MIT Joint Program on the Science and Policy of Global Change

A Modeling Study on the Climate Impacts of Black Carbon Aerosols

Chien Wang

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Chien Wang

Abstract

The role of black carbon (BC) aerosols in climate change is important because of its strong capability in causing extinction of solar radiation. A three-dimensional interactive aerosol-climate model has been used to study the climatic impact of BC. The interannual variations of BC solar forcing derived from 20-year transient integrations are up to 4 times as large as the means mainly related to changes in cloud cover, snow depth (about ±20% over many high- or even mid-latitude regions in Northern Hemisphere) and thus the surface albedo, all caused by BC solar forcing itself. With an absolute amount three times higher than that of the top of the atmosphere (TOA) forcing, the surface forcing of BC is an extremely important factor in analyzing the climate impact of BC. BC aerosols cause a "cloud burning" effect in several polluted regions and a "cloud enhancing" effect in some high-latitude sites. Combined with BC-caused changes in surface albedo, this is defined as a non-Twomey-Albrecht indirect forcing by BC, which alters the radiative budgets by changing cloud cover and some land-surface properties thermodynamically rather than microphysically. The result of this study does not indicate that BC aerosols contribute to a significant increase in land-surface temperature with annual emissions of 14 TgC. The calculated surface temperature change is determined by a subtle balance among changes in surface energy sources and sinks as well as changes in the hydrological cycle, all caused by BC radiative forcing. The result of this study shows that the influence of BC aerosols on climate and environment at the regional scale is more significant than at the global scale. Several important feedbacks between BC radiative effect and climate dynamics revealed in this study suggest the importance of using interactive aerosol-climate models to address the issues related to the climate impacts of aerosols.

Contents

1. INTRODUCTION

Direct radiative forcing of black carbon (BC) aerosols at the top of atmosphere (TOA) has been calculated in various studies and the estimated values range from about +0.16 to +0.80 W/m^2, depending on the treatment of their mixing states (*e.g.*, Haywood and Shine, 1995; Haywood *et al*., 1997; Haywood and Ramaswamy, 1998; Penner *et al*., 1998; Myhre *et al*., 1998; Cooke *et al*., 1999; Tegen *et al*., 2001; Jacobson, 2001). This forcing represents a considerable amount of heating of the atmosphere and has been conjectured as a potential factor causing global warming (*e.g.*, Hansen *et al*., 1998; Jacobson, 2001). However, strong extinction of solar radiation by means of both scattering and absorbing effects of BC leads to a reduction of incoming solar radiation at the Earth's surface, as revealed by observations (*e.g.*, Satheesh and Ramanathan, 2000; Ramanathan *et al*., 2001) or suggested by theoretical studies (*e.g.*, Hansen *et al*., 1998; Haywood and Ramaswamy, 1998), and hence, like sulfate aerosols, introduces a cooling effect to the surface.

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It is obvious that the changes in surface energy budget caused by the direct forcings of aerosols can be quite complicated. The balance between altered surface radiative and heat fluxes caused by aerosols determines the actual impact of aerosols on surface temperature changes. Previous modeling studies suggests that at the current time, the cooling caused by sulfate aerosols represents a significant offset in value to the warming effect caused by so-called greenhouse gases such as CO_2, CH_4, N_2O, and O_3 (Roeckner *et al*., 1999; Ramaswamy *et al*., 2001). In the case of BC aerosols, however, the uneven heating/cooling effect across the atmosphere-Earth's surface interface further complicate the surface energy budget and thus the otherwise near constant relationship between the radiative forcing of a given species and the resultant change in surface temperature established for other radiatively important species (*e.g.*, Manabe and Wetherald, 1967).

Interestingly enough, these BC aerosol induced changes in radiative or heat fluxes and resultant changes in surface temperature can potentially perturb clouds and precipitation, and thus eventually impact on the actual effect of BC's radiative forcing as well as BC's lifetime in troposphere, as the atmospheric abundance of BC is primarily influenced by precipitation (*e.g.*, Liousse *et al*., 1996; Cooke *et al*., 1999). It is clear that the above-mentioned processes can form a two-way feedback mechanism between radiative forcings of BC aerosols and climate dynamics that is essential in determining the climate impact of these aerosols. The close interaction between BC aerosols and climate suggests the importance of using an interactive aerosol-climate model, a model including the transient two-way feedback mechanism between BC aerosols and climate dynamics as well as physics, to investigate the climate impact of aerosols. Unfortunately, due to its complexity, such analyses are still rare for BC aerosols.

This article presents a study using a three-dimensional interactive aerosol-climate system model developed based on the Climate System Model (CSM, see Boville and Gent, 1998) of the National Center for Atmospheric Research (NCAR) to explore the climate impact of BC as an additional component to atmospheric aerosols. The model description is given first, followed by discussion of the modeled results including radiative forcings of BC aerosols as well as their influences on climate dynamics and physics. The principal conclusions are then summarized.

2. MODEL DESCRIPTION

Black carbon aerosols are treated as externally mixed aerosols in a single mode with a given size spectrum in this model (*e.g.*, Wilson *et al*., 2002). Like the majority of current BC models summarized in Penner *et al*. (2001), a single-moment scheme that uses mass mixing ratio as the only prognostic variable to describe atmospheric evolution of BC aerosols is used. The atmospheric transport of aerosols is calculated by the same procedure as other tracers such as water vapor in the NCAR Community Climate Model version 3 (CCM3) at each time step (Kiehl *et al*., 1998). Dry deposition and gravitational sedimentation of aerosols are included. The bulk dry deposition speed of BC aerosols is given to be 0.1 cm/s, adopted from several previous studies (*e.g.*, Cooke and Wilson, 1996; Liousse *et al*., 1996). The gravitational sedimentation speed is given as the dry deposition speed multiplied by an approximated pressure correction factor of $(p_{00}/p)^{0.286}$ (Twomey, 1977a; Wang and Chang, 1993), here p_{00} and p are the sea level air pressure and the air pressure at a given vertical layer, respectively. Wet scavenging of aerosols in the model includes liquid droplet-aerosol and ice particle-aerosol processes based on the impact scavenging calculation (Wang and Chang, 1993; Wang and Prinn, 2000), with an approximate

