Efficacy of black carbon aerosols: the role of shortwave cloud feedback

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Abstract
Using idealized climate model simulations, we investigate the effectiveness of black carbon (BC) aerosols in warming the planet relative to CO$_2$ forcing. We find that a 60-fold increase in the BC aerosol mixing ratio from the present-day levels leads to the same equilibrium global mean surface warming ($\sim$4.1 K) as for a doubling of atmospheric CO$_2$ concentration. However, the radiative forcing is larger ($\sim$5.5 Wm$^{-2}$) in the BC case relative to the doubled CO$_2$ case ($\sim$3.8 Wm$^{-2}$) for the same surface warming indicating the efficacy (a metric for measuring the effectiveness) of BC aerosols to be less than CO$_2$. The lower efficacy of BC aerosols is related to the differences in the shortwave (SW) cloud feedback: negative in the BC case but positive in the CO$_2$ case. In the BC case, the negative SW cloud feedback is related to an increase in the tropical low clouds which is associated with a northward shift ($\sim$7°) of the Intertropical Convergence Zone (ITCZ). Further, we show that in the BC case fast precipitation suppression offsets the surface temperature mediated precipitation response and causes $\sim$8% net decline in the global mean precipitation. Our study suggests that a feedback between the location of ITCZ and the interhemispheric temperature could exist, and the consequent SW cloud feedback could be contributing to the lower efficacy of BC aerosols. Therefore, an improved representation of low clouds in climate models is likely the key to understand the global climate sensitivity to BC aerosols.

Introduction
Black carbon (BC) aerosols, unlike most of the atmospheric aerosols, absorb solar radiation and thus have a warming effect on the planet. This warming effect contrasts with the cooling effect of other aerosols that are primarily scattering (e.g. sulfate aerosols). BC aerosols can also affect the climate system by altering the clouds through local heating (semi-direct effect) or by acting as cloud condensation nuclei (indirect effect). According to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC 2014), BC aerosols are likely the third largest anthropogenic forcing agent after CO$_2$ and CH$_4$, with a present-day radiative forcing of $\sim$0.40 Wm$^{-2}$ relative to the pre-industrial period which is $\sim$25% of the present-day CO$_2$ radiative forcing.

Despite its importance, only few modeling studies (e.g. Cook and Highwood 2004, Roberts and Jones 2004, Hansen et al 2005, Yoshimori and Broccoli 2008, Stjern et al 2017, Smith et al 2018) have examined the effectiveness of BC aerosols in warming the planet relative to CO$_2$ forcing. These studies find that the BC aerosols are less effective in warming the planet compared to CO$_2$. Hansen et al (2005) introduced the concept of ‘efficacy’ to measure the effectiveness of forcing agents. Efficacy is defined as the ratio of the equilibrium global mean temperature change per unit forcing by that agent to the equilibrium temperature change per unit CO$_2$ forcing from the same initial climate state. Basically, it is the ratio of the climate sensitivity to the forcing agent to that due to CO$_2$ forcing.

Hansen et al (2005) find the efficacy of fossil fuel emitted BC aerosols to be 0.93 and biomass burning emitted BC aerosols to be 0.81 when effective radiative forcing (ERF) definition is used to estimate the radiative forcing. They find that the efficacy is closer to unity when the BC aerosols are in the planetary
boundary layer and it reduces drastically as the aerosols are moved higher into the atmosphere. They suggest that this happens because the low-altitude BC aerosols reduces cloud cover by heating the local layers but BC above the boundary layer inhibits underlying convection and therefore leads to moistening of the boundary layer and enhanced cloud cover in the lower layers. This cloud enhancement in the lower troposphere greatly reduces the direct warming effect of the BC aerosols. Such an enhancement in the low clouds is also found in a multi-model ensemble mean in a recent study (Stjern et al 2017) which finds that the multi-model median efficacy of BC aerosols is less than one (0.80).

The decrease in efficacy with altitude of BC aerosols has been found in some recent studies (Ban-Weiss et al 2012, Samset and Myhre 2015). For instance, by prescribing fixed mass of BC aerosols (1 Mt) uniformly around the globe at various altitudes, Ban-Weiss et al (2012) showed that the BC aerosols near the surface causes surface warming in contrast to when they are prescribed at higher altitudes in the troposphere. The variation of the climate response from BC aerosols at different altitudes arises primarily due to different fast climate adjustments which is defined as the climate response before any change in the global mean surface temperature (Bala et al 2010, Ban-Weiss et al 2012). BC aerosols near the surface warms the surface through diabatic heating whereas at higher altitudes the absorbed solar radiation is lost to space in the form of increased longwave radiation without heating the surface. We note that the efficacy decreases with altitude of BC aerosols but the efficiency of BC aerosols to exert larger radiative forcing due to the direct aerosol effect strengthens with altitude (Samset and Myhre 2011, 2015).

