

Global and regional climate changes due to black carbon

Black carbon in soot is the dominant absorber of visible solar radiation in the atmosphere. Anthropogenic sources of black carbon, although distributed globally, are most concentrated in the tropics where solar irradiance is highest. Black carbon is often transported over long distances, mixing with other aerosols along the way. The aerosol mix can form transcontinental plumes of atmospheric brown clouds, with vertical extents of 3 to 5 km. Because of the combination of high absorption, a regional distribution roughly aligned with solar irradiance, and the capacity to form widespread atmospheric brown clouds in a mixture with other aerosols, emissions of black carbon are the second strongest contribution to current global warming, after carbon dioxide emissions. In the Himalayan region, solar heating from black carbon at high elevations may be just as important as carbon dioxide in the melting of snowpacks and glaciers. The interception of solar radiation by atmospheric brown clouds leads to dimming at the Earth's surface with important implications for the hydrological cycle, and the deposition of black carbon darkens snow and ice surfaces, which can contribute to melting, in particular of Arctic sea ice.

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Black carbon (BC) is an important part of the combustion product commonly referred to as soot¹. BC in indoor environments is largely due to cooking with biofuels such as wood, dung and crop residue. Outdoors, it is due to fossil fuel combustion (diesel and coal), open biomass burning (associated with deforestation and crop residue burning), and cooking with biofuels¹. Soot aerosols absorb and scatter solar radiation. BC refers to the absorbing components of soot, often defined using elemental carbon and some condensed organics². Recent findings suggest other secondary organics also contribute to strong absorption in the ultraviolet region of the spectrum, components that were presumably ignored in the original definition of BC³. Dust, which also absorbs solar radiation, is not included in the definition of BC. Globally, the annual emissions of BC are (for the year 1996) ~8 Tg yr⁻¹ (ref. 4), with about 20% from biofuels, 40% from fossil fuels and 40% from open biomass burning. The uncertainty in the published estimates for BC emissions is a factor of two to five on regional scales and at least ±50% on global scales.

High BC emissions (Fig. 1) occur in both the northern and the Southern Hemisphere, resulting largely from fossil fuel combustion and open burning, respectively. Atmospheric brown clouds (ABCs) are composed of numerous submicrometre aerosols, including BC, but also sulphates, nitrates, fly ash and others. ABCs

have been extensively documented by surface observatories, field observations and satellite data⁵⁻¹⁵. Single-particle mass spectrometer data reveal that BC is internally mixed with other aerosol species such as sulphates, nitrates, organics, dust and sea salt¹⁶. BC is removed from the atmosphere by rain and snowfall². Wet removal as well as direct deposition to the surface limits the atmospheric lifetime of BC to about one (±1) week¹⁷.

REGIONAL HOTSPOTS

Until about the 1950s, North America and Western Europe were the major sources of soot emissions, but now developing nations in the tropics and East Asia are the major source regions^{18,19} (Fig. 1). Historical BC emissions are available for fossil fuel combustion and biofuel cooking^{18,19}. Past emissions of BC from biomass burning are very uncertain¹⁹, although, published reports of extensive brown clouds and their possible effects on the atmosphere date back to at least the 1880s²⁰.

Integration of field observations^{7,14} and new satellite aerosol sensors¹⁵ have revealed the global distribution of ABCs and their radiative forcing²¹⁻²³. Their concentrations peak close to major source regions and give rise to regional hotspots of BC-induced atmospheric solar heating (Fig. 1b). Such hotspots have recently been identified²⁴ as the Indo-Gangetic plains in South Asia; eastern China; most of Southeast Asia including Indonesia; regions of Africa between sub-Saharan and South Africa; Mexico and Central America; and most of Brazil and Peru in South America. Populations of about 3 billion are living under the influence of these regional ABC hotspots.