collection coefficient of 0.01 for liquid droplet and 0.005 for ice particles (Pruppacher and Klett, 1997). The nucleation scavenging of BC aerosols are not included (*e.g.*, Flossmann *et al*., 1985). The scavenging rate at a given vertical layer of the model is thus derived as:

$$\frac{1}{q_a}\frac{dq_a}{dt} = C\left(p_{00}/p\right)^{0.286} q_p \qquad (1)$$

Here q_a and q_p are mass mixing ratios of BC aerosol and precipitation particles in kg/kg, respectively; *dt* has an unit of second. For the liquid precipitation scavenging where the temperature is above freezing point, the value of *C* (in kg kg^{-1} second^{-1}) is 4.51×10^{-2} for convective clouds, and 5.18×10^{-2} for stratoform clouds. The value of *C* for ice particle scavenging (assumed to happen when temperature is below freezing point) is calculated based on snowflakes (Wang and Chang, 1993) and appears to be quite close to the *C* value for stratoform clouds multiplied by 0.5 (*i.e.*, 2.59×10^{-2}). To simplify the coding in the program, this value is thus used for the ice particle scavenging calculation. In the model calculation, q_p at a given layer is actually an accumulation from the cloud top to this layer with deduction of evaporation. This is because during the time period between two adjacent wet scavenging calculations (2400 seconds for the T42 run), precipitation particles have a good chance to fall through the clouds. The tropospheric lifetime of BC aerosols yielded using this scheme is about 4.6 days, very close to the result of Liousse *et al*. (1996) (4 to 4.5 days) and Cooke *et al*. (1999) (5.8 days), while it appears to be shorter than that obtained in Cooke and Wilson (1996) for a fossil fuel only calculation (7.85 days). Obviously, implementing microphysical processes into a general circulation model (GCM) with a several hundreds kilometer resolution is "dangerous" work. On-going study incorporating surface property transitions of aerosols due to chemical reactions and microphysics as well as research using a high resolution model with comprehensive physics and chemistry to derive parameterization for GCM and to assess the uncertainty of this type of approximation will further address this issue.

The inclusion of aerosols in the radiation is formulated following Kiehl and Briegleb (1993), and three sets of optical parameters of black carbon aerosols are derived based on previous work in order to investigate the influence of different size spectrum or scattering calculation on the climate impact of BC. The first set of parameters is derived based on Jennings and Pinnick (1980), Twitty and Weinman (1971), and Liousse *et al*. (1996). The second set is adopted from Haywood and Shine (1995), where a lognormal size distribution is applied. The third set is derived based on a revised gamma distribution (*e.g.*, Twitty and Weinman, 1971) and the Mie scattering scheme for spherical particle described in Bohren and Huffman (1983). The values of these parameters for the 0.35 to 0.64 μm waveband are listed in **Table 1**. Model runs using parameter set 1, 2, and 3 hereafter are referred as Model 1, 2, and 3, respectively. The Model 1

Table 1. Optical parameters of BC aerosols for the waveband 0.35 to 0.64 μm used in the model.

	Specific extinction coefficient (m^2/g)	**Single scattering albedo**	**Asymmetry factor**
Model 1	7.93	0.26	0.75
Model 2	12.50	0.25	0.42
Model 3	15.37	0.40	0.42

results will be emphasized in discussions (unless indicated otherwise) in order to reduce the size of this manuscript. By including the aerosol radiative effects into prognostic procedures in the CCM3 radiation module, the aerosols are allowed to interact with climate dynamics when that option is chosen.

The emissions of black carbon from both fossil fuel and biomass burning are used in this study (**Figure 1**). The fossil fuel emissions of BC are derived based on (a) the energy use for coal, gas, and oil specified in the MIT Emissions and Policy Prediction Analysis Model (EPPA) (see Prinn *et al*., 1999); (b) consumption of refined petroleum products from the International Energy Agency (IEA) (http://www.iea.org); and (c) BC and organic carbon emission coefficients for different fuel types (coal, refined oil, and gas) and different sources from Cooke *et al*. (1999). The biomass burning emissions of BC are derived based on Crutzen and Andreae (1990) and Liousse *et al*. (1996). The emissions are mapped into a 1°×1° global grid first. Depending on the source of the emissions, a population density map (Li, 1996) for emissions from fossil fuel combustion and biomass burning in households or the land use maps (Olivier *et al*., 1995) for biomass burning other than the households kind are employed. Details in deriving these emissions are described in Mayer *et al*. (2000). Then, the emissions data for different model resolutions are derived based on this 1°×1° dataset. The annual BC emissions used in this study is about 14 TgC, partitioned as ~8 TgC from fossil fuel burning and ~6 TgC from biomass burning.