Thus, while investigating the climate sensitivity or the efficacy of BC aerosols most of these studies examined the dependence of climate response on the altitude of BC aerosols. However, a process-based mechanistic explanation of the lower efficacy of BC aerosols is still lacking in the literature. In this study, we revisit the efficacy of BC aerosols and provide a systematic explanation of the processes that are associated with the lower efficacy of BC aerosols. We adopt the ERF definition (Myhre et al 2013) to estimate the efficacy of BC aerosols. Past studies (Ban-Weiss et al 2012, Myhre et al 2013, Forster et al 2016) have shown ERF to be a better predictor of equilibrium surface temperature change than the instantaneous or stratosphere-adjusted radiative forcing. In IPCC 2013, the radiative forcing estimates calculated from the Hansen’s prescribed sea–surface-temperature (SST) and Gregory’s regression methods (Gregory et al 2004), discussed in supplementary section S1, which is available online at stacks.iop.org/ERL/14/084029/mmedia, are referred as the ERF (Myhre et al 2013).

To understand the mechanisms associated with lower efficacy of BC aerosols, we decompose the total response of the climate system into fast adjustments and slow response (Andrews et al 2009, Dong et al 2009, Bala et al 2010). The fast adjustment relates to the changes in the thermal structure of the atmosphere and associated variables such as water vapor and clouds before a change in global mean surface temperature, and the slow response or feedback refers to changes that occur in response to changes in global mean surface temperature. However, we caution that no clear separation exists between fast adjustment and slow response as the evolution of climate change is a continuous process (Cao et al 2015).

In this study, we show that the difference in the ERF between CO₂ and BC aerosols are associated with the differing fast adjustments, and the difference in the efficacy of BC aerosols versus CO₂ are primarily linked to difference in the slow response (feedbacks). However, it is also important to note that the differences in the fast adjustments could produce vastly differing climate states from which the slow response begins to evolve and hence the differing fast adjustments are also likely to influence the slow climate feedbacks and hence the efficacy.

Model details

The Community Atmosphere Model, CAM5 (Neale et al 2012) coupled to the Community Land Model CLM4 developed by the National Centre for Atmospheric Research (NCAR) is used in this study. For our simulations, we use CAM5 in two configurations: prescribed-SST and slab ocean model ( SOM). The SOM is a simplification of the full ocean model and it is useful to study equilibrium climate change on decadal time scales. The prescribed-SST configuration is used to compute the radiative forcing and fast climate adjustments. The horizontal resolution of the model is 1.9° × 2.5° (latitude × longitude) and there are 30 vertical levels for the atmosphere (model top is at ~3.5 hPa). The model’s vertical coordinate is a hybrid sigma-pressure system with upper levels in pressure coordinates and lower levels in sigma coordinates (Neale et al 2012).

In the default configuration, CAM5 is coupled to the Modal Aerosol Model (MAM; Liu et al 2012). However, to study efficacy of BC aerosols CAM5 coupled to the Bulk Aerosol Model (BAM; Tie et al 2005, Emmons et al 2010, Lamarque et al 2012) where aerosols are prescribed, is used. The primary reason for adopting CAM5-BAM is to reduce the computational time needed for the model simulations. This aided us to perform several long simulations to study the efficacy of BC aerosols. BAM simulates only the direct and the semi-direct effects of tropospheric aerosols. Hence, in this study, our analysis includes only the
direct effects (radiation, scattering and absorption by the aerosols) and the semi-direct effects (changes in climate due to the local heating from BC aerosols). The indirect aerosol effect, in which aerosol particles act as cloud condensation nuclei, is not considered. We provide further details on BAM in supplementary section S2.

Experiments

We perform four simulations: (i) a ‘CTL’ simulation, where the atmospheric CO2 concentration is prescribed at pre-industrial level (284.7 ppm), (ii) a ‘2 × CO2’ simulation, which is same as CTL but the CO2 concentration is doubled from the pre-industrial level, (iii) a ‘1 × BC’ simulation with prescribed BC aerosol distribution at 2000 climatological level (figure S1) and CO2 is prescribed at the pre-industrial levels and (iv) a ‘60 × BC’ simulation, which is same as 1 × BC, but the prescribed BC mixing ratio is 60 times the 2000 climatology spatial distribution is same as the 2000 climatology, only the magnitude is increased (figure S1). Summary of the experiments are given in tables S1 and S2, and figure S2 shows the surface temperature evolution in all the four simulations. Our choice of increasing the mixing ratio of BC aerosols by 60-folds in the 60 × BC experiment is determined by the constraint that we simulate the same equilibrium global mean surface warming in the 2 × CO2 and 60 × BC cases relative to the CTL and 1 × BC simulations, respectively. In both 2 × CO2 and 60 × BC experiments, we have applied the forcings as step function changes at the start of the simulations. Further details on the experiments is discussed in supplementary section S3.