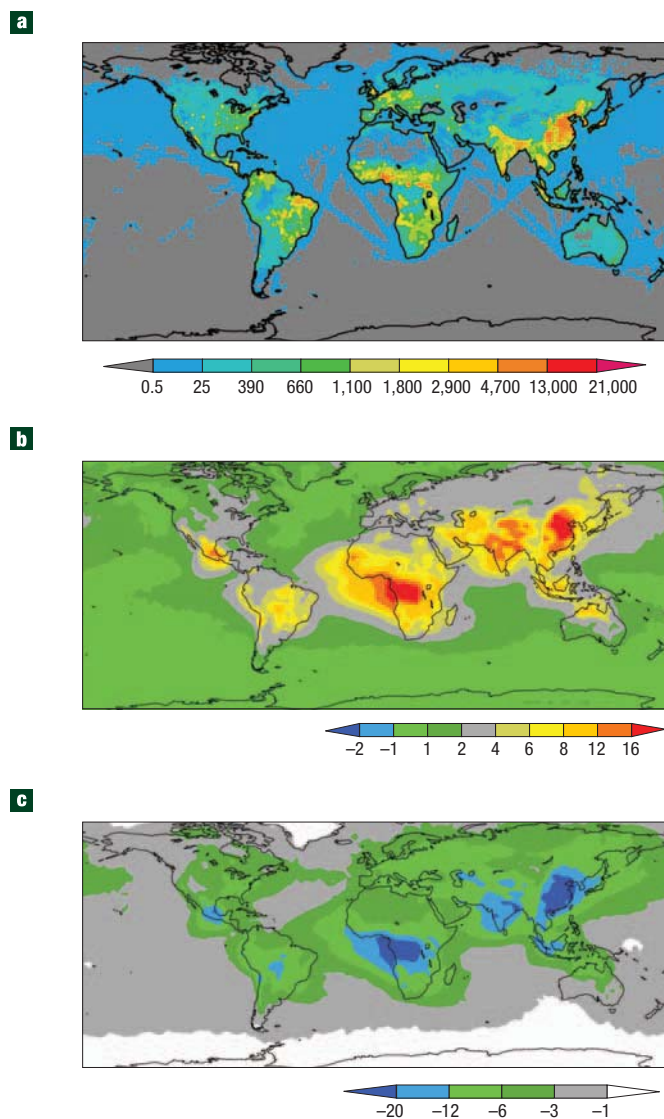


Figure 1 Global distribution of BC sources and radiative forcing. **a**, BC emission strength in tons per year from a study by Bond *et al.*⁴, including emissions from fuel combustion (fossil fuels and biofuels) and open biomass burning (forest fires, savanna burning and outdoor cooking) for the year 1996. The uncertainty in the regional emission is about $\pm 100\%$ or more. **b**, Atmospheric solar heating due to BC from the study by Chung *et al.*²³ for the 2001 to 2003 period. This study integrates satellite aerosol data, surface network of aerosol remote sensing instruments and field observations with an aerosol-transport-chemical model and a radiative transfer model to obtain the forcing. Uncertainty in the forcing is $\pm 30\%$. **c**, As in **b**, but for surface dimming due to ABCs. This shows the reduction in absorbed solar radiation at the surface by all anthropogenic aerosols (BC and non-BC) in ABCs.

RADIATIVE FORCING OF THE CLIMATE SYSTEM

Solar absorption by BC increases inversely with wavelengths from near-infrared (1 μm) to ultraviolet wavelengths with a power law of one to three depending on the source^{3,25}, thus giving the brownish colour to the sky. Unlike the greenhouse effect of CO₂, which leads to a positive radiative forcing of the atmosphere and at the surface²⁶ with moderate latitudinal gradients^{27,28}, black carbon has opposing effects of adding energy to the atmosphere and reducing it at the surface. Furthermore the forcing has

significant latitudinal gradients. It alters the radiative forcing through a complex web of processes⁷.

The first concerns the increase in top-of-the atmosphere (TOA) radiative forcing. This occurs via several pathways: (1) by absorbing the solar radiation reflected by the surface-atmosphere-cloud system, BC reduces the albedo of the planet; (2) soot deposited over snow and sea ice can decrease the surface albedo²⁹⁻³²; (3) soot inside cloud drops and ice crystals can decrease the albedo of clouds by enhancing absorption by droplets and ice crystals³¹⁻³⁴. All three of these processes contribute to a positive TOA forcing. On the other hand, non-BC aerosols such as sulphates, nitrates and organics in ABCs reflect more solar radiation, increasing the albedo of the planet and resulting in a negative TOA forcing. In addition non-BC aerosols also nucleate cloud drops and thus increase the albedo of clouds. This effect is referred to as an indirect effect or 'cloud-albedo effect'³⁵⁻³⁷.

Figure 2 compares the BC forcing (Fig. 2c) with forcing due to all greenhouse gases (GHGs; Fig. 2a), forcing due to CO₂ alone (Fig. 2b) and forcing of all aerosols other than BC (Fig. 2d). The BC forcing includes only the direct forcing from pathway 1 because pathways 2 and 3 are more uncertain and, furthermore, contribute only about 0.1 W m⁻² to the global forcing³³. At the TOA, the ABC (that is, BC + non-BC) forcing of -1.4 W m⁻² (sum of TOA values in Figs 2c,d), which includes a -1 W m⁻² indirect forcing, may have masked as much as 50% ($\pm 25\%$) of the global forcing due to GHGs. The estimated aerosol forcing of -1.4 W m⁻² due to ABCs is within 15% of the aerosol forcing derived in the recent IPCC report³⁷ and is also consistent with other studies^{35,36}.

