Impacts of Global Open Fire Aerosols on Direct Radiative, Cloud and Surface-Albedo Effects Simulated with CAM5

Yiquan Jiang[1,2], Zheng Lu[2], Xiaohong Liu[2,*], Yun Qian[3], Kai Zhang[3], Yuhang Wang[4] and Xiu-Qun Yang[1]

[1] *CMA-NJU Joint Laboratory for Climate Prediction Studies, Institute for Climate and Global Change Research, School of Atmospheric Sciences, Nanjing University, Nanjing, China*

[2] *Department of Atmospheric Science, University of Wyoming*

[3] *Pacific Northwest National Laboratory, Richland, Washington, USA*

[4] *School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, Georgia, USA*

*Corresponding author:

Dr. Xiaohong Liu
Department of Atmospheric Science
University of Wyoming
Laramie, WY 82071
Phone (307) 766-3225
E-mail: xliu6@uwyo.edu

Abstract

Aerosols from open-land fires could significantly perturb the global radiation balance and induce the climate change. In this study, Community Atmospheric Model version 5 (CAM5) with prescribed daily fire aerosol emissions is used to investigate the spatial and seasonal characteristics of radiative effects (REs, relative to the case of no fires) of open fire aerosols including black carbon (BC) and particulate organic matter (POM) from 2003 to 2011. The global annual mean RE due to aerosol-radiation interactions (REari) of all fire aerosols is 0.16±0.01 W m^{-2} (1σ uncertainty), mainly due to the absorption of fire BC (0.25±0.01 W m^{-2}), while fire POM induces a small overall effect (-0.05 W m^{-2} and 0.04±0.01 W m^{-2}, respectively based on two methods). Strong positive REari is found in the Arctic and in the oceanic regions west of southern Africa and South America as a result of amplified absorption of fire BC above low-level clouds, in general agreement with satellite observations. The global annual mean RE due to aerosol-cloud interactions (REaci) of all fire aerosols is -0.70±0.05 W m^{-2}, resulting mainly from the fire POM effect (-0.59±0.03 W m^{-2}). REari (0.43±0.03 W m^{-2}) and REaci (-1.38±0.23 W m^{-2}) in the Arctic are stronger than those in the tropics (0.17±0.02 and -0.82±0.09 W m^{-2}, respectively for REari and REaci), although the fire aerosol burden is higher in the tropics. The large cloud liquid water path over land areas and low solar zenith angle of the Arctic favor the strong fire aerosol REaci (up to -15 W m^{-2}) during the Arctic summer. Significant surface cooling, precipitation reduction and low-level cloud amount increase are also found in the Arctic summer as a result of the fire aerosol REaci based on the atmosphere-only simulations. The global annual mean RE due to

53 surface albedo changes (REsac) over land areas (0.03 ± 0.10 W m^{-2}) is small and
54 statistically insignificant, and is mainly due to the fire BC-in-snow effect (0.02 W m^{-2})
55 with the maximum albedo effect occurring in spring (0.12 W m^{-2}) when snow starts to
56 melt.
57
58 **1. Introduction**
59 Open fires or biomass burning of living and dead vegetation are an integral
60 component of the Earth system, and have significant impacts on the carbon cycle
61 [*Ciais et al.*, 2013] and the climate [*Bowman et al.*, 2009; *Keywood et al.*, 2011; *Liu et*
62 *al.*, 2014; *Sommers et al.*, 2014; *Voulgarakis and Field*, 2015]. On one hand, open
63 fires can perturb the climate system by emitting greenhouse gases and aerosols
64 [*Kaiser et al.*, 2012; *Wiedinmyer et al.*, 2011]. On the other hand, climate states and
65 variabilities can play a critical role in determining the occurrence frequency and
66 intensity of open fires [*Marlon et al.*, 2009; *van der Werf et al.*, 2008; *Westerling et*
67 *al.*, 2006; *Bistinas et al.*, 2014]. However, there are still large unknowns regarding the
68 feedback mechanisms between open fire and climate interactions [*Carslaw et al.*,
69 2010; *Liu et al.*, 2014]. A qualification of radiative forcing of fire aerosols as
70 conducted in this study is the first step to reduce these uncertainties.
71 Particles emitted from open fires can exert significant perturbations to the
72 climate system by scattering and absorbing the solar radiation in the atmosphere (i.e.,
73 direct effect) [*Carslaw et al.*, 2010] and by changing the surface albedo when they are
74 deposited on the snow and ice (i.e., surface albedo effect) [*Flanner et al.*, 2007; *Quinn*
75 *et al.*, 2008; *Randerson et al.*, 2006; *Qian et al.*, 2011, 2015]. In addition, open fire or

76 smoke particles can modify the cloud properties, precipitation efficiency, and the
77 hydrological cycle by changing the atmospheric thermal structure (i.e., semi-direct
78 effect) [*Koch and Del Genio*, 2010; *Andreae et al.*, 2004b] or acting as cloud
79 condensation nuclei (CCN) (i.e., indirect effects) [*Andreae and Rosenfeld*, 2008; *Qian*
80 *et al.*, 2009; *Lu and Sokolik*, 2013].

81 The radiative effect (RE) [*Boucher and Tanre*, 2000] and radiative forcing (RF)
82 [*Forster et al.*, 2007; *Myhre et al.*, 2013a] are typical metrics used to assess and
83 compare anthropogenic and natural drivers of climate change. The aerosol RE
84 represents the instantaneous radiative impact of atmospheric particles on the Earth's
85 energy balance [*Heald et al.*, 2014]. RF is calculated as the change of RE between
86 two different periods, e.g., the pre-industrial and the present-day times [*Heald et al.*,
87 2014; *Liu et al.*, 2007], based on the aerosol and precursor gas emissions in the two
88 periods [*Dentener et al.*, 2006; *Lamarque et al.*, 2010].

89 RF due to aerosol and radiation interactions (RFari) of biomass burning aerosols
90 has been estimated since the IPCC second Assessment Report (AR2). Based on the
91 Aerosol Comparisons between Observations and Models (AeroCom) Phase II
92 simulations [*Bond et al.*, 2013; *Myhre et al.*, 2013b], RFari of biomass burning
93 aerosols in the IPCC Fifth Assessment Report (AR5) is estimated to be 0.0 W m^2
94 (ranging from -0.20 to 0.20 W m^{-2}), and RFari of biomass burning black carbon (BC)
95 and primary organic matter (POM) are of the opposite sign (i.e., 0.10 and -0.10 W m^{-2},
96 respectively).

97 There are also some studies that estimated the RE due to aerosol and radiation

98 interactions (REari) of fire aerosols by comparing the simulation with fire emissions
99 against the simulation with no fire emissions. For example, using the NCAR
100 Community Atmosphere Model version 4 (CAM4) with a bulk aerosol module, *Tosca*
101 *et al.* [2013] reported that the top-of-atmosphere (TOA) REari from global biomass
102 burning aerosols is 0.18±0.10 W m^{-2} averaged for the period of 1997-2009. *Ward et al.*
103 [2012] estimated the REari from biomass burning aerosols in the pre-industrial (for
104 the year 1850), present-day (for the year 2000), and future time periods (for the year
105 2100), and found that the biomass burning aerosol REari for the year 2000 is 0.13 W
106 m^{-2} and -0.27 W m^{-2} in all-sky and clear-sky conditions, respectively.
107 RE due to aerosol and cloud interactions (REaci) of biomass burning aerosols can
108 be comparable in magnitude to or even stronger than the REari [*Liu et al.*, 2014].
109 With a global aerosol-climate model, the REaci of biomass burning aerosols was
110 estimated to range from -1.74 to -1.00 W m^{-2} for the year 2000 in *Ward et al.* [2012].
111 The semi-direct radiative effect of biomass burning aerosols is not independently
112 assessed in IPCC reports. The magnitude was reported to be about 7.0 W m^{-2} in the
113 Southern American biomass burning regions by examining the radiative flux
114 difference with and without the biomass burning aerosol effect on clouds [*Liu*, 2005].
115 The RF or RE due to surface albedo changes (RFsac or REsac) of BC from open
116 fires and other sources has been estimated in previous studies. For biomass burning
117 emissions with a strong (1998) and weak (2001) boreal fire year, RE of fire
118 BC-in-snow was estimated to be 0.011 and 0.006 W m^{-2}, respectively [*Flanner et al.*,
119 2007]. *Randerson et al.* [2006] reported that BC from a boreal forest fire deposited on

120 snow and sea ice introduced a global annual mean RE of 8±5 W per m^2 of burned area
121 in the first year when the fire happened. A summary of BC-in-snow forcing/effect can
122 be found in *Bond et al*. [2013]. They reported that the present-day RE of fire
123 BC-in-snow ranges from 0.006 to 0.02 W m^{-2} based on previous studies [*Jacobson*,
124 2004; *Rypdal et al.*, 2009; *Skeie et al.*, 2011; *Hansen et al.*, 2005; *Flanner et al.*, 2007,
125 2009; *Koch et al.*, 2009].

126 Biomass burning aerosols can have significant impacts on global and regional
127 precipitation and atmospheric circulation. With the change of fire emissions from year
128 1860 to 2000, *Jones et al.* [2007] found that biomass burning aerosols decrease the
129 global near-surface air temperature by about 0.25°C, when considering the feedbacks
130 of sea surface temperature (SST) in the model. As shown in *Tosca et al.* [2013], the
131 direct and semi-direct effects of biomass burning aerosols reduce the precipitation
132 near the equator and weaken the Hadley circulation. With a regional climate model,
133 *Zhang et al.* [2009] found that biomass burning aerosols may warm and stabilize the
134 lower troposphere and thus reinforce the dry season rainfall pattern in the Southern
135 Amazonia. The absorption of shortwave radiation by biomass burning BC could
136 increase the vertical stratification and inhibit both the cloud formation and
137 precipitation [*Ackerman et al.*, 2000; *Tosca et al.*, 2014]. In contrast, biomass burning
138 aerosols could invigorate the convective clouds [*Andreae et al.*, 2004a; *Koren et al.*,
139 2005] through suppressing warm rain processes in the convection, and enhance the
140 latent heat release at higher levels [*Andreae and Rosenfeld*, 2008].

141 Although there have been many studies quantifying the RE of fire aerosols, a

142 further investigation is still needed, as the current estimations of fire aerosol RE are
143 still associated with large uncertainties [e.g., *Myhre and Samset*, 2015; *Chakrabarty et*
144 *al.*, 2014]. The REs of co-emitted fire POM versus BC are even less clear. In this
145 study, we estimate the present day (from year 2003 to 2011) open fire aerosol REs
146 (including REari, REaci and REsac) using the NCAR Community Atmosphere Model
147 version 5.3 (CAM5) with the four-mode version of the modal aerosol module
148 (MAM4). We use two methods to calculate the REari of fire aerosols (total, BC-only,
149 and POM-only). One method estimates the REari based on different model
150 simulations [*Ghan*, 2013], and the other one calculates the REari directly through
151 multiple diagnostic radiation calls in a single simulation. The spatial and seasonal
152 characteristics of fire aerosol REs, and the impacts on the global precipitation and
153 temperature are discussed.

154 Compared to earlier studies of fire aerosol REs [*Tosca et al.*, 2013; *Ward et al.*,
155 2012], a number of improvements are made in this study. First, a higher model
156 horizontal resolution at 0.9° by 1.25° is used versus 1.9° by 2.5°. The higher
157 resolution allows more efficient transport of aerosols from the sources to remote
158 regions [*Ma et al*., 2013; 2014]. Model resolution has also been shown to be important
159 for aerosol REaci [*Ma et al*., 2015]. Second, the latest CAM5 model with MAM4 is
160 used. MAM4 with an additional primary carbon mode explicitly treats the
161 microphysical ageing of primary carbonaceous aerosols (POM/BC) in the atmosphere.
162 MAM4 has higher BC and POM burdens over the earlier three-mode version of
163 MAM (MAM3) in the remote regions by ~30% [*Liu et al.*, 2016]. Third, daily instead

of monthly fire emissions are used, which allows the model to consider the effect of fast changes in the fire emission flux on local atmospheric conditions. It is expected that using the monthly mean emission flux the model can not consider the effect of extremely strong fires, thus it might underestimate the fire aerosol REs for such cases. Finally, a new methodology [*Ghan*, 2013] is used to more accurately diagnose the REs of fire aerosols. Central to this method is that the REari must be calculated in the presence of clouds (i.e., under the all-sky condition), and the REaci be calculated under the condition of no aerosol effects on radiation. With the radiative forcing decomposition of this method, REsac can also be quantified.