Results

Radiative forcing and efficacy estimate using Hansen’s prescribed-SST approach

In Hansen’s prescribed SST method (Hansen et al 1997), the radiative forcing is estimated as the top-of-atmosphere (TOA) net radiative flux change after the introduction of the forcing agent in prescribed-SST simulations. This radiative forcing is termed as ‘prescribed-SST radiative forcing’. In this approach, the TOA radiative flux changes are calculated after the troposphere, stratosphere and land surface have adjusted. Hence, this method considers, for example, some changes in cloud amounts to be part of the climate forcing rather than part of the climate response. In this section, the slow response which is defined as the climate response that is mediated by the change in global mean surface temperature (Bala et al 2010), is calculated as the difference between the total climate response estimated from the slab ocean (SOM) simulations and the fast adjustments estimated from the prescribed-SST simulations (Bala et al 2010). The climate feedback parameters are estimated as the slow response (for the radiative fluxes) normalized by the global mean surface temperature change from the SOM simulations.

As per our experiment design, the global mean equilibrium surface warming is the same (~4.1 K) in both the 2 × CO2 and 60 × BC cases (figures 1(c) and (d)). However, we find that the prescribed-SST radiative forcing is 3.78 ± 0.11 Wm⁻² in the 2 × CO2 case and 5.50 ± 0.13 Wm⁻² in the 60 × BC case (figures 1(a) and (b); table 1). The prescribed-SST radiative forcing is positive throughout the globe in the 2 × CO2 case but in the 60 × BC case, while positive over land, it is negative over most of the ocean (figures 1(a) and (b)). The negative prescribed-SST radiative forcing over the ocean in the 60 × BC case is primarily associated with the increased low clouds over the ocean in the fast adjustment phase (figure S3b).

Thus, we find that for the same surface temperature change, a ~45% larger radiative forcing in the 60 × BC case relative to the 2 × CO2 case is needed, and the efficacy of BC aerosols by definition is 0.69 ± 0.03. Our estimation of lower efficacy of BC aerosols is in agreement with several previous modeling studies (e.g. Cook and Highwood 2004, Roberts and Jones 2004, Hansen et al 2005, Yoshimori and Broccoli 2008) and a recent multi-model study which assessed the efficacy of BC aerosols in nine models (Stjern et al 2017). This multi-model study finds that the median efficacy is 0.80. However, two of the nine models simulate an efficacy larger than one: 1.21 in IPSL-CM5A and 1.12 in CanESM2. Stjern et al (2017) suggest that the larger efficacy in IPSL-CM5A is likely associated with a very large decrease in cloudiness. We can infer from figure 6 and table S2 of Stjern et al (2017) that in IPSL-CM5A the low clouds decreases substantially in the slow response phase which could lead to a strong positive SW cloud feedback unlike a negative SW cloud feedback in our simulations. Another study (Jacobson 2002) also finds the efficacy of BC aerosols to be larger than one (~2.33). This is primarily because Jacobson (2002) estimate the climate response function by considering only the direct aerosol forcing.

In the prescribed-SST simulations, only the SST and sea-ice extent are held fixed, but the land surface temperature is allowed to adjust-land surface warms by 0.27 ± 0.02 K in the 2 × CO2 case while it warms by 0.54 ± 0.02 K in the 60 × BC case (table 1). When we account for this land surface warming (Hansen et al 2005) and the TOA radiative imbalances in the SOM simulations, we find the values of climate feedback factors (slopes in figure S4a) to be \( \lambda_{\text{CO2}} = -0.98 \pm 0.05 \text{ Wm}^{-2} \text{ K}^{-1} \) and \( \lambda_{\text{BC}} = -1.51 \pm 0.05 \text{ Wm}^{-2} \text{ K}^{-1} \) (using equation 4 of (Modak et al 2018), table 1). Using these values of climate feedback parameters, we
estimate the efficacy of BC aerosol as $0.65 \pm 0.04$ (using equation 5 of Modak et al 2018, table 1) which is close to the original estimate which did not account for the land surface warming. Therefore, we find that the inclusion of land surface temperature change from the prescribed-SST simulations and the slight TOA radiative imbalance in the SOM simulations in the definition of climate feedback parameter have little effect on the estimation of efficacy. We note that the slopes in figure S4 are obtained by plotting the equilibrium global-and annual-mean TOA net radiative fluxes (and the SW and LW components) versus the global-and annual-mean surface temperature change from the prescribed-SST (points on the left) and SOM simulations (points on the right) for the $2 \times CO_2$ and $60 \times BC$ cases relative to their respective control simulations.