In this study, comparison of two groups of different model runs, one (REF) which excludes and the other (BCRAD, including three runs) which includes the radiative effect of BC as well as associated influences on climate dynamics, isolates the climate effect of BC. To further concentrate on this, a constant "background" sulfate aerosol field that comes with the CCM3 distribution, instead of a dynamically changed one, is used in the model. Note although this is consistent with the external mixing assumption made in this study, the possible influence of this setup on the modeled results might need to be assessed when a more comprehensive model

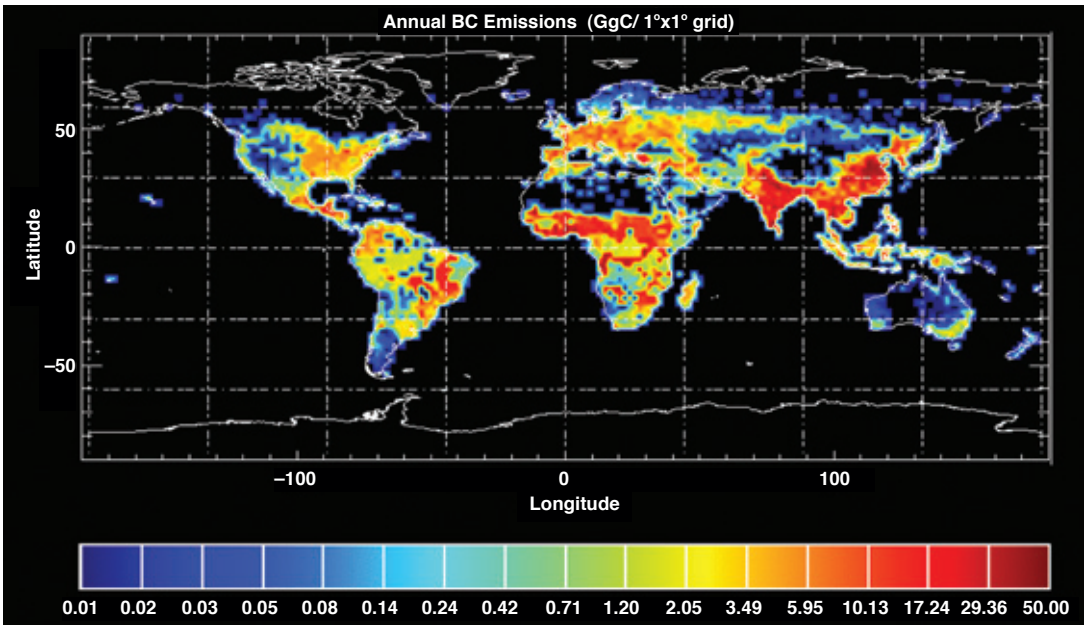

Figure 1. Annual emissions of black carbon aerosols used in the study. The unit is GgC per 1°×1° grid.

that includes internal mixtures of aerosols becomes available. Integration times of both groups of model runs (four in total: one REF and three BCRADs) are 20 years (from 1978 to 1998) in order to take into account the influence of model variability on the results. The BC annual emissions are kept constant through the model integrations in order to emphasize the study of the feedbacks between BC radiative forcings and climate dynamics. To shorten the required computing time of the model runs and to take advantage of the fact that the direct influence of BC on surface temperature is much more significant over the land than the ocean, only the atmospheric model, CCM3 (Kiehl *et al*., 1998) and the Land Surface Model (LSM; Bonan, 1998) in the CSM model family are used. The sea surface temperatures (SST) are prescribed using an analyzed monthly-mean SST dataset (http://www.cgd.ucar.edu/cms/ccm3) that covers the same 20-year time period as the model integrations. The horizontal resolution of the model used in this study is $2.8^{o} \times 2.8^{o}$ (T42). There are eighteen vertical layers. The entire model integrations were processed in a 32-node PC cluster at MIT.

The modeled highest atmospheric loading of BC appears in East Asia, primarily along the eastern part of China (**Figure 2a**). Other areas with high loading of BC aerosols include the Indian subcontinent, eastern and central U.S., Europe, central Africa, and Brazil. Interestingly, model results clearly indicate two transport pathways of BC aerosols, one from East Asia to North America and the other from Europe to the northern Polar Regions. Although the lifetime of black carbon aerosols in the atmosphere is rather short (~5 days in this model), it can be lengthened in the upper troposphere. Model results indicate that in the middle and upper troposphere a considerable amount of BC aerosols that originated from Europe and East Asia have been transported toward higher altitudes and over the northern oceans (**Figure 2b**). The appearance of BC aerosols in these high-latitude areas are relatively persistent due to lower wet scavenging rate compared to the mid- and low latitudes. The modeled characteristics of BC spatial distributions are well supported by the satellite observational data of small particles with biomass or human pollution origins such as the "aerosol index" (AI) derived from the measurements of the POLDER instruments on board the ADEOS satellite shown in **Figure 2c** (Bréon *et al*., 2002). A good correlation between modeled column loading or surface concentrations of BC and satellite AI data, especially for the land regions where the high AI is located, is also found in comparing these two parameters point by point over the model grid (*i.e*., T42, not shown).

In addition to the comparison with satellite data, the modeled surface BC concentrations have also been compared with (somewhat limited) surface observations including point data summarized in Liousse *et al*. (1996) and Cooke *et al*. (1999) (**Figure 3**) and several station data (**Figure 4**). The agreement between modeled surface concentrations and the point data is generally within a factor of 2 except for some oceanic or remote sites, comparable to other nine models reviewed by Penner *et al*. (2001). The modeled results generally agree well with the Barrow and Amsterdam Island station data in both seasonal cycle and concentrations (note that constant emission is used in this study). However, an overestimation in modeled Mauna Loa surface concentration along with a slight switch in seasonal cycle can be clearly seen in the figure. Interestingly, a similar disagreement between modeled and observed results for the Mauna Loa station can be also found in Cooke *et al*. (1999). Therefore, this might suggest a necessary further adjustment in East Asian emissions (*e.g*., Streets *et al*., 2001). The total BC atmospheric burden averaged over 20 years of model integration period is about 0.17 TgC with