The paper is organized as follows. Section 2 introduces the model and experiments. Section 3 describes the methods to diagnose the fire aerosol REs. Section 4 presents the model results of fire aerosol REs, and impacts on global and regional surface temperature and precipitation. Conclusions and discussion are given in Section 5.

2. Model, Experiment Design and Aerosol Radiative Effect Method

2.1 Model

In our study, we use the Community Earth System Model (CESM) version 1.2, with the Community Atmosphere Model version 5.3 (CAM5.3) [*Neale et al.*, 2010] coupled with the Community Land Model version 4 (CLM4) [*Oleson et al.*, 2010]. The SNow, ICe, and Aerosol Radiative model (SNICAR) [*Flanner and Zender*, 2005] is turned on in the simulations to diagnose the biomass burning BC-in-snow effect. CAM5 includes several major updates in its physics parameterizations compared to

187 previous CAM versions. A two-moment stratiform cloud microphysics scheme is
188 included in CAM5 to predict both the mass and number mixing ratios of cloud liquid
189 and cloud ice [*Morrison and Gettelman*, 2008]. MAM4, which was updated from
190 MAM3 [*Liu et al.*, 2012], includes aerosol mass and number mixing ratios in four
191 lognormal modes: Aitken, accumulation, coarse, and primary carbon mode [*Liu et al.*,
192 2016]. An additional primary carbon mode is included in MAM4 on the top of
193 MAM3 to explicitly treat the microphysical ageing of primary carbonaceous aerosols
194 (POM and BC) in the atmosphere. POM and BC in MAM4 are emitted in the primary
195 carbon mode instead of directly in the accumulation mode as in MAM3. MAM4
196 significantly increases the BC and POM concentrations in the remote regions (e.g.,
197 over oceans and Arctic) due to reduced wet scavenging of POM and BC in the
198 primary carbon mode with a lower hygroscopicity than that in the accumulation mode.
199 The increase is relatively small in the land source regions [*Liu et al.*, 2016].
200
201 2.2 Experiment design
202 CAM5 was run with the finite volume dynamics core in a resolution of 0.9°
203 latitude by 1.25° longitude and 30 vertical levels. The model was run for the time
204 period of year 2003 to 2011 (i.e., for 9 years) with prescribed monthly SST and sea
205 ice. The year 2003 was run twice and the first year simulation was used as a model
206 spin-up. Global Fire Emissions Database version 3.1 (GFED 3.1) daily emissions
207 [*Giglio et al.*, 2013] for BC, POM and sulfur dioxide (SO_2) from 2003 to 2011 are
208 prescribed, and the vertical distribution of fire emissions is based on the AeroCom

protocol [*Dentener et al.*, 2006]. Anthropogenic aerosol and precursor gas emissions are from the IPCC AR5 dataset [*Lamarque et al.*, 2010]. We performed our control experiment (FIRE) with the GFED fire emissions turned on and a sensitivity experiment (NOFIRE) with the fire emissions turned off. Differences between FIRE and NOFIRE experiments are used to calculate the REs and atmospheric effects of biomass burning aerosols on temperature and precipitation. Two additional experiments (NOFIREBC and NOFIREPOM) were performed with fire BC and POM emissions turned off, respectively. Differences between the control (FIRE) and these two experiments represent the contribution from biomass burning BC and POM, respectively. Other forcings (e.g., SST, greenhouse gases) of all these experiments are kept the same. We performed ten ensemble members for each of these experiments. Furthermore, we performed the other experiment (FIRE_BBFFBF) using the modified CAM5 model that separately predicts the BC and POM from biomass burning (BB), fossil fuel (FF) and biofuel (BF) sources, while other model features are kept the same as the FIRE experiment. A summary of all the experiments in this study can be found in Table 1.

2.3 Methods of calculating fire aerosol radiative effects

The REs of all fire aerosols, fire BC, and fire POM are calculated from the differences of TOA shortwave fluxes (ΔF) between the FIRE experiment and the three other experiments (NOFIRE, NOFIREBC and NOFIREPOM), respectively. All the atmospheric variables (including temperature, precipitation, and circulation) are

allowed to adjust in the experiments. However, with SST and sea ice prescribed in these experiments, only the rapid adjustments are taken into account. Thus the *effective* radiative effects are actually calculated in this study.

$$\Delta F_{\text{fire aero}} = F_{\text{fire}} - F_{\text{nofire}} \quad (1)$$

$$\Delta F_{\text{fire bc}} = F_{\text{fire}} - F_{\text{nofirebc}} \quad (2)$$

$$\Delta F_{\text{fire pom}} = F_{\text{fire}} - F_{\text{nofirepom}} \quad (3)$$

The total TOA shortwave flux change can be broken into the REari, REaci, and REsac. The aerosol REaci results from both the aerosol effect on clouds via acting as CCN and the aerosol semi-direct effect on clouds via affecting the atmospheric states due to absorbing aerosols. We adopt the method of *Ghan* [2013] to separate the REari, REaci, and REsac from the total effects of all fire aerosols, fire BC and fire POM, respectively. The method is summarized as follows. F_{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. $F_{\text{clean,clear}}$ 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.

$$\Delta F = \underbrace{\Delta(F - F_{\text{clean}})}_{\text{(REari)}} + \underbrace{\Delta(F_{\text{clean}} - F_{\text{clean,clear}})}_{\text{(REaci)}} + \underbrace{\Delta F_{\text{clean,clear}}}_{\text{(REsac)}} \quad (4)$$

In the method above, REaci includes both aerosol indirect and semi-direct effects. The fire BC has a much weaker indirect effect due to its lower mass burden and lower hygroscopicity compared to fire POM [*Koch et al.*, 2011]. Thus the fire aerosol semi-direct effect can be approximately represented by the REaci of fire BC. The fire

aerosol indirect effect can be estimated as the difference of fire aerosol REaci and semi-direct effect. With the sea ice prescribed in these experiments, the radiative effect of fire aerosols on sea ice albedo is not considered in REsac.

We undertake another method to estimate the fire aerosol REari from the experiment (FIRE_BBFFBF). With explicit predictions of fire POM and fire BC in FIRE_BBFFBF, the REari of fire BC and fire POM are estimated by two diagnostic radiation calls, each time neglecting the scattering and absorption of solar radiation of fire BC and fire POM, respectively. This more direct method is named as BBFFBF, and the REari of fire BC and fire POM will be compared with those from the method of *Ghan* [2013]. The fire BC-in-snow effect is calculated from SNICAR, and compared with the REsac estimated from *Ghan* [2013].

3. Results

3.1 Simulation of biomass burning aerosols

The biomass burning BC and POM from forest, grass and agriculture fires are significant contributors to the total BC and POM emissions. Figure 1 shows the seasonal variation of GFED fire emissions (including forest, grass and agriculture fires) in the global, tropical (25°S to 25°N), and Arctic (60°N to 90°N) regions. Global fire emission is the largest during the boreal summer as well as in the boreal autumn (September/October), when it is the fire season in the tropical regions of the Southern Hemisphere (SH). The tropical fire emission contributes the most to the annual global fire emission (80% for BC and 85% for OC, respectively), compared to

other regions. Arctic is the other important fire region, where the emission maximum is found during the boreal summer. In the boreal summer, the OC emission in the Arctic regions is about 50% of that in the tropical region. The BC emission in the Arctic is much smaller than that of the tropical regions even in the boreal summer fire season. The dominant fire type in the SH tropics is deforestation, savanna and grassland fires, while that in the Arctic is the forest fires. The OC to BC ratio (OC/BC) of forest fires is almost three times higher than that of deforestation, savanna and grassland fires [*van der Werf et al.*, 2010].

Figure S1 in the supplemental materials shows the latitudinal and longitudinal distributions of vertically integrated concentrations (column burdens) of BC and POM from BB, FF, and BF sources based on the FIRE_BBFFBF experiment. The BC and POM from BB source are mainly distributed in the tropical and sub-tropical regions (southern Africa, South America and Southeast Asia) and in the mid- to high latitudes (North of 45°N) of the Northern Hemisphere (NH) (Northeast Asia, Alaska and Canada). The largest column burdens of biomass burning aerosols are located in southern Africa and adjacent oceanic areas (1.5 and 20 mg m^{-2} for BC and POM, respectively). The biomass burning aerosols are important aerosol species in the Arctic regions, and contribute up to 53% and 86% to the total burden of BC and POM in the Arctic (from 60° N to 90°N), respectively. In comparison, the maximum column burdens of fossil fuel BC and POM are found in East Asia, South Asia, Western Europe and North America. The maximum column burdens of biofuel BC and POM occur in East Asia, South Asia and Central Africa. The biofuel and fossil

297 fuel sources are dominant contributors to BC and POM in East Asia and South Asia.
298 In other regions of the world, biomass burning is the primary source of BC and POM.
299 Globally, the biomass burning contributes 41% and 70% to the total burdens of BC
300 and POM, respectively. Biomass burning can also emit SO_2. However, it only
301 contributes ~3% to the total global sulfate burden (figure not shown), so only
302 radiative effects of biomass burning POM and BC are discussed in this study.
303 The simulated aerosol optical depth (AOD) and single scattering albedo (SSA)
304 (including aerosols from all sources) are validated with observations from the
305 AErosol RObotic NETwork (AERONET, http://aeronet.gsfc.nasa.gov) at sites
306 significantly affected by biomass burning activity in southern Africa, South America
307 and the Arctic regions, as shown in Figures 2 and 3 (see Figure S2 in the
308 supplemental materials for the site locations). The AERONET AOD and SSA data are
309 averaged for the years from 2003 to 2011 to match the simulation period, although
310 there are missing AERONET data for some periods. We note that *Tosca et al.* [2013]
311 and *Ward et al.* [2012] applied scaling factors (from 1 to 3 varying by regions) to fire
312 emissions to improve modeled AOD magnitudes, whereas here we do not apply any
313 such scaling. In southern Africa, modeled monthly AOD agrees with observations
314 within a factor of 2 for the three sites (Figure 2a-2c). The underestimation of AOD is
315 found in the tropical site (Mongu) (Figure 2a) during the boreal autumn (the fire
316 season). The simulated AOD in the two other sites (Skukuza and Ascension Island) is
317 generally consistent with observations in both the magnitude and seasonal trend. The
318 simulated SSA in southern Africa ranges between 0.75 and 0.95 and generally

matches the observed SSA magnitude and seasonal cycle in the two land sites (Mongu and Skukuza) (Figure 3a-3b). However, an overestimation of SSA is found in the oceanic site (Ascension Island) (Figure 3c). The reason for this overestimation of SSA and thus the underestimation of absorption AOD (AAOD) is unclear and could be due to that the model has not treated the absorption enhancement of aged fire BC during its transport.

The simulated AOD in South America is generally consistent with observations within a factor of 2 (Figure 2d-2f). The seasonal variation of simulated AOD generally matches the observations. The underestimation of AOD in Alta Floresta and Cuiaba-Miranda is most obvious in September and October (the fire season), which may be attributed to the underestimation of fire emissions. However, the modeled AOD is higher than observations before the fire season for Alta Floresta and Rio Branco, which could be due to the overestimation of fire emission in this period. The simulated SSA in South America ranges mostly between 0.87–0.95 and matches the observations reasonably well (Figure 3d-3f). The modeled SSA is too low during the fire season and exhibits too strong a seasonality. It implies that the model underestimation of scattering aerosols (e.g., POM) may be more severe than that of BC during the fire season.

In the Arctic, small AOD (less than 0.3) and large SSA (larger than 0.9) are observed for the three sites. The observed large SSA in the fire season (boreal summer) is consistent with the high OC/BC ratio of fire emissions in the Arctic (Figure 1). The model significantly underestimates the observed AOD in the Arctic in

both fire and non-fire seasons. The underestimation of AOD can be due to (1) the underestimation 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], (2) the excessive scavenging of aerosols during their transport from the NH mid-latitude industrial regions by liquid-phase clouds [*Wang et al.*, 2013a], and (3) the coarse horizontal resolution (~100 km) of the model [*Ma et al.*, 2014]. Although MAM4 increases the column burdens of POM and BC by up to 40 % in many remote regions compared to MAM3, it still underestimates the surface BC concentrations in the Arctic [*Liu et al.*, 2016]. The modeled SSA in the Arctic is lower than observations, which implies that the simulation of AAOD is better than that of AOD and the underestimation of non-absorbing aerosols (e.g., sulfate and POM) in the Arctic may be more severe than that of BC.

