SAT and precipitation changes due to fire aerosols in the AMIP-type simulations are shown in Fig. 5. As expected, the global annual-mean SAT change is very small (-0.03 K ; cooling) and the cooling is constrained in the land regions. The cooling is the largest in the Arctic land regions (-0.5 K ; Fig. 5a), which is correlated to strong REaci of boreal forest fire aerosols (Fig. 3c). In summer, the Arctic cooling could be up to 1.5 K (figure not shown), which is also noticed in Jiang et al. (2016). Although the rapid adjustment of SAT to fire aerosol effect could be significant for some regions (e.g., the Arctic), the land SAT change will eventually be dominated by the ocean response (Fig. 4a). The global annual-mean precipitation is reduced by 0.01 mm day^{-1} before the SST adjustment (fast response). The precipitation change is the largest in the tropical land regions (southern Africa and South America), while the change over the ocean areas is small and insignificant (Fig. 5b).

The surface effect (SE) components are shown in Fig. 6 and Table 3 to understand the SAT and precipitation change. The surface SW radiative flux is significantly reduced by $1.41 \pm 0.2\text{ W m}^{-2}$ (cooling) due to fire aerosols and the reduction is the largest in the tropical

FIG. 4. Annual-mean (a) surface air temperature (K) and (b) precipitation (mm day^{-1}) change due to fire aerosols in CMIP-type simulations. The plus signs denote the regions where the change is statistically significant at the 0.05 level.

FIG. 5. As in Fig. 4, but for the change in AMIP-type simulations.

fire regions (Fig. 6a). The surface dimming is balanced by the decrease of upward longwave radiative flux ($0.15 \pm 0.14 \text{ W m}^{-2}$) and sensible ($0.21 \pm 0.08 \text{ W m}^{-2}$) and latent ($0.28 \pm 0.24 \text{ W m}^{-2}$) heat flux into the atmosphere. All these compensating SE components are the most significant over the land regions, due to the rapid adjustments of land surface. As a result, the net surface energy of land regions could reach a balance

before the SST adjustment. The surface energy of ocean regions remains imbalanced, which needs further SST feedback.

The longwave radiative flux reduction is the largest in the Arctic regions ($\sim 2 \text{ W m}^{-2}$; positive) associated with the Arctic surface cooling (Fig. 6b). The turbulent components (sensible and latent heat flux), however, are the largest in the tropical land regions (Figs. 6c,d). The

FIG. 6. Annual-mean fire aerosol surface effect (SE; W m^{-2}) due to (a) SW radiative flux change, (b) LW radiative flux change, (c) surface sensible heat (SH) flux change, and (d) surface latent heat (LH) flux change. The plus signs denote the regions where the change is statistically significant at the 0.05 level.

TABLE 4. Global annual-mean TOA climate feedbacks and surface feedbacks induced by fire aerosols (both in $\text{W m}^{-2} \text{K}^{-1}$, scaled by $|\Delta T| = 0.61 \text{ K}$). Positive values (warming) mean negative feedback, and vice versa. The boldface values are statistically significant at the 0.05 level.

TOA climate feedback parameter	
Total (LW + SW) feedback	1.28 ± 0.77
SW feedback	-0.80 ± 0.61
LW feedback	2.08 ± 0.65
SW feedback on aerosol–radiation interaction	0.01 ± 0.08
SW cloud feedback	-0.11 ± 0.49
SW clean-and-clear sky feedback	-0.70 ± 0.43
LW feedback on aerosol–radiation interaction	0.00 ± 0.01
LW cloud feedback	0.14 ± 0.31
LW clean-and-clear sky feedback	1.94 ± 0.54
Surface feedback parameter	
Net surface feedback	1.32 ± 0.71
Surface feedback due to SW radiative flux change	0.09 ± 0.69
Surface feedback due to LW radiative flux change	-0.83 ± 0.41
Surface feedback due to sensible heat flux change	-0.40 ± 0.27
Surface feedback due to latent heat flux change	2.47 ± 0.75