The BC forcing of 0.9 W m⁻² (with a range of 0.4 to 1.2 W m⁻²) (Fig. 2c) is as much as 55% of the CO₂ forcing and is larger than the forcing due to the other GHGs such as CH₄, CFCs, N₂O or tropospheric ozone³⁷. Similar conclusions regarding the large magnitude of the BC forcing have been inferred by others³⁸⁻⁴¹ and their estimates range from 0.4 W m⁻² to 1.2 W m⁻². The estimate shown in Fig. 2c is obtained from the observationally constrained study of Chung *et al.*²³. Values generated by many general circulation climate models (GCMs) are mostly in the lower range of 0.2 W m⁻² to 0.4 W m⁻² (refs 37,42,43). There are several reasons for the lower estimates. Many ignore the internally mixed state of BC with other aerosols. Such mixing enhances forcing by a factor of two (ref. 39). Field observations have consistently shown that BC is well mixed with sulphates, organics and others^{16,44}. Another factor contributing to lower BC forcing in GCMs is that observed BC concentrations peak at about 2 km above the surface^{7,14,44} whereas, in most models they are concentrated close to the surface⁴⁵. BC at elevated levels enhances solar absorption significantly because it can absorb the solar radiation reflected by the highly reflective low clouds^{38,40,46}. Furthermore, GCMs with lower positive forcing, in general, ignore biomass burning, which is about 40% of the total BC emission. Column integrated aerosol absorption has been retrieved from a worldwide surface network of solar spectral radiometers, referred to as AERONET⁴⁷. The retrieved aerosol absorption^{11,48} is a factor of two or more larger than the GCM simulated values^{41,49}. The exceptions to the low forcing bias of GCMs are the models that constrain aerosol solar absorption with AERONET values⁵⁰ and models that account for the mixing state of BC with other aerosols and include BC from biomass burning^{39,40}. The BC forcings estimated by these models are in the range of 0.6 to 0.8 W m⁻² (refs 39,40) and 0.8 to 1.2 W m⁻² (refs 41,50).

The second process concerns atmospheric solar heating. In addition to absorbing reflected solar radiation, BC absorbs direct solar radiation and together the two processes contribute to a significant enhancement of lower atmosphere (from the surface

to about 3 km altitude) solar heating, by as much as 50% in the hotspots (that is, regions with 15 W m^{-2} forcing) (see Fig. 1b). Direct measurement of this solar heating has evaded us until now as it requires multiple aircraft flying over the same domain at different altitudes to measure flux divergences (that is, heating rates) for an extensive period of time. These challenges were recently overcome by deploying three lightweight unmanned aerial vehicles (UAVs) with well-calibrated and miniaturized instruments to simultaneously measure aerosols, BC and spectral as well as broadband radiation fluxes^{14,51,52}. The UAV study¹⁴ demonstrated that ABCs with a visible absorption optical depth as low as 0.02 are sufficient to enhance solar heating of the lower atmosphere by 50%. Global average BC solar heating of the atmosphere, as per the present estimate, is 2.6 W m^{-2} (Fig. 2c), which is comparable to the TOA GHG forcing (Fig. 2a).

The third process is the surface dimming. The BC absorption of direct solar radiation reduces the solar radiation reaching the surface and leads to dimming (Fig. 2c). The BC dimming is further enhanced by the direct and indirect effects of non-BC aerosols (Fig. 2d). The total dimming effect is -4.4 W m^{-2} (sum of Fig. 2c,d) — about -3.4 W m^{-2} from the direct effect of ABCs and the remaining -1 W m^{-2} from the indirect effect^{35–37}. The dimming can be as large as 5 to 10% over the regional hotspots (Fig. 1). It is important to note that the surface dimming and absorption of direct solar radiation do not contribute much to TOA forcing as it is simply a redistribution of the direct solar radiation between the surface and the atmosphere. However, globally, this redistribution can weaken the radiative–convective coupling of the atmosphere and decrease global mean evaporation and rainfall²⁶.