3.2 Radiative effect due to aerosol-radiation interactions

The annual mean REari of all fire aerosols (including BC, POM and sulfate) estimated with the method of BBFFBF and with the method of *Ghan* [2013] is shown in Figure 4a-4b. The fire sulfate is not included in the calculation of REari of all fire aerosols with the method of BBFFBF. Its effect is minor since the global annual mean burden of fire sulfate (0.09 mg m^{-2}) is much smaller than that of fire POM (1.25 mg m^{-2}), both of which are light-scattering. The statistical significance of REari estimated with the *Ghan* [2013] method over the interannual variability and ensemble member diversity is shown in Figure 4 (and also later figures). The REari of all fire aerosols

from the two methods agree with each other very well. Thus, we will report the REari of all fire aerosols with the *Ghan* [2013] method below. The global annual mean REari of all fire aerosols is positive (0.16 ± 0.01 W m^{-2}), which indicates a warming effect from all fire aerosols. The REari is positive on the globe except in some land areas (e.g., southern Africa, South America, Great Lakes, North Canada, and East Siberia). The maximum positive REari is located in ocean areas west of southern Africa (~5.0 W m^{-2}) and South America (~1.5 W m^{-2}). Positive REari up to 1 W m^{-2} is found in the Arctic (60°N to 90°N). The different signs of REari between land and ocean areas of southern Africa and South America result from the differences in cloud fraction and cloud liquid water path (LWP) between land and ocean regions. In the fire season (August-September-October) of the SH tropical regions, cloud fraction and cloud LWP over the land areas (10% and 20 g m^{-2}, respectively) are much smaller than those over the adjacent ocean areas (70% and 70 g m^{-2}, respectively). The biomass burning aerosols are transported above the low-level stratocumulus clouds, and when biomass burning BC resides above clouds, its absorption of solar radiation is significantly enhanced due to the reflection of solar radiation by underlying clouds [*Abel et al.*, 2005; *Zhang et al.*, 2016].

A comparison of modeled REari in the boreal autumn (September-October-November) over the South Atlantic Ocean with satellite observations is shown in Figure 5. The observed above-cloud aerosol REari is calculated with the method of *Zhang et al.* [2014] using the Aqua/MODIS and Terra/MODIS products, respectively. The observed above-cloud aerosol REari over

385 southeastern Atlantic Ocean is 3-12 W m^{-2}, with higher values near the coasts. The
386 simulated REari agrees better with Aqua/MODIS observed REari than with
387 Terra/MODIS in both the magnitude and spatial pattern. REari estimated from
388 Terra/MODIS (morning time) is stronger than the one estimated from Aqua/MODIS
389 (afternoon time) due to the larger amount of underlying clouds in the morning [*Min*
390 *and Zhang*, 2014]. Over South America during the fire season (August to September),
391 the clear-sky fire aerosol REari is estimated to be -5.2 W m^{-2} by *Sena and Artaxo*
392 [2015], which is larger than our model result (-2.1 W m^{-2}). This is consistent with the
393 underestimation of modeled AOD in South America compared to the AERONET data
394 (Figure 2).

395 The seasonal variation of REari of all fire aerosols with the *Ghan* [2013] method
396 is shown in the supplemental Figure S3. The REari has a maximum (1.13 W m^{-2}) in
397 the boreal summer (June-July-August, JJA) over the Arctic regions, partially due to
398 the low solar zenith angles there. The maximum positive REari in the tropical regions
399 occurs in the boreal summer and autumn (September, October and November, SON)
400 during the fire season of southern Africa and South America. The REari reaches a
401 positive maximum in Southeast Asia during the fire season in March, April and May
402 (MAM).

403 The REari of fire BC is shown in Figure 4c-4d. The fire BC REari calculated
404 from the two methods are similar in magnitudes and spatial patterns, and there is
405 much less noise with the BBFFBF method. The global annual mean fire BC REari is
406 about 0.25 ± 0.01 W m^{-2} and positive over the globe (the regions with negative values

407 in Figure 4d are in general not statistically significant). Unlike all fire aerosols, fire
408 BC generates a positive forcing in the land regions of southern Africa and South
409 America, and the amplification effect of low-level clouds on fire BC positive forcing
410 can be clearly seen in southern Africa and adjacent Atlantic Ocean.

411 The global annual mean REari of fire POM from the two methods somewhat
412 differs from each other (Figure 4e-4f). The BBFFBF method gives a small negative
413 value (-0.05 W m^{-2}), while the *Ghan* [2013] method shows a small positive value
414 (0.04 ± 0.01 W m^{-2}). The difference is mainly in the Arctic regions where the positive
415 forcing from *Ghan* [2013] is larger than that from the BBFFBF method. This is
416 because the removal of fire POM emissions in the NOFIREPOM experiment affects
417 the properties of aerosol particles within which co-emitted fire BC is internally mixed
418 with fire POM, causing a decrease of BC burden in the Arctic (by ~0.05 mg m^{-2})
419 compared to the FIRE experiment. Thus, one should be careful in using the *Ghan*
420 [2013] method to diagnose the radiative forcing of a single component within
421 co-emitted aerosols. The REari of fire POM is negative in most of the globe. However,
422 positive forcing can be found over oceanic regions west of southern Africa and South
423 America, North Pacific Ocean and the Polar regions where large amount of low-level
424 clouds, sea ice or land ice exist. The multiple scatterings between the above-cloud fire
425 POM and low-level clouds or between the fire POM and the Earth's bright surface
426 with high albedos could reduce the amount of solar radiation reflected by these
427 low-level clouds and bright surface in the case without the fire POM [*Zhang et al.*,
428 2016]. With the BBFFBF method the sum of REari from fire POM and fire BC (i.e.,

0.20 W m^{-2}) is larger than that of all fire aerosols (0.15 W m^{-2}). It reflects the nonlinear interactions among different aerosol components [*Ghan* et al., 2012]. The nonlinearity is stronger with the *Ghan* [2013] method.

3.3 Radiative effect due to aerosol-cloud interactions

The annual mean REaci due to all fire aerosols, fire BC, and fire POM are shown in Figure 6. The REaci diagnosed with the *Ghan* [2013] method includes both aerosol indirect and semi-direct effects. The fire aerosol semi-direct effect (to be discussed below) is much smaller (-0.04±0.03 W m^{-2} on the global mean) than the indirect effect, and the REaci is mostly from the fire aerosol indirect effect. The global annual mean REaci of all fire aerosols is -0.70±0.05 W m^{-2} (Figure 6a). In the tropical regions, the strong negative REaci is located in the adjacent ocean areas of southern Africa, South America and Australia, with the maximum REaci of -8.0 W m^{-2} over the South Atlantic Ocean. The strong negative REaci also occurs in the Arctic (60°N to 90°N). The REaci in East Siberia, Alaska and Canada is as large as -6.0 W m^{-2}.

The fire BC has a weak indirect effect by acting as CCN, but can reduce the cloud amount through its semi-direct effect. The REaci of fire BC (Figure 6b) can approximate the fire BC semi-direct effect with a small global annual mean value of -0.04±0.03 W m^{-2}. However, stronger positive effect can be found in the western Pacific (3.0 W m^{-2}) and Arctic regions (1.0 W m^{-2}). The global annual mean REaci of fire POM is -0.59±0.03 W m^{-2} (Figure 6c), and dominates the cloud effect of all fire aerosols. The sum of REaci from fire BC and POM (-0.62±0.03 W m^{-2}) is smaller

than that of all fire aerosols (-0.70±0.05 W m^{-2}) due to the non-linear interactions of fire BC and fire POM [*Jiang et al.*, 2013] as well as the negative REaci of fire sulfate.

The seasonal variation of all fire aerosol REaci is shown in Figure 7. The maximum of fire aerosol REaci is in the boreal summer (i.e., the fire season in NH) located in the NH high latitudes (60°N to 90°N). The largest summer REaci is found in the land areas and is as large as -15 W m^{-2}. The fire aerosol REaci in the tropical regions is most significant in the boreal summer (up to -15 W m^{-2}) and autumn (up to -10 W m^{-2}) over the ocean areas. The different spatial distributions of fire aerosol REaci in the NH high latitudes and in the tropics result from the difference in cloud distributions between the two regions. During the fire season the cloud LWP over the land areas in the NH mid- and high latitudes is three times larger than that over the ocean areas in the tropics. Larger cloud LWP favors the stronger REaci, because the larger LWP associated with the warm cloud and rain processes favors the aerosol indirect effect via slowing down the autoconversion of cloud water to rain [*Ghan et al.*, 2012; *Jiang et al.*, 2015]. Meanwhile, in the NH high latitudes, the lower solar zenith angle in the boreal summer favors the stronger REaci. Like the fire aerosol REari, the smallest fire aerosol REaci occurs in the boreal spring.

Seasonal variations of zonal mean fire aerosol REari, REaci, cloud LWP, low-level (from surface to 750 hPa) cloud amount, and vertically-integrated (burden) concentrations of fire POM and fire BC are shown in Figure 8. The seasonal variation of fire BC and fire POM burdens is largest in the SH low latitudes (from 30°S to 0°N) and NH mid- and high latitudes (50°N to 90°N). Distinct features of these two areas

473 can also be noticed that the maximum fire BC burden in NH (0.3 mg m^{-2}) is much
474 lower than that in SH (0.8 mg m^{-2}), while the maximum POM burdens in these two
475 areas are comparable. Interestingly, the REari is larger in the boreal summer in NH
476 than that in the boreal autumn in SH, although the fire BC burden is much lower in
477 the NH summer. It is mainly due to the larger amount of low clouds in the NH high
478 latitudes, which enhances the absorption of fire BC. The maximum REari in the NH
479 summer is found near the North Pole (70 °N to 90 °N), and not around 60 °N where
480 the fire aerosol burden is highest. The REaci of fire aerosols is about 3 times larger in
481 the boreal summer in NH than that in the boreal autumn in SH, although the burden of
482 fire POM in NH is comparable to that in SH. The larger cloud LWP in the NH
483 summer around 40-70°N favors the stronger REaci there.
484
485 3.4 Surface albedo effect
486 Here we compare the modeled BC-in-snow (BCS) concentrations with
487 observation data collected from multiple field campaigns over the Arctic [*Doherty et*
488 *al.*, 2010] and Northern China [*Wang et al.*, 2013b; *Qian et al.*, 2014]. Figure 9a
489 shows the simulated (from FIRE and NOFIRE experiments) and observed BCS
490 concentrations as a function of latitude. The range of observed BCS concentrations is
491 between 1 and 200 ng g^{-1} in the Arctic and between 50 and 2000 ng g^{-1} in Northern
492 China, respectively. Both FIRE and NOFIRE experiments capture the meridional
493 gradient in BCS concentrations between the mid-latitudes (Northern China) and high
494 latitudes (Arctic). The mean and median concentrations of BCS are both

495 overestimated in Northern China, implying the high biases from the anthropogenic
496 emissions and/or model physics (Figure 9b). The mean and median BCS
497 concentrations from the FIRE experiment agree slightly better with observations than
498 those from the NOFIRE experiment in the Arctic (Figure 9b). This suggests that fire
499 emissions are important for BCS concentrations in the Arctic.

500 The annual mean REsac of all fire aerosols estimated with *Ghan* [2013] and the
501 fire BCS effect diagnosed from SNICAR are shown in Figure 10a. We note that the
502 radiative effect due to BC deposition on sea ice is not considered since sea ice is
503 prescribed in the simulations. The global annual mean REsac (0.03 ± 0.10 W m^{-2}) is
504 much smaller compared to the REari and REaci. The REsac over land is maximum in
505 spring (0.12 ± 0.27 W m^{-2}) and winter (0.06 ± 0.16 W m^{-2}). The REsac over land in
506 summer and autumn is very small (less than 0.01 W m^{-2}). We note that the mean
507 REsac calculated with *Ghan* [2013] is much smaller than the standard deviation
508 resulted from the internal variability.