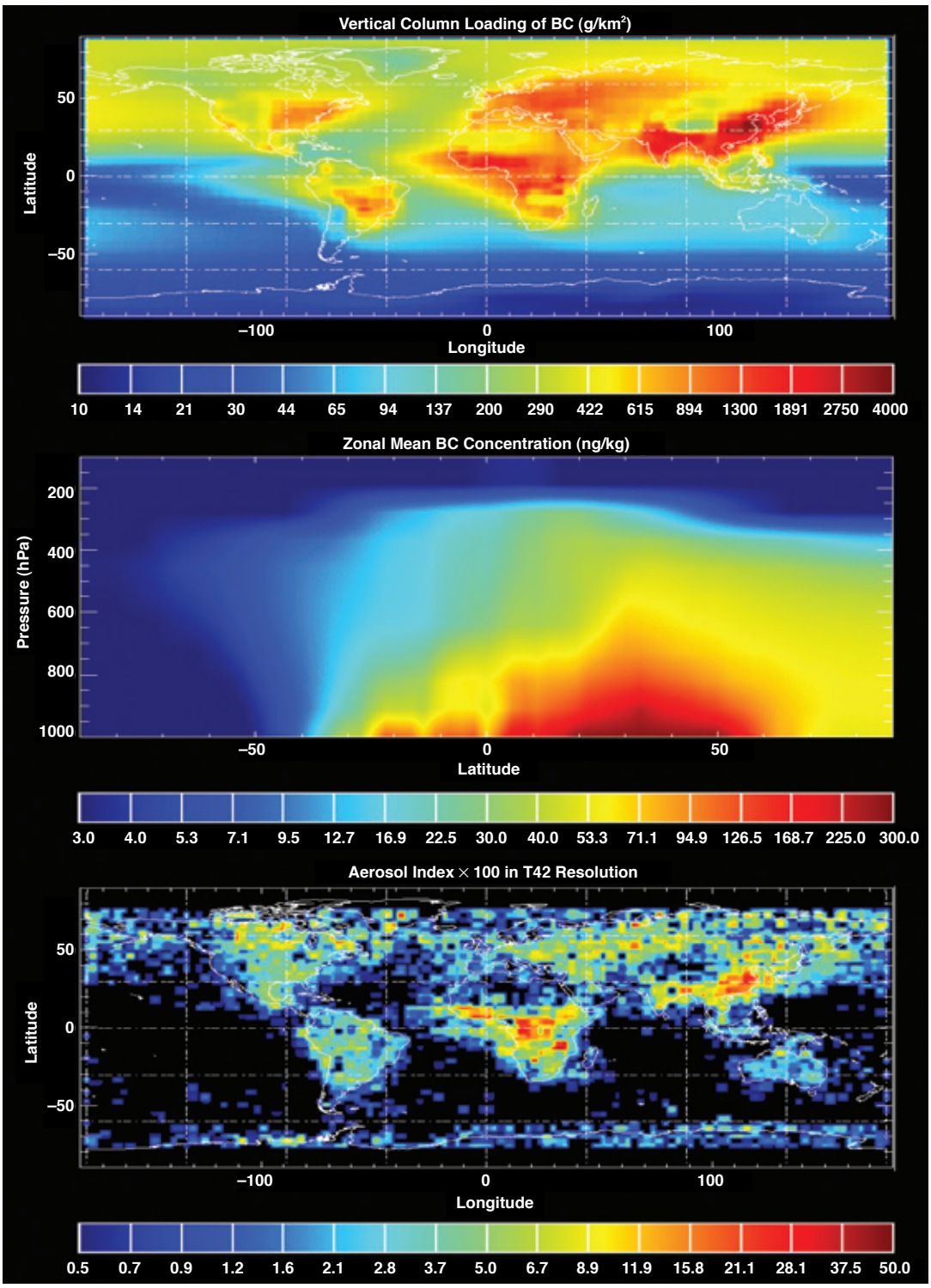

Figure 2. (a) Model simulated vertical column loading of black carbon aerosols (*upper panel*), and **(b)** zonal mean distribution of BC aerosol concentration (*middle panel*), both are 20-year averages of monthly means; and **(c)** aerosol index derived from satellite observations (*lower panel*). The monthly POLDER satellite AI data (kindly provided by F.-M. Bréon and S. Generoso) cover a time period from November 1996 to June 1997. These data are averaged and interpolated into the T42 grid (2.8°×2.8°) used in the model.

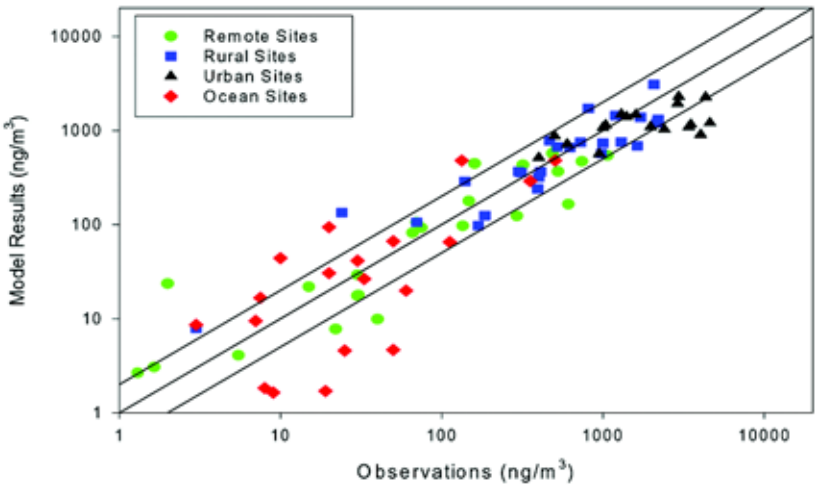

Figure 3. Comparison between observed and modeled surface black carbon concentrations at a number of locations. Observational data are from Liousse *et al*. (1996) and Cooke *et al*. (1999).