As discussed earlier, we also estimate the efficacy of BC aerosols by adopting the Gregory’s regression method (details are in supplementary section S1). We find that for the same equilibrium surface warming, the regressed radiative forcing is $3.39 \pm 0.68 \text{ Wm}^{-2}$ in the $2 \times CO_2$ case and $5.58 \pm 1.00 \text{ Wm}^{-2}$ in the $60 \times BC$ case (figure S5a; table 1). We find the slopes of the linear regression (climate feedback parameters) to be $-0.90 \pm 0.28$ in the $2 \times CO_2$ case and $-1.69 \pm 0.38$ in the $60 \times BC$ case (figure S5a). Thus, we estimate the efficacy of BC forcing as $0.53 \pm 0.20$. In consistent with the findings of past studies (Forster et al 2016, Modak et al 2018), we find that the uncertainty range is larger when the regression approach is used to estimate the efficacy value compared to the prescribed-SST method. This is because the regression approach requires larger ensemble to reduce the standard error of the estimate while the prescribed-SST simulations can be extended to any length of time required to achieve a smaller standard error of the estimate. Therefore, the use of the prescribed-SST method over the regression method for radiative forcing estimates has been recommended recently (Forster et al 2016).

**Differences in fast climate adjustments**

The difference in the radiative forcing ($\sim 1.7 \text{ Wm}^{-2}$) between the $2 \times CO_2$ and $60 \times BC$ cases for the same equilibrium surface warming is likely linked to the differing fast adjustments between the two cases. Unlike $2 \times CO_2$, where the stratosphere cools while the troposphere warms, in the $60 \times BC$ case the entire depth of the atmosphere warms (figure 2(a)) because of the shortwave (SW) absorption by the BC aerosols. This is evident from the vertical profile of change in SW heating rate and temperature in the 1 d simulations which approximate the instantaneous response of the climate system to forcing (figure S6). This SW induced warming causes reduction in the mid- and high-level clouds but an increase in low-level cloudiness (figure 2(d)) in the fast adjustment phase, also found in Stjern et al (2017). The heating of the atmosphere above the boundary layer due to SW
Table 1. The NCAR CAM5 simulated global- and annual-mean changes in some key climate variables for the $2 \times CO_2$ case relative to CTL and for the $60 \times BC$ case relative to the $1 \times BC$ case. The uncertainties in the changes are represented by 2 standard errors. The number of annual means used for estimating the mean and standard error is listed in the 2nd column. The subscripts ‘som’ and ‘psst’ refer to the slab ocean and prescribed-SST simulations, respectively, and the subscripts ‘clr’ and ‘cld’ refer to the clear- and cloudy-sky values.

| Variable | No. of annual means used | $2 \times CO_2$ | $60 \times BC$ |
|----------|--------------------------|----------------|---------------|
| $\Delta TS_{som}$ (Surface temperature change (K)) | 50 | 4.12 ± 0.04 | 4.15 ± 0.04 |
| $\Delta TOA_{som}$ (Steady-state TOA radiative imbalance (W m$^{-2}$)) | 50 | 0.01 ± 0.13 | 0.04 ± 0.13 |
| $\Delta P_{som}$ (Precipitation change (%)) | 50 | 7.89 ± 0.10 | -8.03 ± 0.12 |
| Apparent hydrological sensitivity (%K$^{-1}$) | 50 | 1.91 ± 0.03 | -1.93 ± 0.03 |

Prescribed-SST model results

| Variable | No. of annual means used | $2 \times CO_2$ | $60 \times BC$ |
|----------|--------------------------|----------------|---------------|
| $\Delta TS_{psst}$ (surface temperature change (K)) | 20 | 0.27 ± 0.02 | 0.54 ± 0.02 |
| $\Delta TOA_{psst}$ (net TOA radiative imbalance (W m$^{-2}$)) | 20 | 3.78 ± 0.11 | 5.50 ± 0.13 |
| $\Delta P_{psst}$ (Precipitation change (%)) | 20 | -2.54 ± 0.12 | -17.78 ± 0.10 |
| LW radiative forcing (W m$^{-2}$) | 20 | 3.15 ± 0.08 | -6.90 ± 0.08 |
| SW radiative forcing (W m$^{-2}$) | 20 | 0.63 ± 0.08 | 12.40 ± 0.10 |
| LW$cld$ radiative forcing (W m$^{-2}$) | 20 | 3.71 ± 0.06 | -3.49 ± 0.06 |
| LW$cld$ radiative forcing (W m$^{-2}$) | 20 | -0.56 ± 0.10 | -3.41 ± 0.10 |
| SW$cld$ radiative forcing (W m$^{-2}$) | 20 | 0.26 ± 0.04 | 7.52 ± 0.04 |
| SW$cld$ radiative forcing (W m$^{-2}$) | 20 | 0.37 ± 0.09 | 4.88 ± 0.11 |