Is the planet dimmer now than it was during the early twentieth century? Solar radiometers around the world are indicating that surface solar radiation in the extra tropics was lower by as much as 5% to 10% during the mid-twentieth century^{53,54}, whereas in the tropics such dimming trends have been reported to extend into the twenty-first century. But many of these radiometers are close to urban areas and it is unclear if the published trends are representative of true regional to global averages⁵⁵. The Indian Ocean Experiment⁷ used a variety of chemical, physical and optical measurements to convincingly demonstrate that ABCs can lead to dimming as large as 5–10% (Fig. 1c) over widespread regions in the North Indian Ocean and South Asia. In order to get a handle on the global average dimming, Chung *et al.*²³ integrated field observations with satellite data and aerosol transport models to retrieve an observationally constrained estimate. As seen from Fig. 1c, over large regions the reduction of solar absorption at the surface exceeds 10 W m^{-2} (>5%), which is consistent with the dimming reported from surface observations. The global annual average dimming (for 2001–2003), however, is -4.4 W m^{-2} , as opposed to the -10 W m^{-2} estimated by surface radiometers. Thus, great care should be exercised to extrapolate surface measurements over land areas to global averages. The global dimming of -4.4 W m^{-2} has been compared to the GHGs forcing of 3 W m^{-2} from 1850 to present⁵⁴. Such comparisons, without a proper context could be misleading because, as shown in Fig. 2, for BC, the surface forcing is negative whereas the TOA forcing is positive (Fig. 2c).

GLOBAL CLIMATE EFFECTS

The TOA BC forcing implies that BC has a surface warming effect of about 0.5 to 1 °C, where we have assumed a climate sensitivity of 2 to 4 °C for a doubling of CO₂. Conversely, ABCs have a cooling effect of about -0.75 to -2.5 °C (ref. 35). Because BC forcing results in a vertical redistribution of the solar forcing,

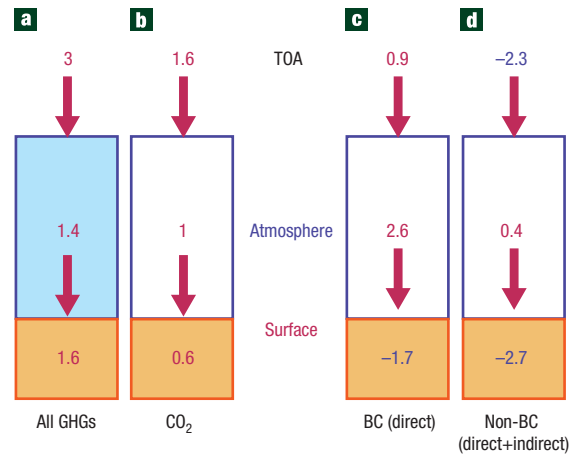


Figure 2 Comparison of the global mean radiative forcing due to greenhouse gases (GHGs) with that of ABCs. **a,b**, Forcing for all GHGs (CO₂, CH₄, N₂O, halons and ozone) (**a**), and for CO₂ (**b**). The number at the top of the atmosphere box (blue box) is the top-of-the atmosphere (TOA) forcing; the number within the atmosphere box is the atmospheric forcing; and the number within the brown box is the forcing at the surface. The TOA forcing is the sum of the forcing of the atmosphere and the surface. The forcing values represent the change in radiative forcing due to increase in gases for the year 2005, which is the same as the forcing from pre-industrial to present. The TOA numbers are taken from ref. 68 and the atmospheric and surface forcings are derived from an atmospheric radiative transfer model. The numbers at the surface and the atmosphere are slightly adjusted to agree with the TOA IPCC forcing. The uncertainty in the forcing values is $\pm 20\%$. **c**, BC forcing obtained by running the Chung *et al.* analysis²³ with and without BC. The forcing values are valid for the 2001–2003 period and have an uncertainty of $\pm 50\%$. **d**, Non-BC forcing. This includes the direct and the indirect forcing. The direct forcing is obtained by subtracting the total anthropogenic forcing in Chung *et al.* from the BC forcing shown in **b**. The indirect forcing (of about 1 W m^{-2} at the TOA and at the surface) is an average of values derived from recent studies^{35–37}.

a simple scaling of the forcing with the CO₂ doubling climate sensitivity parameter may not be appropriate^{40,56,57}. For example, GCMs suggest that the reduction of sea ice and snow albedo by BC is three times as effective as CO₂ forcing for global average surface warming⁵⁷.

BC and non-BC aerosols perturb the hydrological cycle significantly. The surface and atmospheric warming due to GHGs would lead to an increase in atmospheric humidity (owing to an increase in saturation vapour pressure) and rainfall (owing to an increase in the radiative heating at the surface)^{26,58}. With respect to ABCs, the overall negative forcing at the TOA, as well as the surface dimming, should lead to a decrease in evaporation and rainfall^{7,37}. It is difficult to predict the net effect of GHGs and ABCs on global rainfall, given the large positive forcing at the TOA and the larger negative forcing at the surface. We can not resort to observed rainfall trends to infer the net anthropogenic effect on global rainfall as long-term rainfall measurements are only available for land regions.