tropospheric heat budget is analyzed to better illustrate such a distribution. The net tropospheric heat budget should be very small, since the atmosphere has a relatively small heat capacity (Andrews et al. 2009). If the radiative effect at the surface is not the same as that at the TOA (i.e., strong atmospheric absorption), then there must be an induced turbulent component in the surface effect to maintain the tropospheric heat balance. In the tropical fire regions, the TOA radiative effect (Fig. 3a) is much smaller than the surface radiative effect (Figs. 6a,b), which is due to strong fire BC absorption ($\sim 8 \text{ W m}^{-2}$). As a result, both the latent and sensible heat fluxes (turbulent components) are significantly reduced (positive) to conserve the tropospheric heat budget. A reduction of latent heat flux implies less transportation of moisture from the surface to the atmosphere, which could explain why the precipitation significantly reduces in the tropical land regions. In the Arctic regions, the atmospheric absorption ($\sim 2 \text{ W m}^{-2}$) is weaker due to the lower BC/OC ratio. As a result, the compensating latent heat flux ($\sim 1 \text{ W m}^{-2}$) there is much smaller than that in the tropical fire regions, which explains why precipitation change in the Arctic fire region is relatively small and insignificant.

d. Climate feedback

As shown in section 4b, the fire radiative effect reduces the TOA net radiation by $-0.78 \pm 0.29 \text{ W m}^{-2}$, which leads to an energy imbalance at TOA. Such an imbalance should be compensated by the negative

FIG. 7. Annual-mean TOA (a) total (SW + LW), (b) SW, and (c) LW cloud feedback ($\text{W m}^{-2} \text{K}^{-1}$) due to fire aerosols.

climate feedback process ($1.28 \pm 0.77 \text{ m}^{-2} \text{K}^{-1}$), so that the net energy budget of climate system becomes balanced again (Table 4). The longwave feedback acts as a main negative feedback ($2.08 \pm 0.65 \text{ W m}^{-2} \text{K}^{-1}$), while the shortwave feedback tends to amplify the energy imbalance ($-0.80 \pm 0.61 \text{ W m}^{-2} \text{K}^{-1}$; positive feedback).

Both SW and LW feedbacks on aerosol–radiation interaction are small and insignificant (Table 4), because the global fire aerosol distribution does not change much associated with the SST adjustment (Fig. 1). The global annual-mean cloud feedback (Fig. 7a) is also very small ($0.03 \text{ W m}^{-2} \text{K}^{-1}$). The negative cloud feedback (positive) mainly distributes in high-latitude ocean areas (Arctic Ocean and Southern Ocean), while the positive cloud feedback (negative) is

FIG. 8. Annual-mean TOA (a) SW and (b) LW clean-and-clear-sky climate feedback ($\text{W m}^{-2} \text{K}^{-1}$) due to fire aerosols.

mainly located in boreal midlatitude ocean areas (e.g., North Pacific Ocean). The net cloud feedback of tropical ocean regions is relatively small, since the SW (Fig. 7b) and LW (Fig. 7c) cloud feedbacks tend to offset each other.

The SW and LW feedbacks in a clean-and-clear-sky (CC) situation are the most important positive and

negative climate feedbacks, respectively. The SW feedback (Fig. 8a) is the largest in the Arctic, which is correlated to the Arctic sea ice extent increase (figure not shown) as well as the surface albedo increase (cooling). Positive ice-albedo feedback further decreases the SAT in the Arctic, which could explain why the maximum surface cooling is near the North Pole (Fig. 4a). The

FIG. 9. Annual-mean fire aerosol surface feedback ($\text{W m}^{-2} \text{K}^{-1}$) due to (a) SW radiative flux change, (b) LW radiative flux change, (c) surface sensible heat (SH) flux change, and (d) surface latent heat (LH) flux change. The plus signs denote the regions where the change is statistically significant at the 0.05 level.

TOA energy budget is finally balanced by the LW feedback in the CC situation ($2.08 \pm 0.65 \text{ W m}^{-2} \text{ K}^{-1}$). The blackbody radiation feedback is the most important negative feedback, and the distribution of longwave flux change (Fig. 8b) is in general consistent with the tropospheric cooling. The tropospheric cooling is the largest in the mid- to high-latitude regions of Northern Hemisphere, which is associated with strong boreal forest fire aerosol indirect effect and associated ice-albedo feedback.