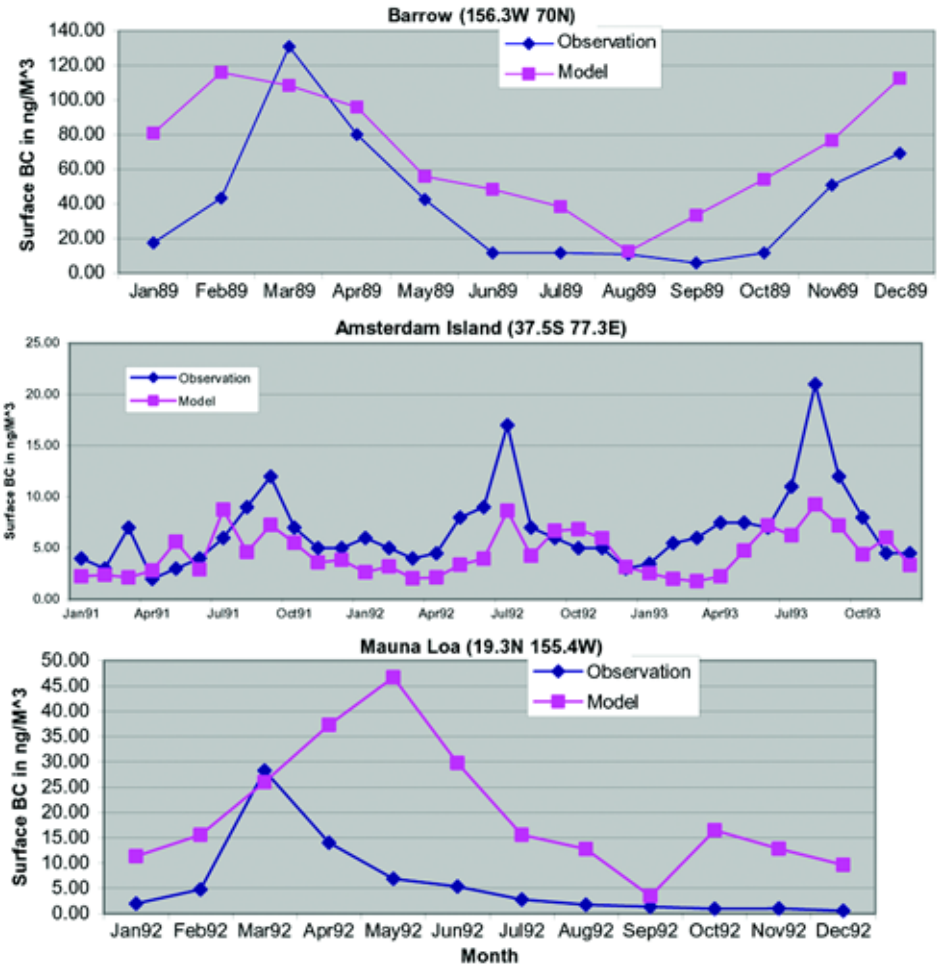

Figure 4. Comparison between modeled and observed surface concentrations of BC aerosols at three stations: **(a)** Barrow, **(b)** Amsterdam Island, and **(c)** Mauna Loa. Observational data of Barrow and Mauna Loa site are from Bodhaine (1995), the Amsterdam Island data are from Liousse *et al*. (1996).

interannual variations smaller than ±2%. It appears to be smaller than the central value (0.133 TgC from biomass burning plus 0.133 TgC from fossil fuel—note the simple addition might not be valid here due to nonlinear precipitation scavenging) obtained in the IPCC model intercomparison (Penner *et al*., 2001), however, it is still within the range of commonly estimated values (*i.e*., 0.15 to 0.25 TgC, Cooke *et al*., 1999).

3. SOLAR RADIATIVE FORCING CAUSED BY BLACK CARBON

Four types of radiative forcings caused by black carbon, namely forcings at the TOA and at the surface, both for all sky (including clouds) and clear sky, have been derived by subtracting the results of the REF run from those of the BCRAD runs:

$$\Delta F_{BC}(x, y, h, t) = \Delta F_{BCRAD}(x, y, h, t) - \Delta F_{REF}(x, y, h, t) \quad (2)$$

Here, x, y, and t are indices of longitude, latitude, and time, respectively; h represents either the TOA or the Earth's surface; ΔF_{BC} represents the radiative forcing caused by BC; the two ΔF terms on the right hand side of the equation are the net downward fluxes of solar radiation in the BCRAD and REF runs, respectively. The global mean radiative forcings by BC appear to be positive at the TOA (warms the atmosphere) but negative at the Earth's surface (cools the ground). In addition, the magnitudes of the surface forcing are about 3 times larger than those at the TOA. This ratio is nearly constant throughout the whole 20 years of integration time. This result is consistent with an estimate based on local direct observations (Satheesh and Ramanathan, 2000). The 20-year means of BC solar forcing using Model 1 are 0.27 and 0.22 W/m^2 at the TOA for clear sky and all sky and –0.57 and –0.64 W/m^2 at the surface for clear sky and all sky, respectively. The TOA forcings of two other model runs are quite close to the Model 1 results, while the surface forcings differ from the Model 1 results slightly more (**Table 2**).

These forcings (even as annual means) appear to have large ranges, *e.g.*, in the Model 1 run, 0.28 to 0.85 W/m^2 at the TOA for clear and all sky, and 0.24 to 0.98 W/m^2 at the surface for clear sky and all sky, with corresponding minima/maxima of 0.12/0.40, –0.27/0.58, –0.73/–0.48, and –1.21/–0.23, respectively (**Figure 5** and **Table 3**), implying the possible error ranges of a "snapshot" estimate of BC radiative forcing. Note the interannual variation of BC atmospheric burden is too small to cause these perturbations in BC radiative forcings. It has been found that the interannual variation of clear sky forcing is primarily related to the surface albedo changes caused by BC aerosol forcing through altering the surface snow depth, mainly in mid- and high-latitudes of the Northern Hemisphere (**Figure 6** and **7**). Modeled results have shown that BC aerosols cause decreases in surface snow depth as large as 15 mm over a vast area of mid- and high-latitude lands, with exceptional increases in high latitudes of North America, west Greenland, and several sites over the Eurasian continent, most related to relatively wet regions. These changes in snow depth can be caused by BC aerosols through several different processes, including atmospheric heating that causes clouds to dissipate or precipitation to decrease, and the reduction of solar radiation at the surface to preserve the snow layer or the adiabatic cooling or heating at the boundary layer depending on the vertical locations of aerosol layers (*e.g.*, Hansen *et al*., 1997) that can impact the formation and dissipation of low level clouds as well as the accumulation/evaporation of snow. The location appears to be important because the