REGIONAL CLIMATE EFFECTS

We have just begun to comprehend the chain of response and feedbacks on the regional climate due to BC^{9,12,14,23,59–65}. In regions where radiative–convective coupling of the surface and the atmosphere is strong (for example, equatorial oceans and tropical land during wet seasons), the surface–atmosphere response will

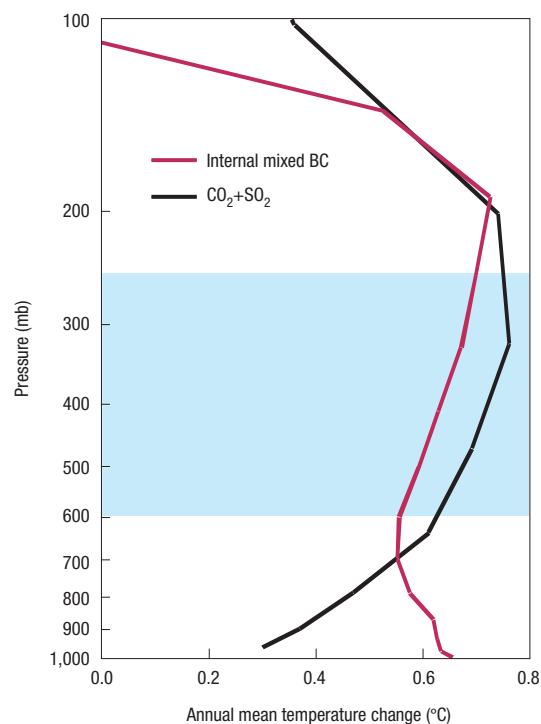


Figure 3 Simulated atmospheric temperature change due to GHGs and BC for the South Asian region. The values are annual mean temperature changes over the South Asian region, averaged from 20° N to 40° N and from 70° E to 100° E. The blue line is the change due to the increase in all GHGs and sulphate aerosols as simulated by ref. 60. The red line is the estimated temperature change due to BC taken from the global circulation model study of Chung and Seinfeld⁶⁷.

be determined by the TOA forcing, and as a result BC by itself will lead to a warming of both the surface (in spite of the surface dimming) and the atmosphere (in spite of the atmospheric solar heating), whereas ABCs will lead to a cooling of both the surface and the atmosphere. In regions where such coupling is weak (for example, dry seasons in the tropics), the surface dimming due to ABCs can lead to surface cooling and thus can mask the greenhouse warming⁶⁶, whereas the atmospheric solar heating by BCs can lead to a warming of the atmosphere and intensify the greenhouse warming of the troposphere. GCMs that include just the BC forcing^{14,64,67} show that BC leads to a warming from the surface to about 12 km altitude, by as much as 0.6 °C over most of the Northern Hemisphere including the Arctic region (for example, see Fig. 11 in Chung and Seinfeld⁴⁰, and ref. 64). The magnitude of the BC atmospheric warming is comparable to the simulated warming due to GHGs forcing⁶⁸. Regionally, the combined effect of ABCs is to cause a surface cooling⁶⁵ over South Asia while warming the atmosphere by as much as 0.6 °C during winter and spring^{14,60}. Such differential warming of the atmosphere with respect to the surface over the South Asian region has also been observed with microwave satellite sensor observations of the trends from 1979 to 2003^{14,60}.

BC atmospheric heating may be an important contributing factor to the retreat of Himalayan glaciers. Analysis of temperature trends on the Tibetan side of the Himalayas reveals warming in excess of 1 °C since the 1950s. This large warming trend at the elevated levels is proposed as the causal factor for the retreat of glaciers through melting^{69,70}. GCM simulations suggest that advection of the warmer air heated by BC from South

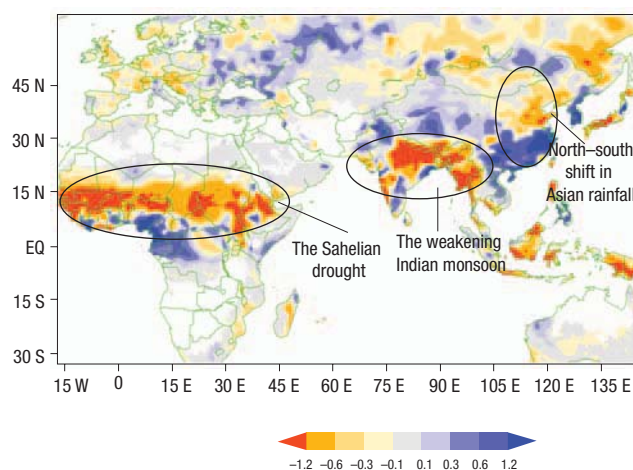


Figure 4 Precipitation trend from 1950–2002. The plot is adopted from ref. 65. (units: change in mm per day from 1950 to 2002). The red and dark blue shaded regions are statistically significant.