Combined slab ocean and prescribed-SST model results

| Efficacy which do not account for land warming and TOA imbalance | — | 0.69 ± 0.03 |
| $\Delta TOA_{som} - \Delta TOA_{psst}$ (W m$^{-2}$) | -3.77 ± 0.17 | -5.46 ± 0.18 |
| $\Delta TS_{som} - \Delta TS_{psst}$ (K) | 3.85 ± 0.05 | 3.61 ± 0.05 |
| Radiative forcing (W m$^{-2}$) which accounts for land warming and TOA imbalance | 4.04 ± 0.14 | 6.32 ± 0.15 |
| $\lambda_{psst}$ (Feedback parameter (W m$^{-2}$ K$^{-1}$) which accounts for land warming and TOA imbalance) | -0.98 ± 0.05 | -1.51 ± 0.05 |
| Efficacy which accounts for land warming and TOA imbalance | — | 0.65 ± 0.04 |

Results from the regression method

| Regressed radiative forcing (W m$^{-2}$; intercept in figure S5a) | 10 | 3.39 ± 0.68 | 5.58 ± 1.00 |
| Feedback parameter (W m$^{-2}$ K$^{-1}$; slope in figure S5a) | 10 | -0.90 ± 0.28 | -1.69 ± 0.38 |
| Efficacy | — | 0.53 ± 0.20 |
| Fast precipitation adjustment (%; intercept in figure S19) | 10 | -2.94 ± 0.50 | -17.34 ± 0.64 |
| Hydrological sensitivity (%K$^{-1}$; slope in figure S19) | 10 | 2.66 ± 0.20 | 2.16 ± 0.24 |

Absorption by BC aerosols stabilizes the atmosphere. This prevents export of water vapor from the boundary layer resulting in moistening of the boundary layer and a concomitant increase in the low-level cloudiness (Hansen et al. 2005).

The SW absorption by BC aerosols results in positive SW clear-sky radiative flux change at TOA (figure S7b) while the negative LW clear-sky radiative flux change is associated with the increase in LW radiation to space because of the upper atmospheric warming in the $60 \times BC$ case (figures 2(a) and S7b). It is apparent from figure S7 that the larger prescribed-SST radiative forcing in the $60 \times BC$ case relative to the $2 \times CO_2$ case is associated with the differences in the net cloudy-sky radiative fluxes at TOA and hence are linked to the differences in the fast cloud adjustments between $2 \times CO_2$ and $60 \times BC$ cases. Although the low-level clouds increase in the $60 \times BC$ case, a decrease in the mid- and upper level clouds associated with the SW induced warming leads to a significant positive SW cloudy-sky forcing (figures 2(d) and S7b). A decrease in high clouds leads to the negative LW cloudy-sky forcing (figures 2(d) and S7b). It is interesting to note that Stjern et al. (2017) found that while cloud increase in the tropics caused positive rapid adjustment flux response there, the global mean rapid adjustment is negative. Thus, we should recognize that the details of the cloud responses can alter the sign of the rapid adjustment depending on the model details.

The larger SW cloudy-sky forcing in the $60 \times BC$ case (figure S7c) leads to a net positive cloudy-sky forcing which is close to the difference in the prescribed-SST radiative forcing ($\sim$1.7 W m$^{-2}$) between the $2 \times CO_2$ and $60 \times BC$ cases. This confirms that the difference in the prescribed-SST radiative forcing between the $2 \times CO_2$ and $60 \times BC$ cases is associated with the differences in the net cloudy-sky radiative fluxes at TOA and hence due to the differing fast cloud adjustments.
Figure 2. The vertical profile of the NCAR CAM5 simulated (a) fast adjustment, (b) total response and (c) slow response in global- and annual-mean temperature (K) and (d) fast adjustment, (e) total response and (f) slow response in the global- and annual-mean cloud (fraction) for the 2 × CO₂ (black) and the 60 × BC (red) cases relative to their respective control simulations. The slow response is estimated by subtracting fast adjustment from the total response. Last 20 years of the 40 year prescribed-SST simulation are used to estimate the fast adjustment and last 50 years of the 100 year SOM simulation are used to estimate the total response.