509 The annual mean fire BCS effect calculated from SNICAR is shown in Figure
510 10b and 10c. The spatial distribution of the fire BCS effect is similar to the fire REsac,
511 implying that the fire REsac has a significant contribution from the fire BCS effect.
512 Averaged when only snow is present, the fire BCS effect is larger (0.048 W m^{-2}). The
513 global mean fire BCS effect (with the presence of snow) can be as large as 0.06 W m^{-2}
514 in spring. The maximum fire BCS effect (up to 1 W m^{-2}) is located in Greenland and
515 the very northern reaches of Canada, while that in the other Arctic regions and North
516 China is smaller.

517 The positive REsac in Siberia, North America and Canada can be a result of BCS
518 effect. However, the REsac in these regions is larger than the BCS effect especially in
519 spring. The snow melting and snow depth change due to the BCS warming may
520 induce a larger positive REsac than the albedo change due to BCS itself. The negative
521 REsac over land can be a result of atmospheric feedbacks caused by fire aerosols
522 [*Ghan*, 2013].
523
524 3.5 Fire aerosol effects on shortwave radiation, global temperature and precipitation
525 Here, we show the annual mean net shortwave flux change at TOA (i.e., total
526 radiative effect), in the atmosphere and at surface, and changes in surface air
527 temperature, convective and large-scale precipitation due to all fire aerosols in Figure
528 11 and Table 2. The global mean net shortwave flux change at TOA due to all fire
529 aerosols is -0.55±0.07 W m^{-2}, which indicates that fire aerosols lead to the reduction
530 of shortwave flux into the Earth's system. The zonal mean TOA shortwave flux
531 reduction in the Arctic regions (-1.35±1.03 W m^{-2}) is much larger than that in the
532 tropical regions (-0.66±0.09 W m^{-2}). The cooling at TOA is mostly from fire aerosol
533 REaci. The maximum negative RE is located in the land areas of the Arctic and ocean
534 areas of the tropics. Although the global mean total radiative effect is negative,
535 positive effect is found in some land areas (e.g., Africa, Greenland).
536 The shortwave atmospheric absorption change in the tropical regions is larger
537 than that in the Arctic regions. It is because BC burden in the tropics (0.17 mg m^{-2}) is
538 larger than that in the Arctic (0.09 mg m^{-2}). Strong absorption (~8 W m^{-2}) in the

atmosphere is found in the land areas of southern Africa and South America and in the Southeast Atlantic. The surface shortwave flux change in the Arctic is mostly from the TOA shortwave flux reduction due to the fire aerosol REaci, while the surface shortwave flux change in the tropics is mostly due to the fire BC absorption in the atmosphere.

The fire aerosols lead to the reduction of the global mean surface air temperature (T_s) by 0.03±0.03 K, consistent with the reduction of shortwave fluxes at TOA and at surface. The largest surface cooling is found in the Arctic and tropical regions by up to 0.6 K. The cooling of the Arctic is related to the strong fire aerosol REaci, while the cooling in the tropics is mainly from the surface shortwave flux reduction due to the fire BC absorption. The T_s change in the ocean areas is very small since the SST is prescribed in our simulations.

The global mean total precipitation is reduced by 0.010±0.002 mm day^{-1} due to all fire aerosols (Table 2). Unlike the T_s change, the precipitation reduction in the tropics (0.016±0.01 mm day^{-1}) is much larger than that in the Arctic (0.001±0.02 mm day^{-1}, not statistically significant). The reduction in the tropics is mainly from the large-scale precipitation decrease (0.015±0.003 mm day^{-1}). The net decrease in the convective precipitation is very small in the tropics (0.001±0.009 mm day^{-1}, not statistically significant), as the convective precipitation is significantly decreased near the equator and increased in the regions away from the equator, partly consistent with the results of *Tosca et al.* [2013]. The precipitation reduction in southern Africa is consistent with the recent findings of *Hodnebrog et al.* [2016]. The shortwave flux

561 reduction at surface leads to a stabilization of the atmospheric boundary layer and a
562 suppression of the convection near the equator. The strong atmospheric absorption by
563 fire BC leads to the reduction of low-level clouds and large-scale precipitation in the
564 tropics. Both effects lead to a significant reduction of total precipitation near the
565 equator. The precipitation decrease in the NH high latitudes is mainly from the
566 reduction of convective precipitation. We note that the temperature and (especially)
567 precipitation changes reported here do not represent the complete impact of fire
568 aerosols, since the SSTs are fixed in our simulations. Fully-coupled atmosphere and
569 ocean models will be used to further investigate the impact of fire aerosols.

570 Figure 12 shows the changes of T_s, total precipitation, cloud LWP, and low-level
571 cloud cover in the summer due to all fire aerosols. The T_s is reduced by more than 1 K
572 in most of land areas around 60°N. The maximum cooling (larger than 1.5 K) is found
573 in East Siberia, Alaska and Canada. A decrease of total precipitation (by about 0.2
574 mm day^{-1}) is found in these regions. Accompanying the surface cooling and
575 precipitation reduction, a significant increase of cloud LWP and low-level cloud cover
576 is found there. This is a result of the indirect effect of fire aerosols in the land areas of
577 the Arctic (60°N to 90°N). The fire POM leads to the reduction of cloud droplet
578 effective radius and the increase of cloud droplet number concentration, consistent
579 with observed fire effects on clouds in Canada and the United States [*Peng et al.*,
580 2002].

581

582 **4. Discussion and Conclusions**

583 Although many studies have been conducted on the fire aerosol RE and RF [e.g.,
584 *Bond et al.*, 2013; *Myhre et al.*, 2013b; *Ward et al.*, 2012; *Tosca et al.*, 2013], the
585 current estimations are still associated with large uncertainties. In this study, the fire
586 aerosol RE (including REari, REaci and REsac) is calculated based on a new method
587 from *Ghan* [2013]. In addition, the fire aerosol REari and fire BC-in-snow effect are
588 diagnosed from an experiment of CESM which tracks the open fire BC and POM
589 separately from fossil fuel and biofuel sources and compared with the estimates from
590 the *Ghan* [2013] method.

591 The BC and POM burdens from open fires are largest in the tropical regions
592 (southern Africa, South America and Southeast Asia) and in the NH mid- to high
593 latitudes (North of 45°N) (Northeast Asia, Alaska and Canada). Fire aerosols
594 contribute 41% and 70% to the global burden of BC and POM, respectively. When
595 comparing with the AERONET AOD and SSA data, modeled monthly AOD agrees
596 with observations within a factor of 2 for most of the southern African and South
597 American sites. The model underestimation of AOD is found in the South American
598 sites near fire source regions, which is most obvious in the fire season (September and
599 October). The model underestimates the observed AOD in the Arctic regions in both
600 fire and non-fire seasons. The modeled SSA in southern Africa and South America is
601 generally in agreement with observations, while the modeled SSA in the Arctic is
602 lower.

603 The annual mean REari of all fire aerosols is 0.16 ± 0.01 W m^{-2} and positive over
604 most areas except in some land areas (e.g., southern Africa, North Canada, and East

605 Siberia). The annual maximum REari is found in the oceanic areas west of southern
606 Africa (5 W m^{-2}) and South America (1.5 W m^{-2}). The positive REari over the land
607 regions of southern Africa and South America is smaller, although the fire aerosol
608 burdens are higher. The annual zonal mean REari in the Arctic regions can reach 0.43
609 ±0.028 W m^{-2}, and is larger than that in the tropical regions (0.17±0.017 W m^{-2}),
610 although the fire aerosol burden is higher in the tropics. The annual mean REari of
611 fire BC is about 0.25±0.01 W m^{-2} and positive over the globe. Fire POM induces a
612 weak negative REari globally (-0.05 W m^{-2}) with the BBFFBF method and a small
613 positive value (0.04±0.01 W m^{-2}) with the *Ghan* [2013] method. The positive REari
614 of fire POM is found over oceanic areas west of southern Africa and South America,
615 North Pacific, and polar regions where the low-level cloud coverage is large or the
616 surface albedo is higher.

617 The global annual mean REaci of all fire aerosols is -0.70±0.05 W m^{-2} and the
618 maximum effect is located in the ocean areas west of southern Africa and South
619 America and land areas of the NH high latitudes. The maximum fire aerosol REaci
620 occurs in the NH high latitudes in the boreal summer, which results from the large
621 cloud LWP over the land areas and the low solar zenith angle. Associated with the
622 strong indirect effects of fire aerosols in the Arctic summer, significant surface
623 cooling, precipitation reduction, and low-level cloud cover increase are found in these
624 regions.

625 Modeled BCS concentrations from the FIRE experiment are evaluated against
626 observations in Northern China and in the Arctic, and generally agree with the

627 observations for the mean and median values in the Arctic regions. The high bias of
628 modeled BCS concentrations in Northern China may not result from the fire BC
629 because differences in BCS concentrations between FIRE and NOFIRE experiments
630 are very small in North China. The global annual mean REsac is 0.03 ± 0.10 W m^{-2}
631 (statistically insignificant) with the maximum effect in spring (0.12 W m^{-2}). The
632 REsac is mainly due to the effect of fire BC deposit on snow (0.02 W m^{-2}) diagnosed
633 from SNICAR with the maximum effect as large as 0.06 W m^{-2} (when snow is present)
634 in spring.

635 The fire aerosols reduce the global mean surface air temperature (T_s) by $0.03 \pm$
636 0.03 K and precipitation by 0.01 ± 0.002 mm day^{-1}. The maximum cooling (~1 K) due
637 to fire aerosols occurs around 60°N in summer, and a suppression of precipitation
638 (~0.1 mm day^{-1}) is also found there. The strong cooling is a result of the strong
639 indirect effects (-15 W m^{-2}) in the land areas of the Arctic regions (60°N to 90°N). A
640 significant reduction of precipitation in southern Africa is also noticed. We note that
641 these results are based on the simulations with fixed SSTs and may not represent the
642 full climate responses.

643 In our study, the global radiative effect of fire aerosols is estimated from
644 simulations performed with the 4-mode version Modal aerosol module (MAM4) [*Liu*
645 *et al.*, 2016], daily fire emissions with prescribed vertical emission profiles, and
646 higher model resolution (0.9° by 1.25°) compared to earlier modeling studies of fire
647 aerosols [*Tosca et al.*, 2013; *Ward et al.*, 2012]. In their studies, the GFED fire
648 aerosol emissions were increased by a factor of 1-3 depending on regions to match the

observed AOD. In our study, we do not apply the scaling factor to the fire aerosol emissions. Our global annual mean REari of fire aerosols (0.16 ± 0.01 W m^{-2}) is, however, close to 0.18 W m^{-2} in *Tosca et al.* [2013] and 0.13 W m^{-2} in *Ward et al.* [2012]. The similar fire aerosol REari from our study but with smaller fire emissions than these previous studies can result from (1) the use of MAM4 in our study which more realistically represents the external/internal mixing of BC with other soluble aerosol species; (2) the more accurate estimation of REari of fire aerosols in the presence of low-level clouds with the method of *Ghan* [2013]; and (3) the inclusion of vertical emissions of fire aerosols, which allows more efficient transport of fire aerosols from sources. The REaci due to fire aerosols in our study (-0.70 ± 0.05 W m^{-2}) is smaller than -1.64 W m^{-2} in *Ward et al.* [2012] due to the lower fire POM emissions used in this study compared to *Ward et al.* [2012].

We note that there are limitations and uncertainties with our study. The model still underestimates observed AODs (mostly within a factor of 2) at the sites predominantly influenced by biomass burning aerosols during the fire season, which implies that the fire aerosol radiative forcing can be stronger than estimated in this study. The RE estimates of fire POM and fire BC with the *Ghan* [2013] approach may not be accurate due to the internal mixing of co-emitted fire components (POM and BC). In our simulations, sea ice is prescribed, and thus the fire BC effect on sea ice albedo is not considered. The brown carbon component of POM [*Feng et al.*, 2013] is not treated in the current CESM model, which may result in an underestimation of atmospheric absorption of fire aerosols.