The surface energy budget also reaches a balance through the surface feedback process ($1.32 \pm 0.71 \text{ W m}^{-2} \text{ K}^{-1}$). The longwave radiative flux change ($-0.83 \pm 0.41 \text{ W m}^{-2} \text{ K}^{-1}$; Fig. 9b) and sensible heat flux change ($-0.40 \pm 0.27 \text{ W m}^{-2} \text{ K}^{-1}$; Fig. 9c) act as positive feedbacks (cooling), which further amplifies the surface energy imbalance. It is because the cooling of free troposphere is larger than that at surface (e.g., the tropical regions), which increases upward sensible heat flux (negative) and decreases downward longwave flux (negative) at surface. The surface shortwave flux change ($0.09 \pm 0.69 \text{ W m}^{-2} \text{ K}^{-1}$; Fig. 9a) is small and insignificant. The latent heat flux change is the most important negative surface feedback ($2.47 \pm 0.75 \text{ W m}^{-2} \text{ K}^{-1}$; Fig. 9d). The latent heat change is the most significant over the ocean regions, which is associated with the cooling of sea surface temperature (SST). In the tropical regions, the latent heat flux reduction is the largest over the eastern Pacific Ocean and Atlantic Ocean, since low SST in these regions favors a positive SST–LH flux correlation (Zhang and McPhaden 1995).

Decreased upward latent heat flux (positive) tends to reduce the global precipitation by $-0.05 \text{ mm day}^{-1}$ so as to keep global water balance, and the reduction is mostly over the oceanic areas (Fig. 10a). The precipitation reduction is the largest over the deep tropical ocean areas (5°N) and mostly from the convective precipitation change (Fig. 10b). An increase of precipitation is found in Southern Hemisphere tropical ocean areas (15°S). Such a redistribution in tropical precipitation could be associated with large-scale vertical circulation change, since there is abundant water vapor in the tropics. Thus, the tropical vertical meridional circulation change will be analyzed in next subsection.

e. Tropical circulation change

The fire aerosol-induced zonally averaged vertical motion change and its fast and slow components are shown in Fig. 11. The change is mostly from slow climate feedback (Fig. 11d), and its fast response components is small and insignificant (Fig. 11c). The climatological vertical velocity shows maxima in the tropics (Fig. 11a),

FIG. 10. Annual-mean (a) total precipitation, (b) convective precipitation, and (c) large-scale precipitation (all in mm day^{-1}) change due to fire aerosol climate feedbacks. The plus signs denote the regions where the change is statistically significant at the 0.05 level.

consistent with the positions of the intertropical convergence zone (ITCZ). The vertical velocity change due to fire aerosols exhibits wavelike perturbations, featuring anomalous sinking motions in the deep tropics (5°N) and rising motions in its south sides (15°S ; Fig. 11b), which could be seen as a southward shift of ITCZ. Strong tropical sinking (rising) motions could be accompanied with the compensating reversal rising (sinking) motions on its subtropical side, which could explain the wavelike perturbation (Lau and Kim 2015). The precipitation is significantly reduced in the deep tropics (5°N) but increased at 15°S (Fig. 12a), which is consistent with the vertical motion change.

FIG. 11. Zonally averaged (annual) vertical velocities ($-10^{-2} \text{ Pa s}^{-1}$) from (a) FIRE_CMIP simulation and (b) the difference between the FIRE_CMIP and NOFIRE_CMIP simulation. (c),(d) As in (b), but for its fast and slow components, respectively. Positive (negative) values indicate upward (downward) velocities. The plus signs in (b)–(d) denote the regions where the change is statistically significant at the 0.05 level.