Figure 3. (a) The TOA shortwave clear-sky (SWclr), longwave clear-sky (LWclr), shortwave cloudy-sky (SWcld), longwave cloudy-sky (LWcld) and (b) net, SW and LW feedback factors (Wm⁻² K⁻¹, slopes from figure S4b) for the 2 × CO₂ (black) and 60 × BC (red) cases relative to their respective control simulations.
Differing climate feedback factors

We recognize that the differences in the fast cloud adjustments are likely to cause the difference in the radiative forcing but what causes the climate feedback factors (slopes in figure S4a) to differ between the $2 \times CO_2$ and $60 \times BC$ cases? From figure 3 which shows the SW and LW components of the climate feedback factors (slopes in figure S4b), it can be seen that while the SW clear-sky, LW clear-sky and the LW cloud feedbacks are similar for both $2 \times CO_2$ and $60 \times BC$ cases, the SW cloud feedback is of opposite sign—positive in the $2 \times CO_2$ case but negative in the $60 \times BC$ case (figure 3(a)). In the $2 \times CO_2$ case, reduced low and mid-clouds predominately in the sub-tropics (figure S8j-k) leads to a positive SW cloud feedback (Modak et al. 2016). However, in the $60 \times BC$ case, increase in the low and mid-level clouds mainly over the tropics (figures S8j-k and S9b-c) causes a negative SW cloud feedback (figure 3(a)). This differing SW cloud feedback between the $2 \times CO_2$ and $60 \times BC$ cases is mainly responsible for the difference in the net climate feedback factors (figure 3(b)) and hence for the lower efficacy of BC aerosols.

We find that the increase in the low clouds over the tropics in the $60 \times BC$ case is related to the slow response of the low clouds discernible in the annual zonal mean plots of cloud response (compare figures S8i and S8j). As discussed in (Modak et al. 2018), it is likely that the differing climate states produced after the fast adjustments likely influences the slow response (SW cloud feedback in this case) of the climate system. We find that the increase in clouds in the $60 \times BC$ case over the tropics (figures S8i-l and S9) is primarily associated with the northward movement and intensification of the Intertropical Convergence Zone (ITCZ) and the ITCZ shift is larger in the slow response phase than during the fast adjustments (compare figure 4(b) and S10; figure S11).

The schematic in figure 4(a) illustrates how the low cloud increase over the NH as well as SH tropics causes a negative SW cloud feedback in the $60 \times BC$ case. For a global mean surface warming of $\sim 4.1$ K in $60 \times BC$ case, the NH warms $\sim 5$ K more than the SH (table S3,
figure 1(d) and S12) because of the presence of more BC aerosols in the NH. This interhemispheric surface temperature gradient drives the ITCZ to move into the NH in association with the Hadley cells (Chiang and Bitz 2005, Yoshimori and Broccoli 2008, Donohoe et al 2013, Devaraju et al 2015). Shifting of the ITCZ ensures the transport of the excess energy available in the warmer NH to the relatively cooler SH. As the NH is warmer, the ITCZ in the new location intensifies and consequently there is a substantial increase in the clouds in the NH tropics—the average low cloud fraction in the total response increases by \(\sim 25\%\) in the 60 × BC case, ±2° latitude around the ITCZ (figure 4(b)). However, we find that the low clouds also increase over the SH tropics especially in the eastern Pacific and Atlantic oceans (figure S9b).

The enhanced low clouds in the SH tropics is also associated with the northward movement of the ITCZ. When the ITCZ shifts to the north, the downwelling branch of the SH Hadley cell is located over the SH tropics. This subsidence and the associated moistening of the boundary layer increase the low clouds over the SH tropics. The enhanced low clouds both over the NH and SH tropics tends to cool the surface and thus, act as a negative feedback. Increased low clouds in the SH tropics cools the SH, consequently intensifying the interhemispheric surface temperature gradient which further pushes the ITCZ towards the north and thus strengthens the negative feedback. Our study suggests that a feedback between the interhemispheric temperature gradient and the location of the ITCZ could exist, and the consequent SW cloud feedback could be contributing to the lower efficacy of BC aerosols.

How much does the ITCZ shifts?

We estimate the location of ITCZ by using a metric called precipitation centroid (Pcent, Frierson and Hwang 2012, Donohoe et al 2013, Devaraju et al 2015)). It is defined as the median latitude of zonal mean area-weighted precipitation between 20°S and 20°N. In order to identify small shifts in the ITCZ position, we interpolate the ~2° latitude grid in to a 0.01° grid (Donohoe et al 2013, Devaraju et al 2015, Nalam et al 2018).

We find that there is a substantial northward shift of 7.03° ± 0.15 in the location of the annual mean ITCZ in the 60 × BC case from the annual mean position in the 1xBC case (~−1.5°S). However, in the 2 × CO₂ case there is relatively negligible northward shift of only 0.31° ± 0.20. Further, we find that the ITCZ shifts are largest during the months of May and June (~10°) in the 60 × BC case (figure S13a). In the control simulation the location of Pcent varies from 6° S to 8° N on the seasonal timescale and the interhemispheric temperature difference varies from −10 to 10 K (figure S13b). In the 60 × BC case, we find that the magnitude of the location change in annual mean Pcent is comparable to the seasonal migration of the position of ITCZ in the control case (figure S13).