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Figure Captions

Figure 1. Seasonal variation of GFED monthly fire (a) organic carbon (OC) and (b) black carbon (BC) emissions (Tg C $month^{-1}$) averaged for the period of year 2003 to 2011 in the global, tropical (25°S to 25°N) and Arctic (60°N to 90°N) regions.

Figure 2. Comparison of modeled seasonal variations of aerosol optical depth (AOD) for the period of 2003-2011 with observations for the same period from the AERONET sites. The upper, middle, and bottom panels represent the sites in southern Africa, South America, and the Arctic, respectively.

Figure 3. Same as Figure 2, but for the comparison of single scattering albedo (SSA) at 550 nm.

Figure 4. Annual mean radiative effect due to aerosol-radiation interactions (REari) (W m^{-2}) averaged over the period of 2003-2011 due to (a) all fire aerosols, (c) fire BC, and (e) fire POM estimated with the method of BBFFBF (left panels), and with the method of Ghan (2013) ((b), (d), and (f) in the right panels). The plus signs in Figure 4(b), (d) and (f) denote the regions where the radiative effect estimated with Ghan [2013] is statistically significant at the 0.05 level.

Figure 5. (a) September-October-November (SON) mean fire aerosol radiative effect due to aerosol-radiation interactions (REari) (W m^{-2}) for the period of 2003-2011 over the Southeast Atlantic Ocean due to all fire aerosols. (b) and (c) are the same as (a), but for the above-cloud aerosol REari for the period of 2007-2011 estimated using Aqua/MODIS and Terra/MODIS products [*Zhang et al.*, 2014], respectively.

Figure 6. Annual mean radiative effect due to aerosol-cloud interactions (REaci) (W m^{-2}) averaged over the period of 2003-2011 due to (a) all fire aerosols, (b) fire BC, and (c) fire POM. The plus signs denote the regions where the radiative effect is statistically significant at the 0.1 level.

Figure 7. Seasonal variation of radiative effect of all fire aerosols due to aerosol-cloud interactions (REaci) (W m^{-2}) for the period of 2003-2011 for (a) December-January-February (DJF), (b) March-April-May (MAM), (c) June-July-August (JJA), and (d) September-October-November (SON). The plus signs denote the regions where the radiative effect is statistically significant at the 0.05 level.

Figure 8. Month-latitude cross sections of zonal mean and monthly (a) vertically-integrated concentrations (mg m^{-2}) of fire BC and (b) fire POM, (c) cloud liquid water path (LWP, in g m^{-2}), (d) low-level cloud cover (CLDLOW, in %), (e) radiative effect due to aerosol-radiation interactions (REari, in W m^{-2}), and (f) radiative effect due to aerosol-cloud interactions (REaci, in W m^{-2}) of all fire aerosols.

Figure 9. Evaluation of CAM5 simulated black carbon (BC) concentration for the period of 2003-2011 (in ng g^{-1}) in the top snow layer against observations in the Arctic [*Doherty et al.*, 2010] and Northern China [*Wang et al.*, 2013b]. The top snow layer ranges in thickness from 1 to 3 cm. Configuration of the two CAM5 simulations (FIRE and NOFIRE) is summarized in Table 1. Panel (a) shows the comparisons at different latitudes. The box and whisker plot in panel (b) shows the minimum and maximum value with the bar, the 25th and 75th percentiles with the box, the 50th percentile (i.e., median) by the bar within the box, and the mean value with the dot.

Figure 10. (a) Annual mean radiative effect due to surface albedo changes (REsac, W m^{-2}) averaged over the period of 2003-2011 of all fire aerosols over land regions, and annual mean surface effect of fire BC-in-snow calculated from SNICAR averaged (b) over all times and (c) only when snow is present. The plus signs in (a) denote the regions where the radiative effect is statistically significant at the 0.1 level.

Figure 11. Annual mean net shortwave flux changes (W m^{-2}) over the period of 2003-2011 (a) at top of the atmosphere, (b) in the atmosphere, (c) at surface, and changes of (d) surface air temperature (TS, K), (e) convective precipitation (mm d^{-1}), and (f) large-scale precipitation (mm d^{-1}) due to all fire aerosols. The plus signs denote the regions where the change is statistically significant at the 0.1 level.

Figure 12. Changes in (a) surface air temperature (K), (b) total precipitation (mm d^{-1}), (c) cloud liquid water path (g m^{-2}), and (d) low-level cloud cover (%) due to all fire aerosols in the boreal summer (JJA) averaged for the period of 2003-2011. The plus signs denote the regions where the change is statistically significant at the 0.1 level.

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

Experiment	Ensembles	Fire BC	Fire POM	Fire SO_2
FIRE	10	On	On	On
NOFIRE	10	Off	Off	Off
NOFIREBC	10	Off	On	On
NOFIREPOM	10	On	Off	On
FIRE_BBFFBF	1	On	On	On

Table 2. Global, tropics (25°S to 25°N) and Arctic (60°N to 90°N) annual mean fire aerosol (POM and BC) burdens, fire aerosol AOD, total fire aerosol radiative effect (RE) at TOA, radiative effects due to aerosol-radiation interactions (REari), due to aerosol-cloud interactions (REaci), and due to surface albedo changes (REsac), and changes in cloud liquid water path (LWP), low-level cloud cover, net solar fluxes at surface, shortwave atmospheric absorption, surface air temperature, and precipitation (total, convective, and large-scale) due to all fire aerosols. Standard deviations about the 10-ensemble means are included. The change shown in bold character is statistically significant at the 0.05 level.

	Global	Tropics (25°S to 25°N)	Arctic (60°N to 90°N)
Fire POM burden (mg m^{-2})	**1.25±0.01**	**1.87±0.01**	**1.70±0.08**
Fire BC burden (mg m^{-2})	**0.106±0.001**	**0.17±0.001**	**0.09±0.004**
Fire aerosol optical depth	**0.008±0.001**	**0.012±0.001**	**0.007±0.0004**
Total radiative effect (W m^{-2})	**-0.55±0.07**	**-0.66±0.09**	**-1.35±1.03**
Radiative effect due to ARI (W m^{-2})	**0.16±0.01**	**0.17±0.017**	**0.43±0.028**
Radiative effect due to ACI (W m^{-2})	**-0.70±0.05**	**-0.82±0.09**	**-1.38±0.23**
Radiative effect due to surface albedo changes (over land, W m^{-2})	0.03±0.10	-0.04±0.06	0.09±0.80
Cloud liquid water path (g m^{-2})	**1.62±0.01**	**1.95±0.13**	**2.59±0.25**
Low-level cloud cover (%)	0.012±0.06	**-0.055±0.05**	**0.46±0.45**
Net solar flux at surface (W m^{-2})	**-1.38±0.05**	**-1.91±0.12**	**-2.27±1.04**
Shortwave atmospheric absorption (W m^{-2})	**0.83±0.03**	**1.25±0.04**	**0.92±0.05**
Surface air temperature (K)	**-0.03±0.03**	**-0.024±0.011**	-0.15±0.20
Total precipitation rate (mm day^{-1})	**-0.010±0.002**	**-0.016±0.01**	-0.001±0.02
Convective precipitation rate (mm day^{-1})	**-0.003±0.002**	-0.001±0.009	**-0.005±0.003**
Large-scale precipitation rate (mm day^{-1})	**-0.007±0.002**	**-0.015±0.003**	0.004±0.019

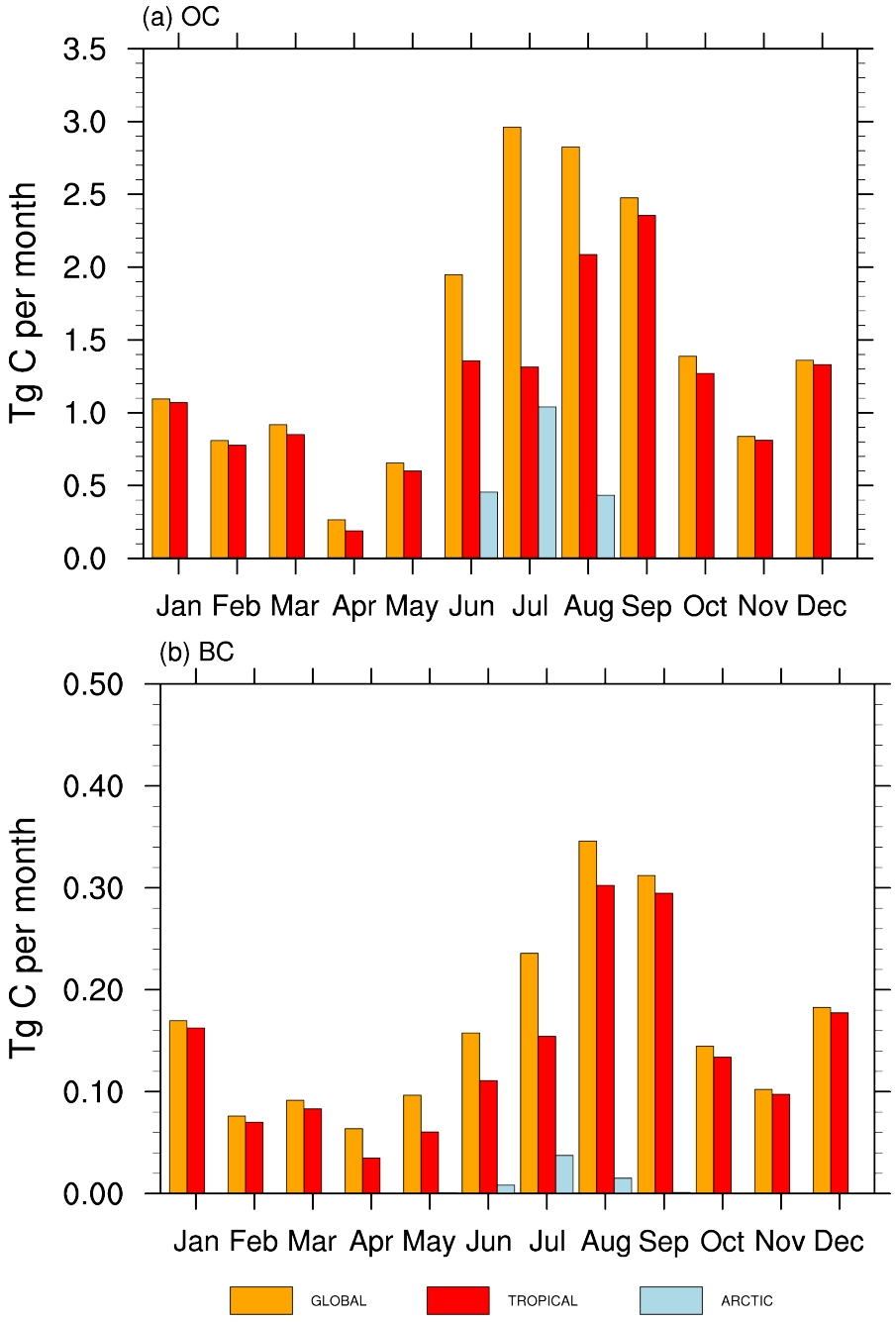

Figure 1. Seasonal variation of GFED monthly fire (a) organic carbon (OC) and (b) black carbon (BC) emissions (Tg C month^{-1}) averaged for the period of year 2003 to 2011 in the global, tropical (25°S to 25°N) and Arctic (60°N to 90°N) regions.