The zonally averaged mass streamfunction (MSF) is shown in Fig. 13 to examine the meridional overturning circulation change. The climatological annual-mean MSF exhibits two cells which are generally symmetric about the equator (5°N ; Fig. 13a), which represent two cells of Hadley circulation (HC). The climatological Southern (Northern) Hemisphere HC is anticlockwise (clockwise), which implies that low-level (upper-level) transportation is equatorward (poleward) for both hemispheres. The MSF anomaly due to fire aerosols is positive (clockwise) for most of the tropical regions, and the clockwise anomaly is the largest between 20°S and 5°N (Fig. 13b). It implies that the Southern Hemisphere (SH) HC is weakened, while the Northern Hemisphere (NH) HC is slightly enhanced. The variations of HC are strongly correlated to the temperature contrast between the Northern and Southern Hemispheres (i.e., the inter-hemispheric temperature asymmetry), as noticed in many previous studies (e.g., Chiang and Friedman 2012). Both the boreal forest indirect effect and ice-albedo feedback induce greater cooling in Northern Hemisphere (Fig. 12b). Such an interhemispheric temperature asymmetry could be compensated by the weakening Southern Hemisphere HC. Associated with the HC weakening, its ascending branch shifts southward and leads to an anomalous cross-equatorial mass transport from the SH (NH) to the NH (SH) in the upper (lower) troposphere.

More atmospheric energy is transported from the SH to the NH, since the moist static energy is larger in the upper troposphere. Increased atmospheric energy from the SH to the NH could partly compensate larger NH

FIG. 12. Zonally averaged (annual) (a) precipitation (mm day^{-1}) and (b) tropospheric temperature (K) changes in CMIP-type simulations. The plus signs in (b) denote the regions where the change is statistically significant at the 0.05 level.

FIG. 13. Zonally averaged (annual) mass streamfunctions (10^9 kg s^{-1}) from (a) FIRE_CMIP simulation and (b) the difference between the FIRE_CMIP and NOFIRE_CMIP simulation. (c),(d) As in (b), but for its fast and slow components, respectively. The positive (negative) values denote to the clockwise (counterclockwise) rotation. The plus signs in (b)–(d) denote the regions where the change is statistically significant at the 0.05 level.

radiative cooling and thus decrease the interhemispheric temperature asymmetry.

5. Conclusions and discussion

In this study, fire aerosol impacts on global energy budget and climate are investigated with CESM. The fire aerosols-induced AOD change mainly distributes in two latitude zones: one is in the tropical regions (southern Africa, South America, and Southeast Asia) and the other is in the middle to high latitudes of Northern Hemisphere. The fire BC to OC ratio (BC/OC) in the tropical regions is about 3 times higher than that in the high latitudes of Northern Hemisphere, which results in a stronger atmospheric absorption in the tropics.

Accordingly, the global fire RE ($-0.78 \pm 0.29 \text{ W m}^{-2}$; cooling) is the largest in the tropics and in the Arctic. Both the SW REaci ($-0.70 \pm 0.20 \text{ W m}^{-2}$) and LW REaci ($-0.25 \pm 0.14 \text{ W m}^{-2}$) contribute mostly to the cooling, which emphasizes the role of the fire aerosol indirect effect. The SW REaci is the largest in the tropical ocean areas as well as in the Arctic land regions, which is due to strong aerosol indirect effect in those areas. The LW REaci is the largest in the tropical land regions (cooling), which is correlated to the convection suppression due to fire aerosols. The REari is relatively

small ($0.16 \pm 0.04 \text{ W m}^{-2}$) and could partly compensate the cooling of REaci. The energy imbalance at TOA (i.e., fire RE) should be balanced by the climate feedback processes associated with global SAT change ΔT ($-0.61 \pm 0.16 \text{ K}$). Longwave feedback is the most important negative feedback ($2.08 \pm 0.65 \text{ W m}^{-2} \text{ K}^{-1}$) and mostly from the blackbody radiation feedback associated with the atmospheric radiative cooling. The SW feedback tends to amplify the energy imbalance ($-0.80 \pm 0.61 \text{ W m}^{-2} \text{ K}^{-1}$, cooling), which is the largest in the Arctic regions. It is significantly increased and induces a positive shortwave albedo feedback (cooling), which significantly reduces the SAT in the Arctic. The global-mean feedback on aerosol–radiation interaction ($0.01 \text{ W m}^{-2} \text{ K}^{-1}$) and cloud feedback ($0.03 \text{ W m}^{-2} \text{ K}^{-1}$) are relatively small and insignificant.