and East Asia over the Himalayas contributes to a warming of about 0.6 °C (annual mean) in the lower and mid troposphere (see Fig. 3) of the Himalayan region^{14,64}. This is as large as the warming trend due to the GHGs (Fig. 3), leading to the inference¹⁴ that BC forcing is as important as GHGs in the observed retreat of over two thirds of the Himalayan glaciers⁷¹.

BC contributes to melting of snow through another process. When soot is deposited over snow and sea ice, it darkens the snow and significantly enhances solar absorption by snow and ice^{30,32}. Recent studies suggest that this is one of the important contributors to the retreat of the Arctic sea ice (see summary of earlier studies in ref. 57). Simulations by Flanner *et al.*⁵⁷ showed that the deposition of BC from sources in North America and Europe over the Arctic sea ice may have resulted in an Arctic surface warming trend of as much as 0.5 to 1 °C (ref. 72). In addition, the study estimated that BC-induced reduction of snow albedo is a major forcing term (about 20 W m⁻²) in the Tibetan side of the Himalayas. Ice-core records of BC deposition over Greenland from the early nineteenth century onwards have now provided a historical record for examining the role of BC forcing in the retreat of sea ice⁷³.

Atmospheric heating and dimming by BC and non-BC aerosols can perturb the monsoon significantly. Precipitation trends over many regions of the tropics during the last 50 years have been negative, particularly over Africa, South Asia and northern China (Fig. 4)⁶⁸. These drying patterns can not be explained solely by global warming^{74,75}. Natural variability and anthropogenic aerosol forcing are emerging as major players in the observed trends^{60,74–76}. The impacts of ABCs and BC on the South Asian monsoon have received attention recently^{40,59,60,62,63,65,67,74}. Precipitation over land is driven by evaporation from the land surface and long-range transport of moisture from the surrounding Indian Ocean. These model studies reveal that ABCs have three distinct effects on the long-range transport of moisture and its convergence over South Asia:

(i) A decrease in the evaporation of the Indian Ocean owing to dimming. Emissions of BC and other aerosol precursors from South Asia have increased significantly since 1950s^{18,19}. This has resulted in a dimming trend of about 7% as detected by surface radiometers in India⁶⁰. Similar dimming has also occurred over the Indian Ocean⁷ (See Fig. 1c). As about 75% or more of the surface radiative heating is balanced by evaporation²⁶, the