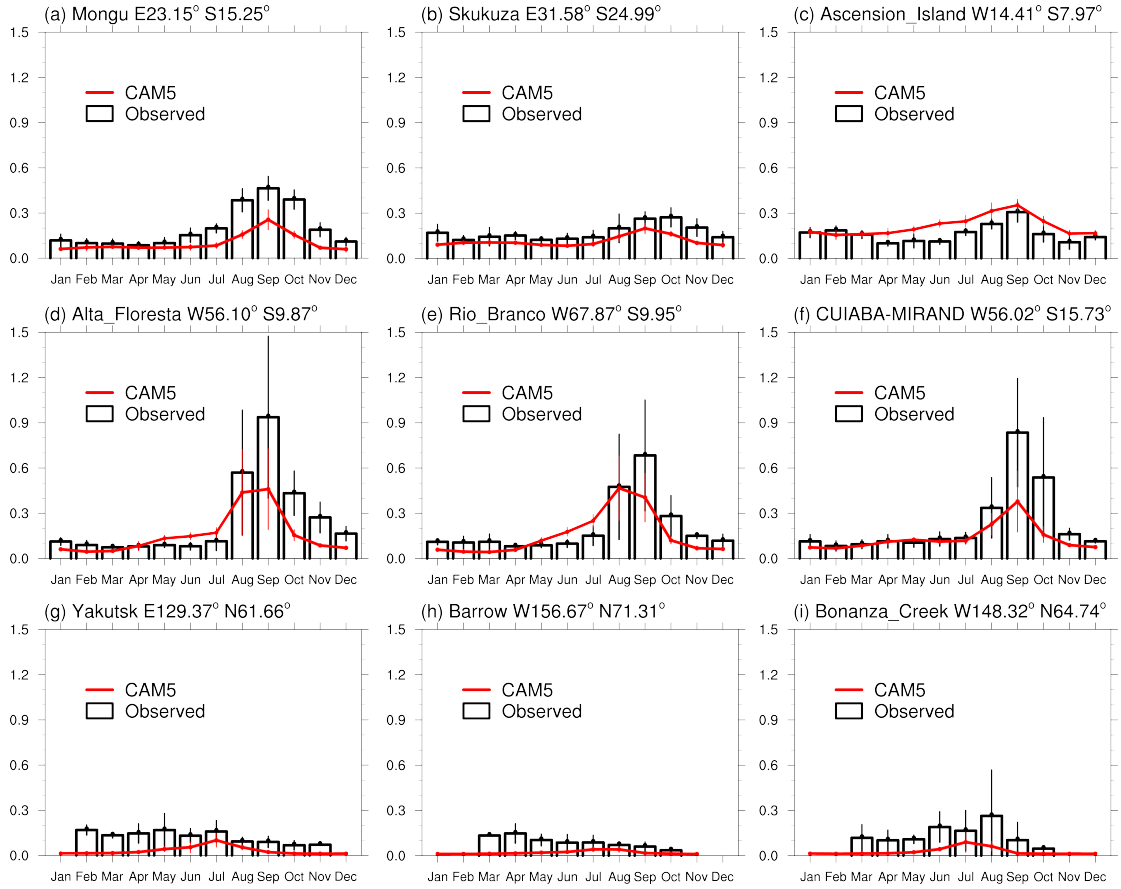

Figure 2. Comparison of modeled seasonal variations of aerosol optical depth (AOD) for the period of 2003-2011 with observations for the same period from the AERONET sites. The upper, middle, and bottom panels represent the sites in southern Africa, South America, and the Arctic, respectively.

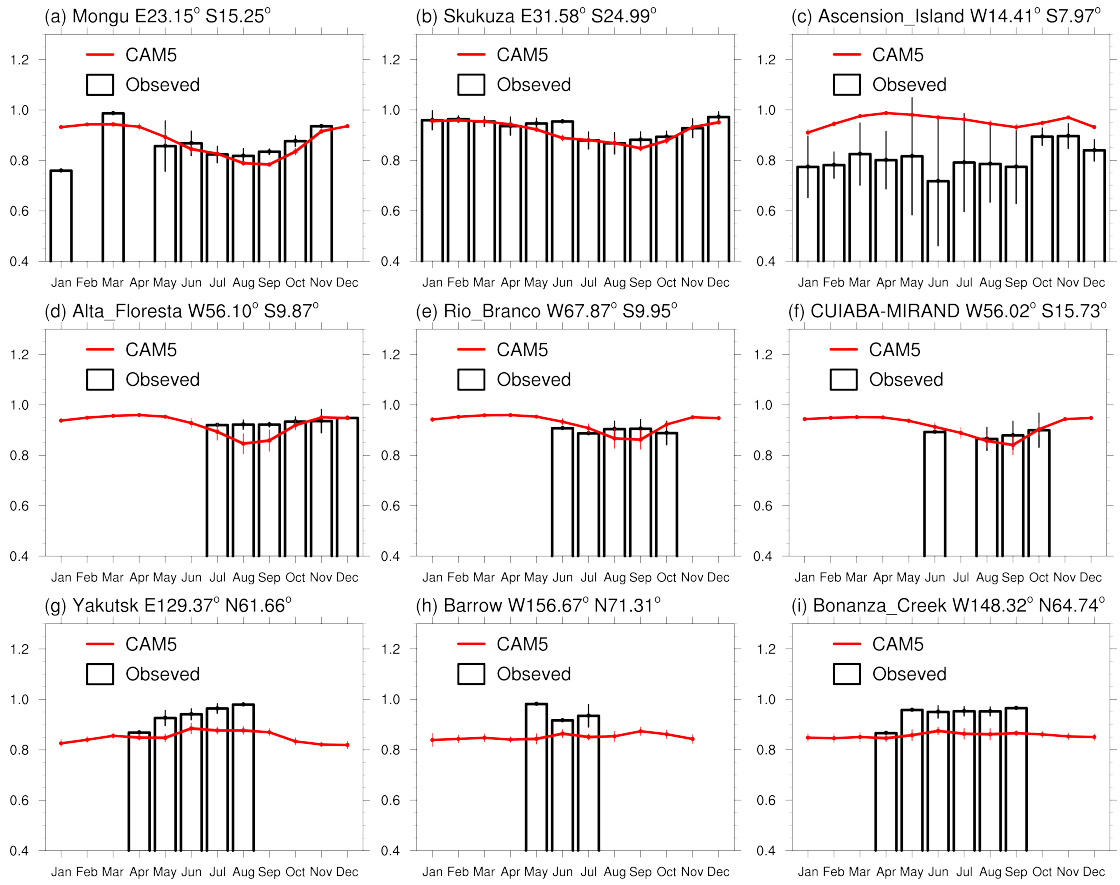

Figure 3. Same as Figure 2, but for the comparison of single scattering albedo (SSA) at 550 nm.

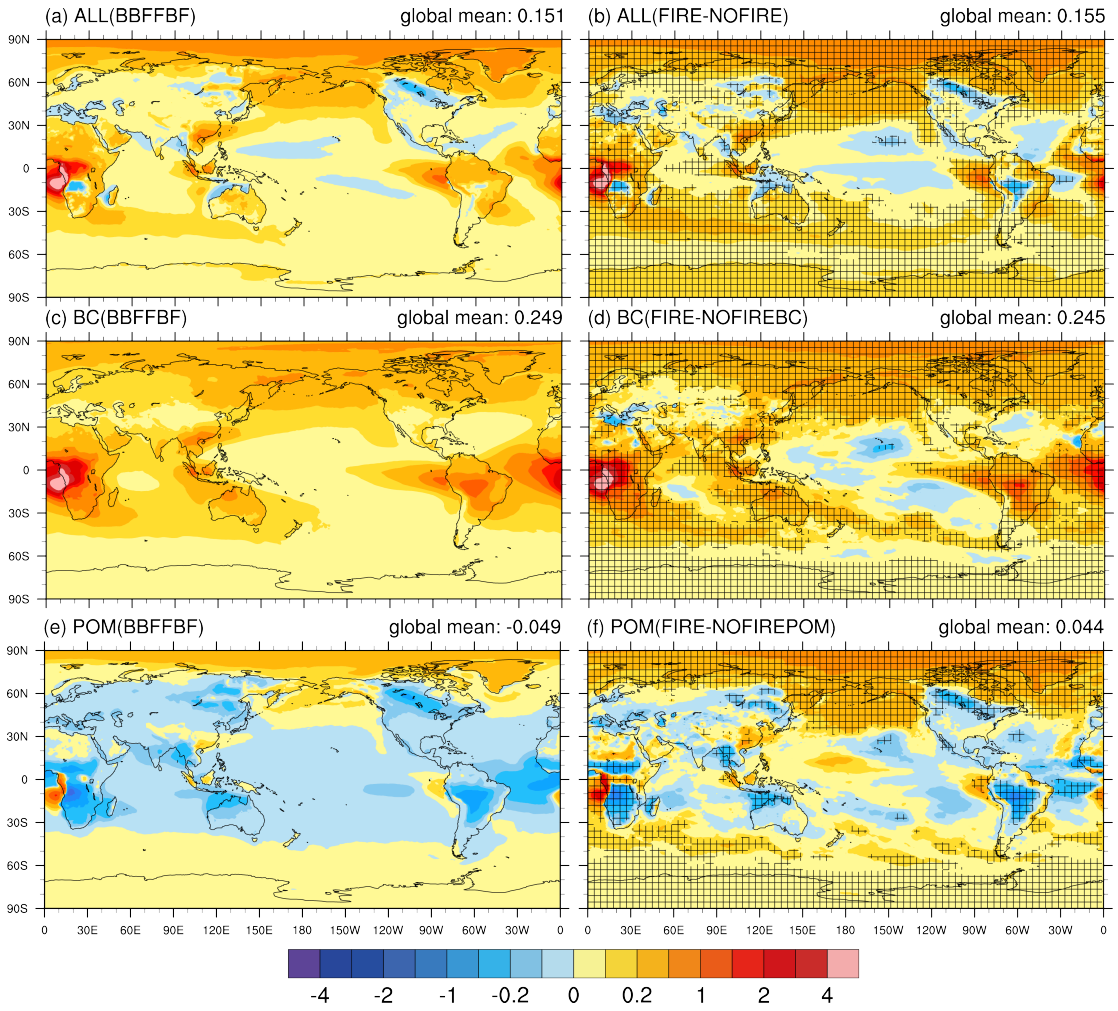

Figure 4. Annual mean radiative effect due to aerosol-radiation interactions (REari) (W m^{-2}) averaged over the period of 2003-2011 due to (a) all fire aerosols, (c) fire BC, and (e) fire POM estimated with the method of BBFFBF (left panels), and with the method of Ghan (2013) ((b), (d), and (f) in the right panels). The plus signs in Figure 4(b), (d) and (f) denote the regions where the radiative effect estimated with Ghan [2013] is statistically significant at the 0.05 level.