thermodynamic perturbation of BC is much more effective in low temperature. As shown in Figures 6 and 7, because the model predicted snow depth in most areas in the high latitudes of the Northern Hemisphere is equal or lower than 100 mm except for Greenland, the change in snow depth caused by BC (quite often ±20%) produces a considerable change in surface albedo in these areas where the snow cover really matters. Note that there are some BC-caused changes in snow depth in the Antarctic, however, these changes are relatively small due to the large background value of snow depth (> 300 mm) over Antarctica, the resultant effect on albedo hence is almost invisible.

Table 2. Annual and global means of various radiative forcings of BC averaged over the 20-year integration periods, all in W/m^2.

	TOA Clear Sky	TOA All Sky	Surface Clear Sky	Surface All Sky
Model 1	0.27	0.22	–0.57	–0.64
Model 2	0.32	0.18	–0.76	–0.57
Model 3	0.28	0.20	–0.80	–0.92

Table 3. Minima/maxima of annual and global means of various radiative forcings of BC based on the 20-year results, all in W/m^2.

	TOA Clear Sky	TOA All Sky	Surface Clear Sky	Surface All Sky
Model 1	0.12/0.40	–0.27/0.58	–0.73/–0.48	–1.21/–0.23
Model 2	0.19/0.46	–0.24/0.44	–0.91/–0.62	–0.84/–0.32
Model 3	0.14/0.40	–0.11/0.65	–0.99/–0.42	–1.33/–0.42

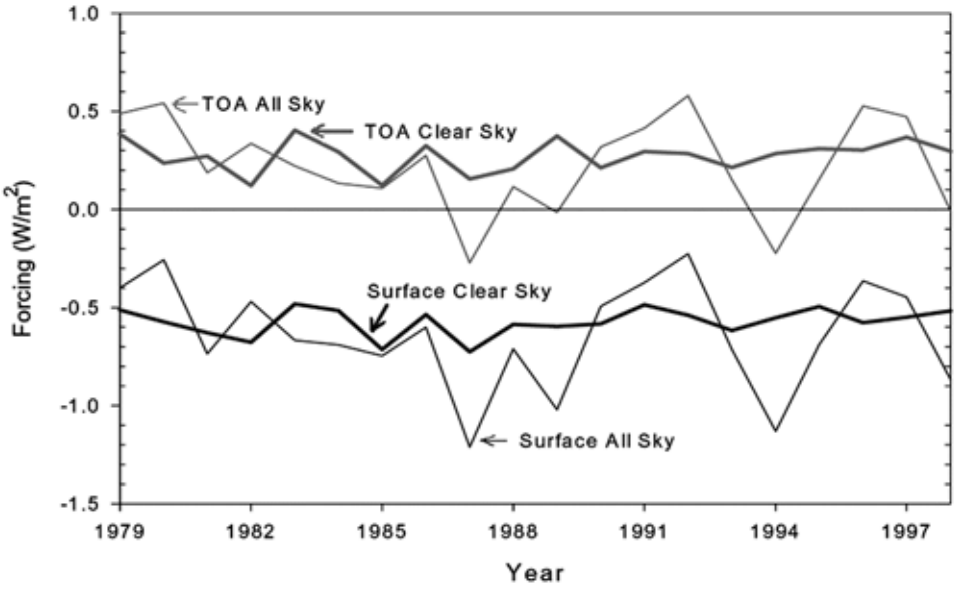

Figure 5. Annual-mean clear sky and all sky radiative forcings induced by black carbon aerosols, presented as a function of time. The global mean values are derived based on area-weighted averages of the forcings as functions of longitude and latitude for a given vertical layer derived using Equation 2.