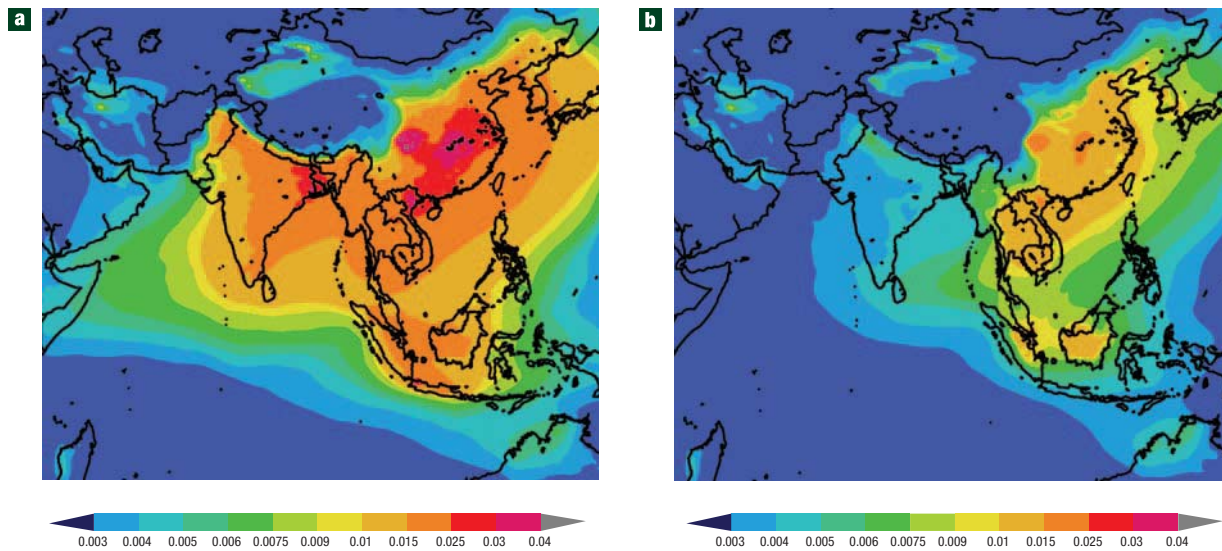


Figure 5 The effect of biofuel cooking on Asian BC loading. **a**, The simulated annual mean optical depth of BC aerosols for 2004–2005 using the regional aerosol/chemical/transport model described in ref. 96. The values include BC emissions from biofuel cooking (indoor cooking with wood/dung/crop residues), fossil fuels and biomass burning. **b**, As for **a**, but without biofuel cooking.

dimming trend leads to a decrease in evaporation from the North Indian Ocean⁶⁰, so less moisture is fed to the monsoonal inflow into South Asia.

(ii) A decrease in meridional sea surface temperature (SST) gradient. Because ABCs are concentrated over the North Indian Ocean (Fig 1), the dimming is suppressing the greenhouse warming over the North Indian Ocean while the GHGs warming is proceeding unabated over the southern Indian Ocean. As a result, the summertime north-to-south SST gradient (with warmer waters over the North Indian Ocean) has decreased since the 1950s, as has been seen from observations^{60,74}. The weakening of the SST gradient weakens the monsoonal circulation, as shown by numerous studies^{60,74,75}, and in turn weakens the monsoonal rainfall during summertime. It is important to note that, although the ABC dimming peaks in winter and spring, the SST response is delayed until summer owing to the slower response time of the ocean^{60,64,74}.

(iii) An increase in atmospheric meridional heating gradient. The stronger BC solar heating of the atmosphere over South Asia (Fig. 1b) strengthens the monsoonal outflow with stronger rising motions over the subcontinent, accompanied by a stronger moisture flux into South Asia^{60,63,64,74}. This effect, which increases rainfall, peaks during spring when the BC heating is at its peak value.

The atmospheric heating shown in Fig. 1b is solely due to BC, whereas the dimming is due to both the BCs and non-BC aerosols in ABCs (Figs 1c and 2d). The larger dimming over the land regions compared with the adjacent oceans also suggest that the dimming decreases the land–sea contrast in surface temperature — a major monsoon forcing term. In order to account for the delayed oceanic response to the dimming, fully coupled ocean–atmosphere models are required. To date, three such studies have been published^{60,62,64} and all of them estimate an increase in pre-monsoon rainfall during spring followed by a decrease in summer monsoon rainfall, in agreement with observed trends (Fig. 4; ref. 60). The link between dimming, the north–south SST gradient and a decrease in land rainfall has also been invoked to explain the Sahel drought⁷⁵ of the 1970s and 1980s.

CLIMATE SYSTEM RESPONSE AND FEEDBACKS

The immediate response of the atmosphere to ABCs is to increase or decrease cloud cover. The non-BC aerosols, by nucleating more cloud drops, decrease the effective radius of the cloud drop. This can suppress formation of larger drizzle drops, extend the lifetime of clouds, and thus increase cloud cover³⁷. On the other hand, BC solar heating can decrease the relative humidity of the cloud layer, leading to evaporation of cloud drops and thus decreasing low cloud fraction and albedo. This semi-direct effect can enhance the positive climate forcing by BC³⁷. A relative comparison of the increase in cloud cover by non-BC aerosols and the decrease due to BC was undertaken⁷⁷ empirically with a surface network of sun photometers⁴⁷. This study suggested that the non-BC effects dominate overall, except for in heavily polluted regions with absorption optical depths exceeding 0.05 (for example, the Amazon during the burning season; Africa during Savanna burning season; and urban regions in South and East Asia). An alternative scenario is that BC solar heating induces convection and consequently leads to cloud formation⁷⁸. The global magnitude of the semi-direct effect is highly uncertain.

Spring season dust storms from Asia and Africa transport large quantities of dust across the Pacific Ocean^{79,80} and the Atlantic Ocean⁸¹. The dust is transported either as individual layers or mixed with industrial soot. Such dust–soot mixtures increase the atmospheric solar heating and surface dimming significantly^{79,80} and can also serve as nuclei for ice clouds and feedback on precipitation⁸². For the first time, such dust–soot mixtures were tracked in an aircraft all the way across the Pacific Ocean from near the surface to about 14 km altitude⁸³.