We also estimate the corresponding annual mean northward heat transport (Donohoe et al 2013, Nalam et al 2018) in the 60 × BC case (figure S14). We find that for a ~7° shift in the ITCZ position, the magnitude of annual mean cross-equatorial atmospheric northward heat transport (AHTeq) to the SH from the NH is \(\sim −1.2\) PW. This implies that the rate at which the ITCZ shifts to the NH for a given BC induced forcing per unit change in the AHTeq is \(\sim 5.8\°\ PW\(^{-1}\).

Hydrological cycle

For the same equilibrium global mean surface warming, the global mean total precipitation (ΔP\(_{\text{tot}}\)) increases by 7.87% ± 0.10% in the 2 × CO₂ case while it decreases by 8.03% ± 0.12% in the 60 × BC case (figures 5(a) and (b)). The large precipitation decrease in the 60 × BC case is because of the precipitation reduction associated with the fast adjustment (ΔP\(_{\text{fast}}\) is −17.78% ± 0.10%) which dominates the temperature driven slow precipitation response (ΔP\(_{\text{slow}}\) is 9.75% ± 0.16%; figure 5(d), table S3). In the 2 × CO₂ case, the ΔP\(_{\text{fast}}\) is −2.54% ± 0.12% and ΔP\(_{\text{slow}}\) is 10.41% ± 0.16% (figure 5(c), table S3).

In the 2 × CO₂ case, the suppression of precipitation during the fast adjustment (−2.5%) occurs due to (i) over ocean: the stability of the lower troposphere increases for CO₂ forcing which leads to a precipitation reduction and (ii) over land: the CO₂ physiological effect—elevated levels of CO₂ cause a decrease in plant stomatal conductance and hence reduced transpiration and reduced precipitation as shown in past studies (e.g. Dong et al 2009, Bala et al 2010, Cao et al 2010, 2012, Modak et al 2016). In the 60 × BC case, the large decrease in precipitation on fast timescale (−17.8%) is because of (i) the reduction in the downwelling solar flux reaching the surface (−29.08 ± 0.08 W m\(^{-2}\)) due to the increased BC aerosol concentration which blocks the solar radiation—the decreased downwelling solar flux inhibits surface evaporation—leading to a decreased surface latent heat flux (15.94 W m\(^{-2}\) Yoshimori and Broccoli 2008, Ming et al 2010, Ban-Weiss et al 2012), and (ii) the increase in the atmospheric stability due to strong heating at the upper atmosphere because of the SW absorption by the BC aerosols (figure 2(a)); (Ming et al 2010, Frieler et al 2011). We present further discussion on precipitation response in the NH versus SH and the hydrological sensitivity in supplementary section S4.

Summary and discussion

Using a global climate model, we investigate the physical mechanisms associated with the lower efficacy of BC aerosols. Past studies (Cook and Highwood 2004, Roberts and Jones 2004, Hansen et al 2005, Yoshimori and Broccoli 2008, Ban-Weiss et al 2012, 2013, Devaraju et al 2015).
Stjern et al (2017) estimated the efficacy of BC aerosols to be less than one but none of them examined the processes that are associated with the lower efficacy in detail. Most of these studies showed that the efficacy varies with the altitude of BC aerosols—when BC aerosols are closer to the surface, they cause surface warming while at higher altitude they cause surface cooling. In this study, we estimate the efficacy of BC aerosols as 0.65 ± 0.04 when we use the prescribed-SST approach for estimating the ERF and it is 0.53 ± 0.20 when the regression method is used (table 1, figures 3(a) and S15). Therefore, it is not sufficient to compare only the ERF of BC aerosols and CO₂, the efficacy factor should be included to assess the global warming potential of BC aerosols. However, the difference in regional patterns of radiative forcing should be recognized—while the radiative forcing from CO₂ is nearly uniform, BC forcing is highly localized and hence the regional pattern of warming is very different between the two cases (figure 1). Nevertheless, it should be also noted that it is not necessary to have a direct link between the pattern of forcing and the pattern of temperature change (figure 1).