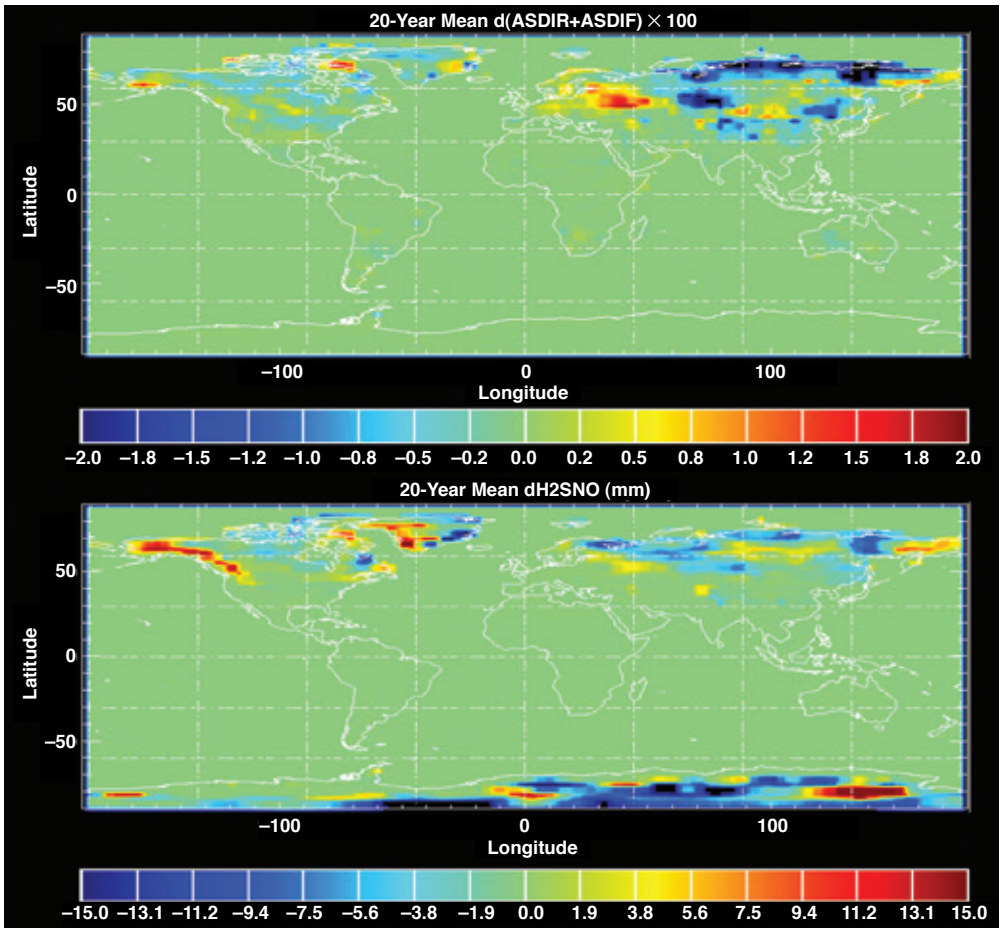

Figure 6. Changes in surface solar albedo (sum of direct and diffusive, or ASDIR + ASDIF; *upper panel*) and surface snow depth, or H2OSNO (in mm; *lower panel*) caused by BC forcing. These are derived by deducting REF run results from the Model 1 BCRAD run results (similar to Equation 2), and are presented as average values of monthly mean data over the 20-year integration period.

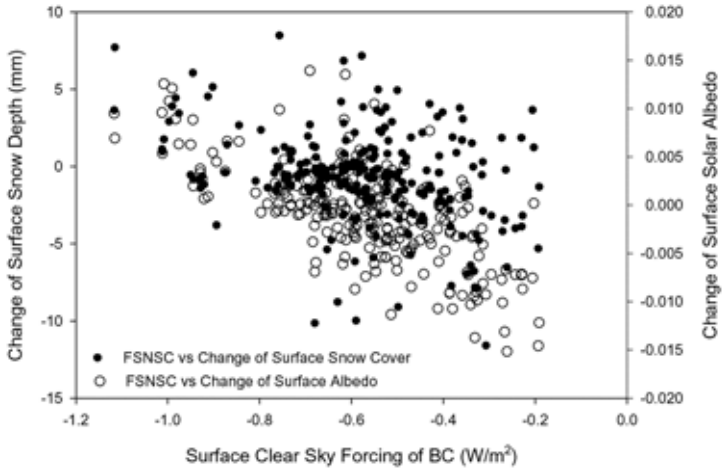

Figure 7. Global-mean of surface clear sky forcing of BC and BC-caused changes in the Northern Hemispheric mean of snow depth (in mm) and surface solar albedo. All data are derived based on monthly means of REF and Model 1 BCRAD runs.

The modeled results indicate that the direct forcing of black carbon aerosols is primarily concentrated in the Northern Hemisphere with some clear extensions over the Southern Hemisphere (**Figure 8**). Generally speaking, contrasting with the positive atmospheric forcings (forcings at the TOA) are significant negative surface forcings located mainly in East Asia, South Asia, and Central Africa. However, it has been shown that the decrease of surface albedo caused by BC radiative forcings in several high-latitudinal areas lead to positive clear sky forcings at both the TOA and the surface over these regions, which would not otherwise happen. These results indicate the close feedbacks between BC radiative forcings and resultant processes and thus strongly suggest the use of interactive aerosol-climate models to address the issue of climatic impact of absorbing aerosols.