An increase in drought intensity due to global warming can intensify occurrence of forest fires, as has been documented for California⁸⁴. Increase in forest fires, such as the boreal forest fires of 1996, can increase soot deposition in sea ice and enhance its melting⁵⁷. Surface cooling occurring simultaneously with lower atmosphere warming (due to BC and dust) can stabilize the boundary layer during the dry season and increase the lifetimes of aerosols in ABCs and increase persistence of soot-filled fog. Soot can also influence precipitation formation mechanisms^{85,86}.

Two extreme scenarios have been proposed for such feedbacks. For South Asia, GCM simulations suggest that a two- to threefold increase in soot loading (from present day levels) is sufficient to substantially weaken the monsoon circulation, decrease rainfall by more than 25% and increase drought frequency significantly⁵⁹. As wash out by rain is a major sink for BC, large decreases in rainfall can have a positive feedback on BC concentrations. The other scenario is the so-called nuclear winter scenario^{87–89}, in which large-scale increase in BC from fires resulting from a global-scale nuclear war can nearly shut down sunlight at the ground (total dimming), which can collapse the troposphere and decrease rainfall drastically.

REDUCING FUTURE BLACK CARBON EMISSIONS

Given that BC has a significant contribution to global radiative forcing, and a much shorter lifetime compared with CO₂ (which has a lifetime of 100 years or more), a major focus on decreasing BC emissions offers an opportunity to mitigate the effects of global warming trends in the short term (see, for example, refs 90–92). Reductions in BC are also warranted from considerations of regional climate change and human health^{93,94}.

It is clear from Fig. 2 that air pollution mitigation steps can have significant impacts on future climate changes. The logical deduction from Fig. 2a,c,d is that the elimination of present day ABCs through emission reduction strategies would intensify surface warming by about 0.4 to 2.4 °C (see also, ref. 35). If only the non-BC aerosols were controlled, it could potentially add 2.3 W m⁻² to the TOA forcing and push the system closer to the 3 °C cumulative warming (since 1850s), which is a likely threshold for unprecedented climate change⁹⁵. If on the other hand, the immediate target for control shifts entirely to BC (owing to its health impacts) without a reduction in non-BC aerosols, the elimination of the positive forcing by BC will decrease both the global warming and the retreat of sea ice and glaciers. It is important to emphasize that BC reduction can only help delay and not prevent unprecedented climate changes due to CO₂ emissions.

ASIAN EMISSIONS AND FUTURE TRENDS

Given the fact that technology exists for large reductions of soot emissions, we explore the impact of a major focus on soot reductions. We focus on Asia, where emissions from China and India alone account for ~25 to 35% of global BC emissions and the regional climate responses to BC are (expected to be) large. In addition, with the economies of China and India expanding with double digit growth rates, Asia can become a much larger source of ABCs, depending on the energy path taken to sustain this growth rate. In fact new estimates indicate that BC emissions for China in 2006 have doubled since 2000, whereas SO₂ emissions have grown during this period by more than 50% (D. G. Streets, manuscript in preparation, data available at http://www.cgrrer.uiowa.edu/EMISSION_DATA_new/summary_of_changes.html). East Asia and South Asia also represent a different mix of emissions, and therefore can illustrate potentials for various control options that are also representative of global choices. The majority of soot emission in South Asia is due to biofuel cooking, whereas in East Asia, coal combustion for residential and industrial uses plays a larger role. The large BC emissions are reflected in the geographical extent of the large absorbing component of aerosol optical depth, simulated with a regional aerosol-chemistry transport model⁹⁶ (see areas with BC optical depth > 0.01 in Fig. 5a).

What are the opportunities to reduce the positive forcing by BC? Providing alternative energy-efficient and smoke-free cookers and introducing transferring technology for reducing

soot emissions from coal combustion in small industries could have major impacts on the radiative forcing due to soot⁹⁷. Figure 4b shows the impact of replacing biofuel cooking with BC-free cookers (solar and bio and natural gas) in South and East Asia. The impacts are dramatic: over South Asia, a 70 to 80% reduction in BC heating; and in East Asia, a 20 to 40% reduction. The impact on human health will potentially be even more dramatic as over 400,000 annual fatalities among women and children are attributed to smoke inhalation during indoor cooking^{93,94}. However, changes in BC alone do not tell the entire story as the climate response also depends on how the BC to non-BC aerosol fraction responds to future emissions. As BC is co-emitted with non-BC aerosols, it is necessary to evaluate how various mitigation strategies impact this fraction. With an emphasis on the opportunities discussed here, this ratio would probably decrease in the future, more quickly in East Asia, amplifying the effectiveness of BC reductions⁹⁸.

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