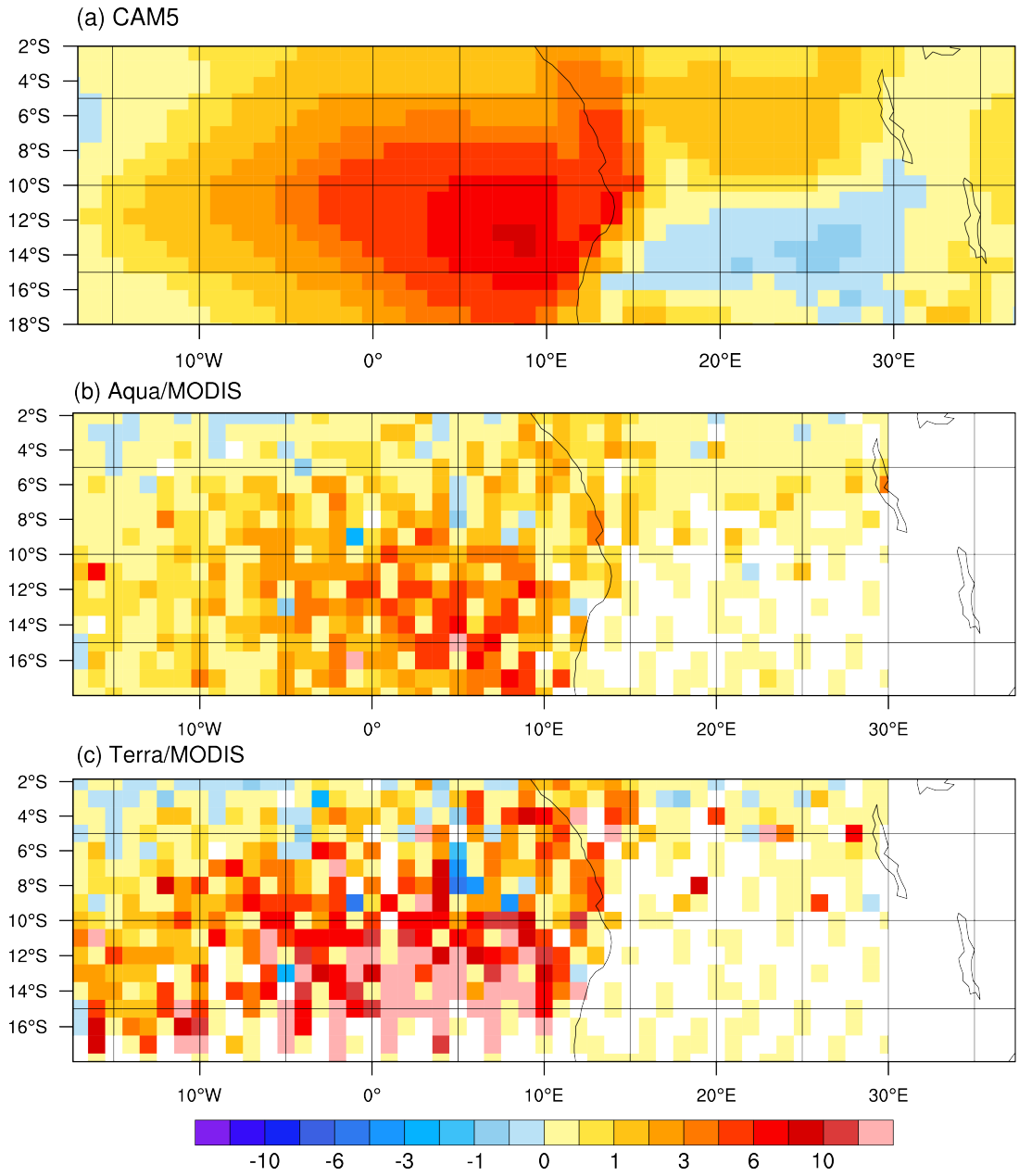

Figure 5. (a) September-October-November (SON) mean fire aerosol radiative effect due to aerosol-radiation interactions (REari) (W m^{-2}) for the period of 2003-2011 over the Southeast Atlantic Ocean due to all fire aerosols. (b) and (c) are the same as (a), but for the above-cloud aerosol REari for the period of 2007-2011 estimated using Aqua/MODIS and Terra/MODIS products [*Zhang et al.*, 2014], respectively.

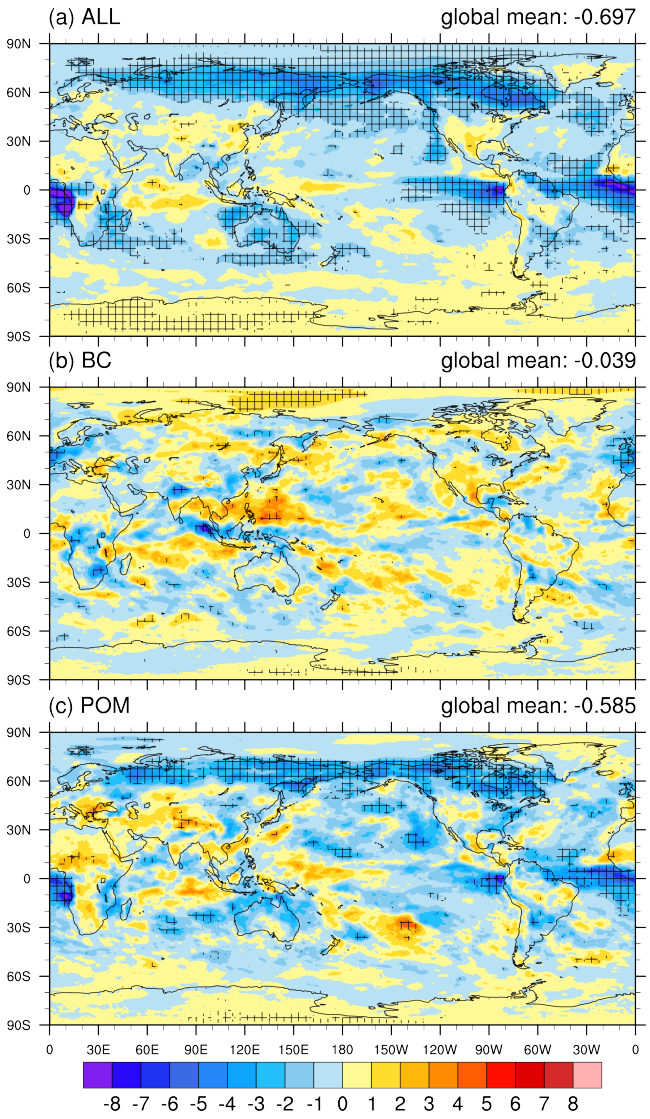

Figure 6. Annual mean radiative effect due to aerosol-cloud interactions (REaci) (W m^{-2}) averaged over the period of 2003-2011 due to (a) all fire aerosols, (b) fire BC, and (c) fire POM. The plus signs denote the regions where the radiative effect is statistically significant at the 0.1 level.

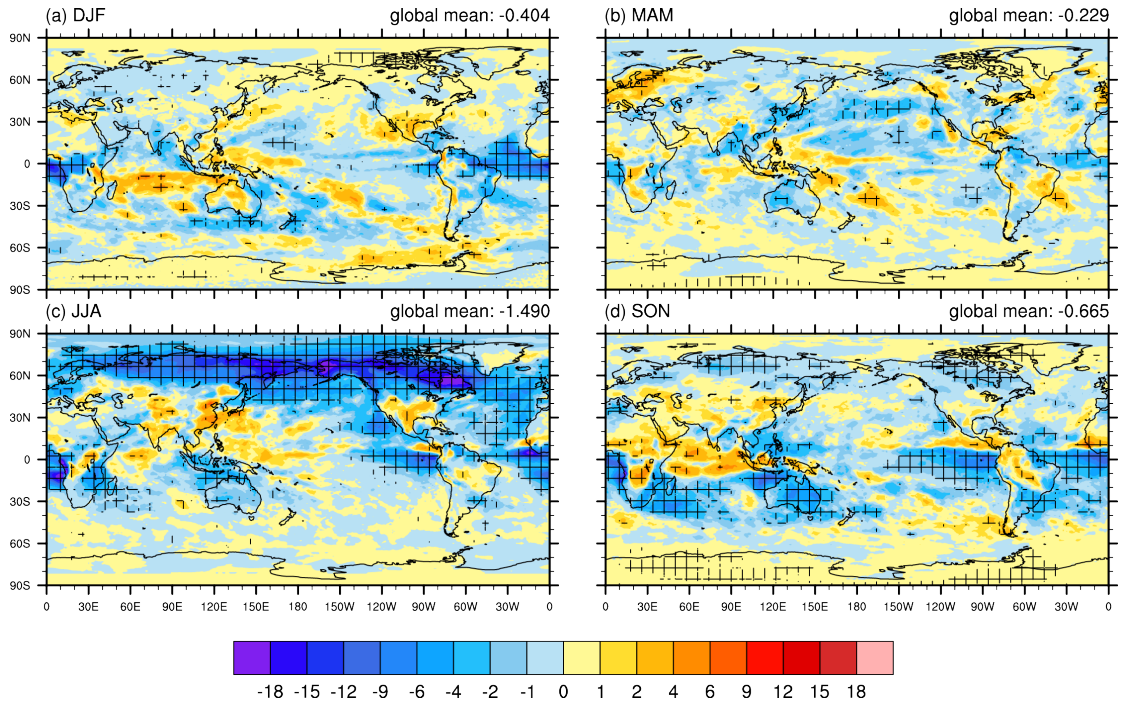

Figure 7. Seasonal variation of radiative effect of all fire aerosols due to aerosol-cloud interactions (REaci) (W m^{-2}) for the period of 2003-2011 for (a) December-January-February (DJF), (b) March-April-May (MAM), (c) June-July-August (JJA), and (d) September-October-November (SON). The plus signs denote the regions where the radiative effect is statistically significant at the 0.05 level.

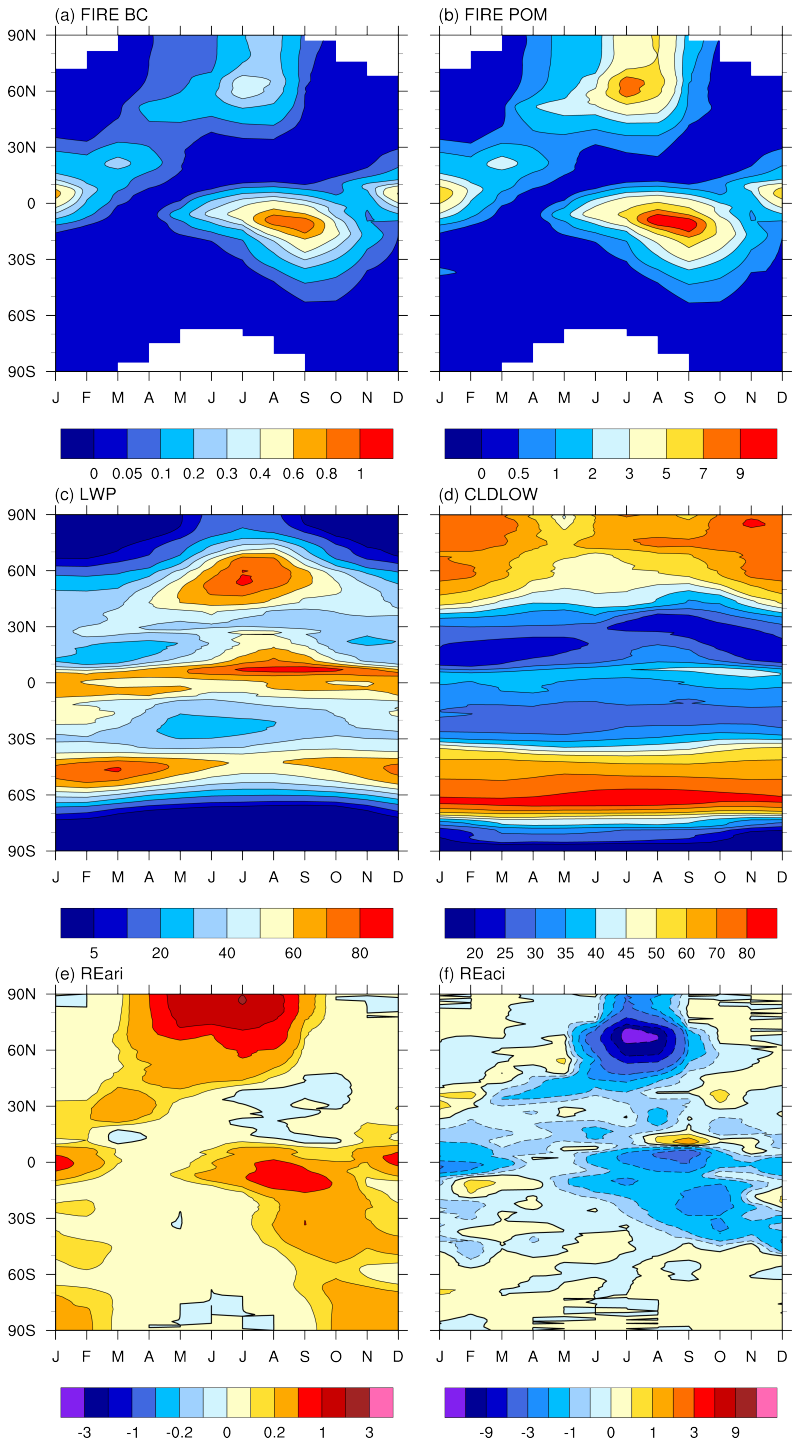

Figure 8. Month-latitude cross sections of zonal mean and monthly (a) vertically-integrated concentrations (mg m^{-2}) of fire BC and (b) fire POM, (c) cloud liquid water path (LWP, in g m^{-2}), (d) low-level cloud cover (CLDLOW, in %), (e) radiative effect due to aerosol-radiation interactions (REari, in W m^{-2}), and (f) radiative effect due to aerosol-cloud interactions (REaci, in W m^{-2}) of all fire aerosols.