Both surface effect and surface feedback (summarized in Fig. 14) are analyzed to understand the fast and slow responses of precipitation. For the fast adjustments (i.e., surface effect; Fig. 14a), the significant surface SW radiation reduction ($-1.41 \pm 0.2 \text{ W m}^{-2}$, cooling) is balanced by the decrease of all the other components (LW, SH, and LH) into atmosphere. The latent heat flux change is the most important compensating SE components ($0.28 \pm 0.24 \text{ W m}^{-2}$), and the reduction is the largest in the tropical land regions. Strong fire BC

FIG. 14. The components of (a) surface effect and (b) surface climate feedback due to fire aerosols.

atmospheric absorption in the tropics should be balanced by the turbulent (SH and LH) flux reduction to conserve the tropospheric heat budget. Significant latent heat flux reduction in the tropics could explain the significant precipitation decrease in the tropical land regions. The net surface effect exhibits a cooling ($-0.76 \pm 0.29 \text{ W m}^{-2}$), which needs to be further balanced by the surface feedbacks. The latent heat flux (LH) feedback is the most important negative surface feedback ($2.47 \pm 0.75 \text{ W m}^{-2} \text{ K}^{-1}$), while the LW and SH feedbacks act as positive feedbacks (Fig. 14b). Significant latent heat flux reduction could explain why most (80%) of the global precipitation change is associated with the climate feedback processes.

The precipitation reduction is the most significant in deep tropical regions (5°N). Accordingly, boreal ITCZ shift southward and the Southern Hemisphere Hadley cell is weakened. The cross-equatorial moist static energy transport is increased, which could partly

compensate interhemispheric temperature asymmetry (Fig. 15a). Such an asymmetry is associated with the significant NH radiative cooling induced by boreal forest aerosol indirect effect. The Hadley cell change is different from the response found in Tosca et al. (2013). In their study, only fire aerosol direct effect is considered and the Hadley cells of both hemispheres are weakened, and the weakening is in general symmetry about the equator (Fig. 15b), since the interhemispheric temperature asymmetry induced by fire aerosol direct effect is small. Besides, both the global precipitation ($-0.03 \text{ mm day}^{-1}$) and SAT change (-0.13 K) shown in Tosca et al. (2013) are much smaller than those in our study, which further emphasized the important role of fire aerosol indirect effect.

Our model results could provide some implications for fire–climate interaction under the global warming background. While the greenhouse gas (CO_2) tends to deepen the ITCZs (Lau and Kim 2015) and induces a

FIG. 15. Schematic diagram of fire aerosol effects on Hadley circulation for (a) all (direct and indirect) fire aerosol effects and (b) only fire aerosol direct effect; (b) is based on results of Tosca et al. (2013). The boreal forest aerosol indirect effect induces anomalous NH cooling, and clockwise anomalous Hadley circulation transports energy northward (red arrow) to compensate the interhemispheric asymmetry. The fire aerosol direct effect weakens the Hadley cells of both hemispheres (symmetry about the equator), since the interhemispheric temperature asymmetry induced by the fire aerosol direct effect is small.

significant warming in the Arctic (Cohen et al. 2012), the fire aerosol effect tends to compensate the impact of greenhouse gases in both the tropics and Arctic. The global warming tends to increase wildfire activities and emit more CO₂, which could produce a positive feedback. The wildfire-emitted aerosols, however, could compensate the impact of CO₂ and act as a negative feedback.

Recent study indicated that a slab-ocean model could exaggerate the response of cross equatorial heat transport in the atmosphere and the ITCZ shift to aerosol forcing, as compared with a fully coupled model (Zhao and Suzuki 2019). They highlighted the importance of using a fully coupled model in modeling studies of the tropical climate response to aerosols. The fully coupled simulations will be applied in our future study to investigate fire aerosol effects. The results of this study are based on CESM simulations, which could be a first step to understand fire aerosol effects. Also, we will try to find more evidence from observation by comparing features between strong and weak fire years in the future studies.

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