In summary, we show that an increase in BC aerosol mixing ratio by sixty times with the spatial distribution same as the present-day levels causes the Northern Hemisphere (NH) to warm more (∼5 K) than the Southern Hemisphere (SH). This increase in interhemispheric temperature contrast shifts the ITCZ to the north from its mean position concomitantly enhancing the cloud cover over the NH tropics. A shift in the ITCZ corresponds to a shift in the location of Hadley cells. We find that the subsiding branch of the SH Hadley cell shifts into the SH tropics. A moistened boundary layer associated with the subsidence enhances the low cloud cover over the SH tropics. Increased low clouds over both the NH and SH tropics induces a negative SW cloud feedback which causes a lower efficacy of BC aerosols.

What other factors could influence the efficacy? The altitudinal and latitudinal distribution of radiative forcing could influence the efficacy of a forcing agent (Hansen et al 1997, Forster et al 2000, Shindell et al 2001, Kang and Xie 2014, Pithan and Mauritsen 2014). For instance, a forcing at higher latitude causes a larger global mean temperature response compared to low latitude forcing which can be explained by cloud radiative and temperature feedbacks (both Planck and lapse rate feedback) (Kang and Xie 2014). Similarly, a forcing at higher altitude causes a larger fraction of energy loss to space without warming the surface thus leading to less effect on the surface climate change (Hansen et al 1997). In the BC case, we find that the net
radiative forcing is larger at higher altitudes relative to the CO$_2$ case (figure S16) as in the case for Solar and CH$_4$ forcings (Modak et al. 2016, 2018). This indicates that the efficacy of BC forcing would be less than CO$_2$ forcing. A recent observational study (Ditas et al. 2018) showed that the BC aerosols from the wildfires can reach the lower stratosphere and consequently could influence the radiative forcing.

Since our aim in this paper is to understand the fundamental mechanisms associated with the non-unity efficacy of BC aerosols, we have performed highly idealized simulations. Hence, there are some limitations to our study. First, this is a single model study and our results depend on the sign and magnitude of cloud feedbacks. As the estimation of climate sensitivity is sensitive to cloud parameterizations in models (Boucher et al. 2013, Caldwell et al. 2016, Ceppi et al. 2017), further investigations using a multi-model inter-comparison framework is needed for assessing the robustness of our results. Second, we recognize that increasing the BC mixing ratio by 60 times relative to the 2000 climatological level induces a very large forcing and resembles nuclear war scenarios. However, to increase the signal-to-noise ratios and gain clear insight it is important to perform simulations with large forcing as done in a recent study (Kovilakam and Mahajan 2016). Further, we performed several simulations where we increased the BC mixing ratio by 25 times, 45 times, 50 times and 65 times the 2000 climatological level. We find that the global mean surface temperature scales approximately linearly with the increased BC aerosol loading (figure S17). Hence, the results shown in our study could be scaled down linearly to obtain values corresponding to any realistic magnitude of BC aerosol forcing.

Third, in our simulations, only the magnitude of BC forcing distribution is increased by sixty times while keeping the spatial distribution same as in the control simulation (2000 climatology). Hence, the physical mechanism that we present in this study that the differing SW cloud feedback is associated with the lower efficacy of BC aerosols might not hold true with other BC forcing distribution. Hence, the robustness of our result should be assessed by altering the spatial distribution of BC aerosols, for instance, by varying the altitude or the latitude of BC aerosols. Using an earlier version of the model used in this study, Ban-Weiss et al. (2012) investigated the dependence of climate response on the altitude of BC aerosols. We note that in our BC forcing experiment most of the aerosols are confined below 500 hPa which is ~5 km (figure S1a). In Ban-Weiss et al. (2012) when BC aerosols are prescribed at 4 km uniformly around the globe, they find the efficacy to be 0.28. Therefore, we infer that for our design of BC forcing, the efficacy would be less than one which is consistent with our estimation. We decided to keep the same spatial distribution of BC in the 60 × BC case as in the control case because it is more closely related to the present-day BC distribution and hence is more realistic. A systematic investigation of the efficacy by changing the latitude location of BC forcing is beyond the scope of this study and would be the subject for a future study.

Our analysis includes only the aerosol direct (i.e. radiation scattering and absorption by aerosols) and semi-direct effects (i.e. changes in climate due to the local heating from BC aerosols). We do not model the aerosol indirect effects in which particles act as cloud condensation nuclei. As clouds are among the most poorly constrained elements of climate models (Boucher et al. 2013), inclusion of aerosol indirect effects in a study of this type is important for future studies. We also do not include the direct impact of BC aerosols on surface albedo (i.e. deposition of aerosols on the Earth surface) and the cloud absorption effects (i.e. absorption by BC inside the clouds (Jacobson 2012)). Further, this model configuration neglects the fact that different aerosol species are usually internally mixed (Clarke et al. 2004). Inclusion of this could enhance absorption of sunlight by up to a factor of two (Jacobson 2001, 2002). However, we believe that the qualitative fundamental results of our study would be unchanged even after including the above effects of BC aerosols.

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