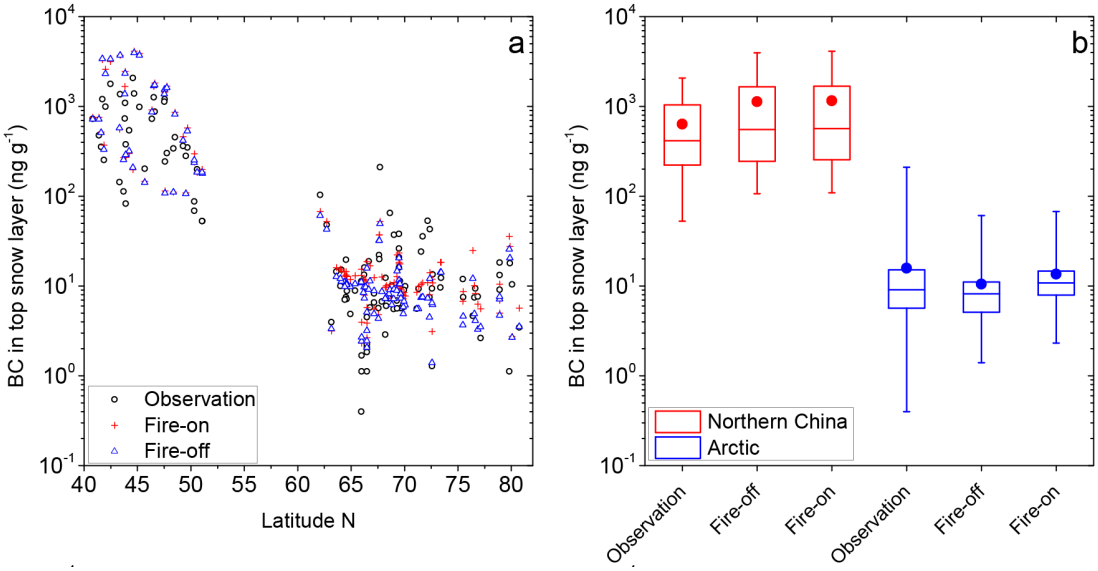

Figure 9. Evaluation of CAM5 simulated black carbon (BC) concentration for the period of 2003-2011 (in ng g^{-1}) in the top snow layer against observations in the Arctic [*Doherty et al.*, 2010] and Northern China [*Wang et al.*, 2013b]. The top snow layer ranges in thickness from 1 to 3 cm. Configuration of the two CAM5 simulations (FIRE and NOFIRE) is summarized in Table 1. Panel (a) shows the comparisons at different latitudes. The box and whisker plot in panel (b) shows the minimum and maximum value with the bar, the 25th and 75th percentiles with the box, the 50th percentile (i.e., median) by the bar within the box, and the mean value with the dot.

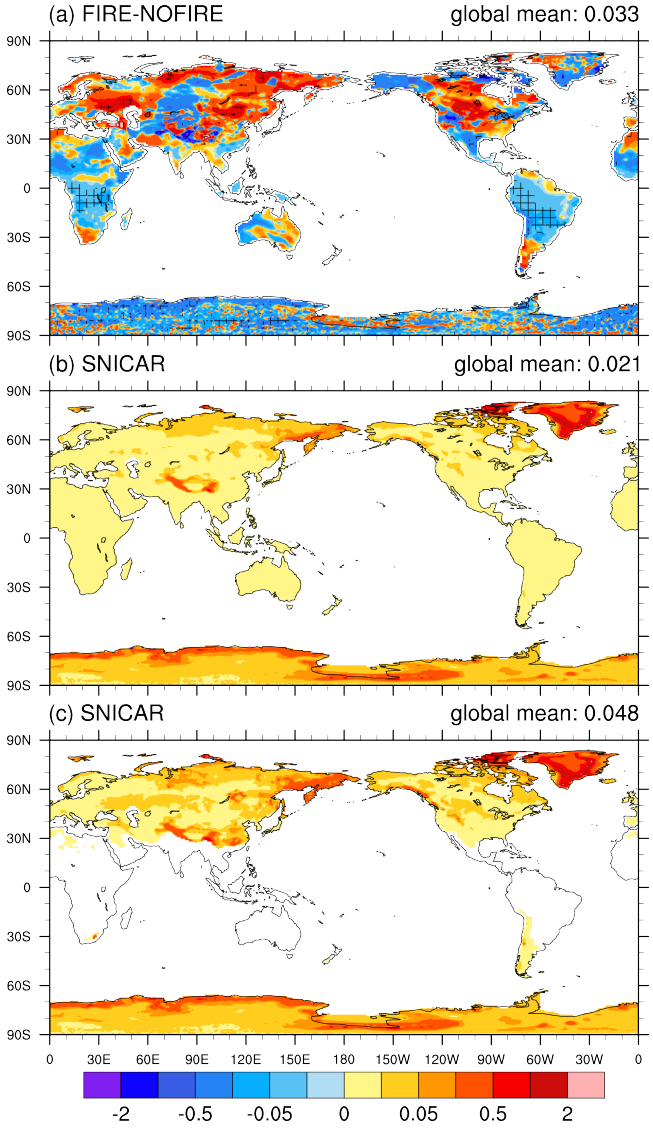

Figure 10. (a) Annual mean radiative effect due to surface albedo changes (REsac, W m^{-2}) averaged over the period of 2003-2011 of all fire aerosols over land regions, and annual mean surface effect of fire BC-in-snow calculated from SNICAR averaged (b) over all times and (c) only when snow is present. The plus signs in (a) denote the regions where the radiative effect is statistically significant at the 0.1 level.

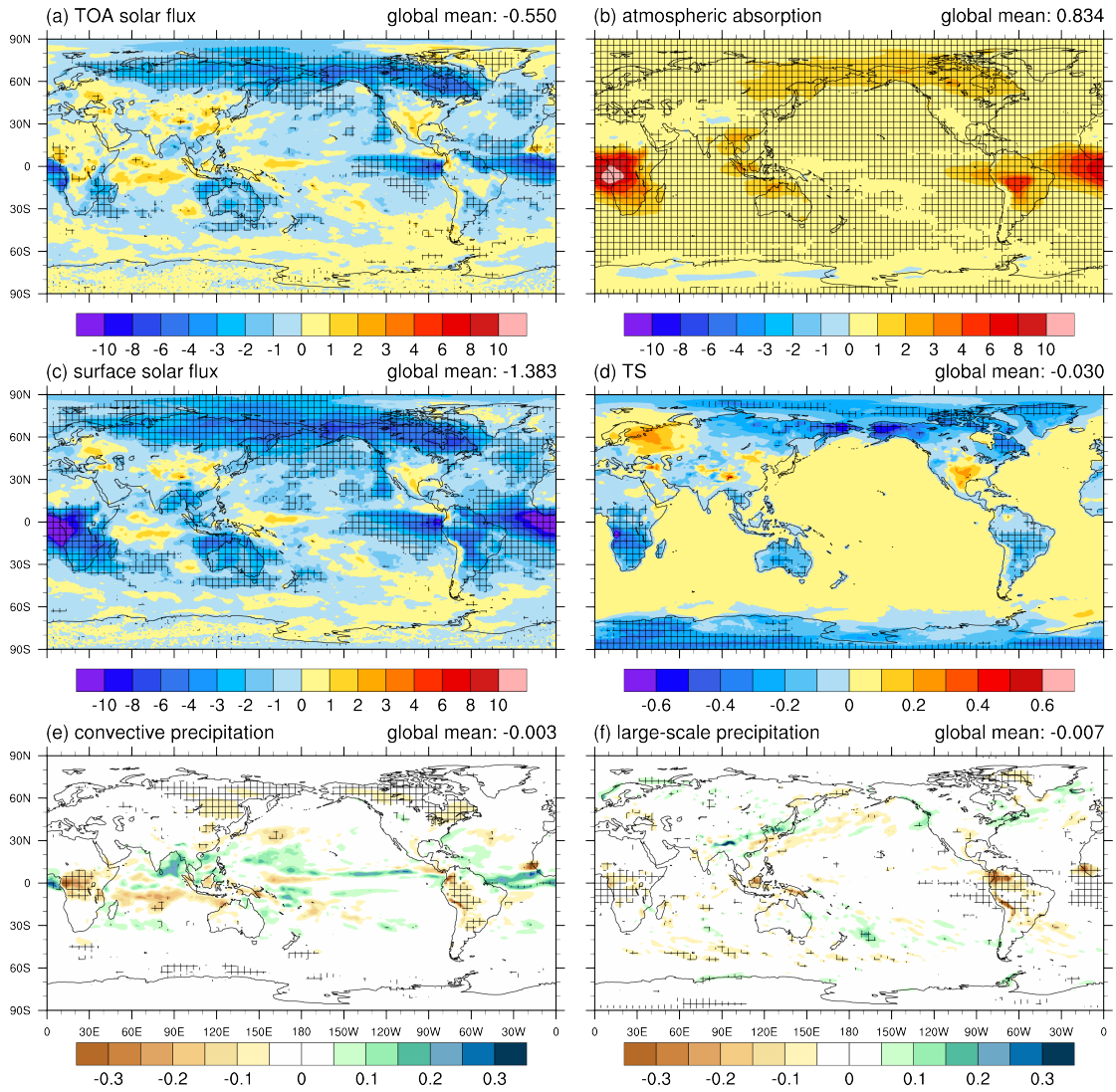

Figure 11. Annual mean net shortwave flux changes (W m^{-2}) over the period of 2003-2011 (a) at top of the atmosphere, (b) in the atmosphere, (c) at surface, and changes of (d) surface air temperature (TS, in K), (e) convective precipitation (mm d^{-1}), and (f) large-scale precipitation (mm d^{-1}) due to all fire aerosols. The plus signs denote the regions where the change is statistically significant at the 0.1 level.

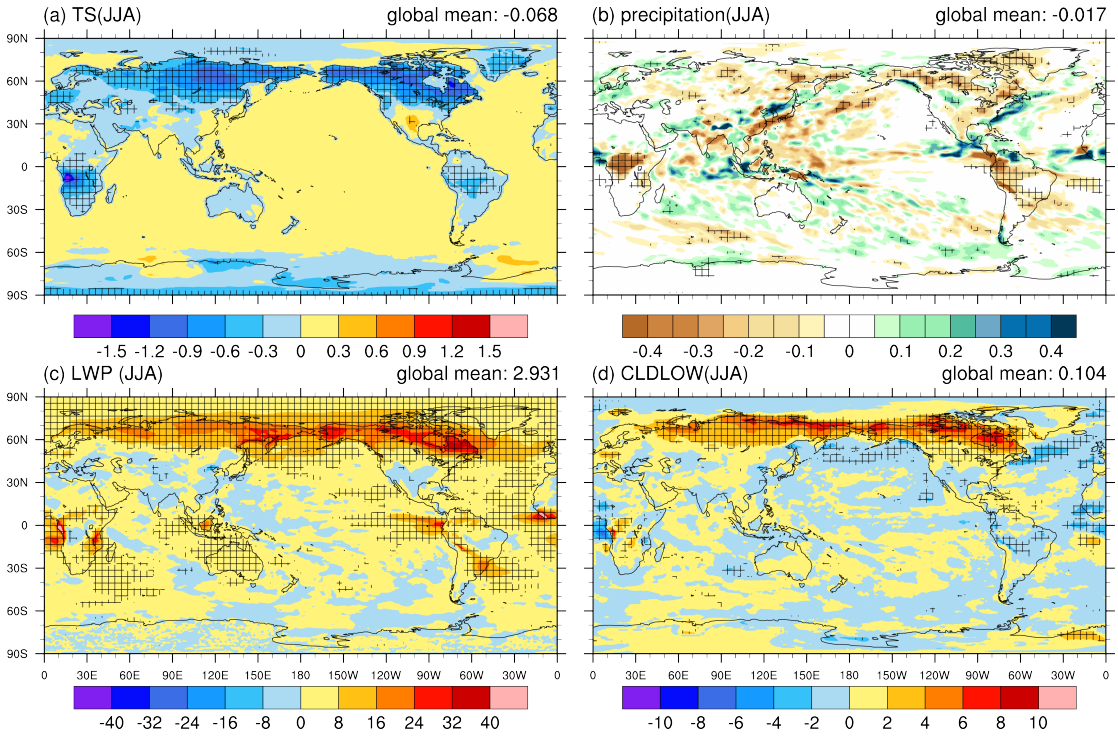

Figure 12. Changes in (a) surface air temperature (TS, in K), (b) total precipitation (mm d^{-1}), (c) cloud liquid water path (LWP, in g m^{-2}), and (d) low-level cloud cover (CLDLOW, in %) due to all fire aerosols in the boreal summer (JJA) averaged for the period of 2003-2011. The plus signs denote the regions where the change is statistically significant at the 0.1 level.