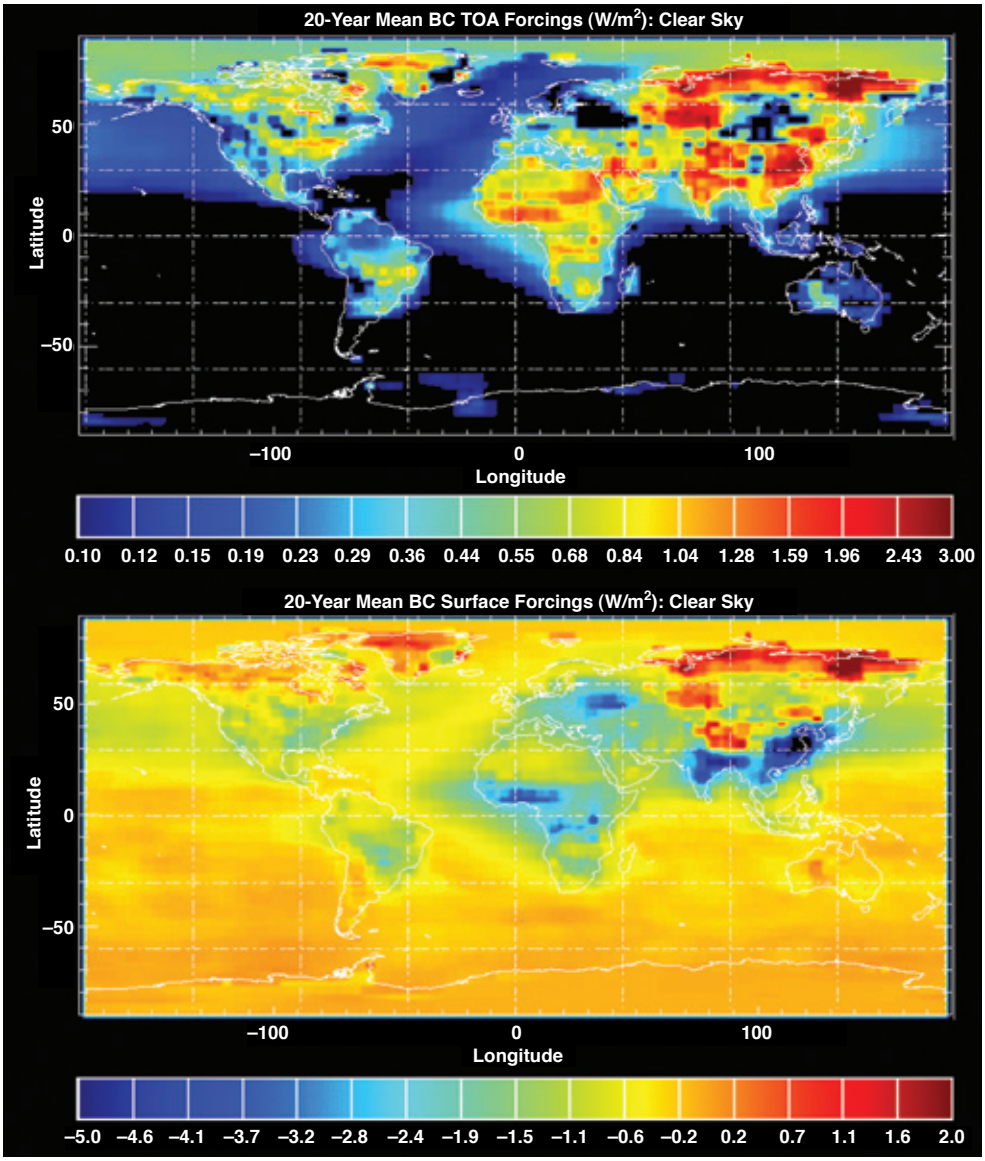

Figure 8. Modeled annual-mean clear sky radiative forcings of BC at the top of atmosphere (*upper panel*) and the Earth's surface (*lower panel*). Results are derived from the 20-year means.

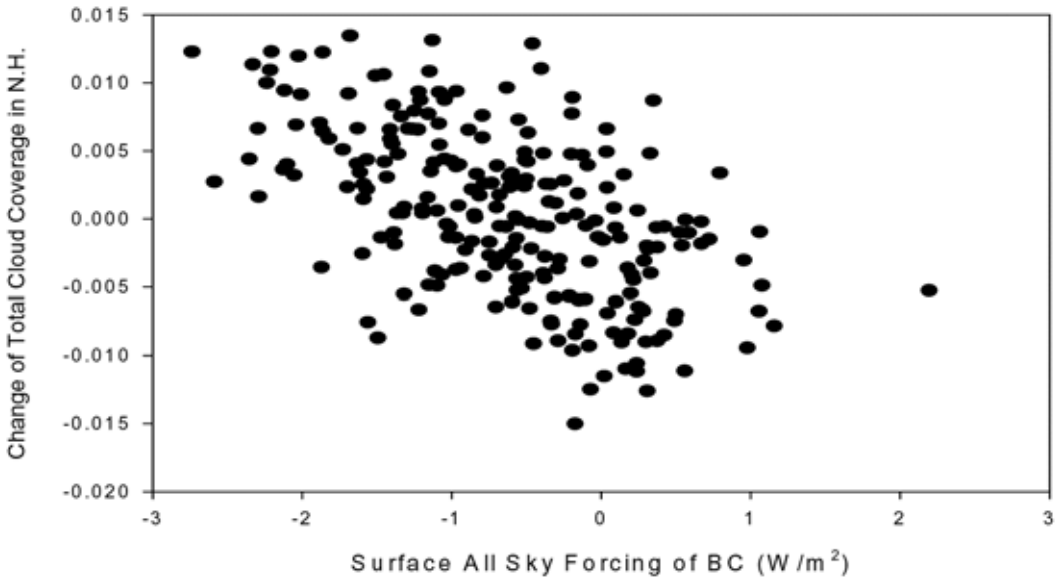

Figure 9. Modeled global-mean all sky forcings of BC at Earth's surface (W/m^2) and BC-caused changes in the Northern Hemispheric mean total cloud cover. Both are derived from monthly means of REF and Model 1 BCRAD runs.

Although the 20-year means of all sky radiative forcings caused by BC are close to those of clear sky diagnostics as shown in Figure 5 and Table 2, the ranges of variation of all sky forcings are more than 3 times larger than those of clear sky forcings as indicated earlier. It has been found that the variations in all sky forcings are closely related to the changes of cloud coverage induced by BC radiative forcing (**Figure 9**). In addition, the spatial distributions of all sky forcings caused by BC (not shown) are more scattered compared with the clear sky ones and spread out over both land and ocean. All these results indicate an enhancement of BC radiative forcing by clouds and also suggest the possible existence of "indirect forcing" by BC.

The indirect forcing of aerosols traditionally refers to a process in which aerosols impact climate not by directly altering atmospheric radiation, but by changing cloud properties and thus the radiative budget, through an increase of the atmospheric concentration of cloud condensation nuclei (CCN) that produce optically thicker and more reflecting clouds (Twomey, 1977b), or lead to a decrease of precipitation and hence an increase of abundance of certain clouds (Albrecht, 1989). Besides these two mechanisms, BC aerosols may alter cloud properties through thermodynamic processes such as changing the vertical temperature or radiation profile and thus the atmospheric stability, or enhancing evaporation by warming the cloudy air to cause dissipation or partial dissipation of clouds. The latter process has been discussed in Ackerman *et al*. (2000) as well as in Lohmann and Feichter (2001) (more generic discussions can be also found in Haywood and Boucher, 2001; Ramanathan *et al*., 2001). The impact of BC on the surface properties such as evaporation and cooling rate, heat fluxes, and surface albedo as shown in previous discussion, can also change the water and energy budget particularly in the planetary boundary layer and hence consequently influence cloud formation or dissipation. The radiative forcing formed through these above-discussed (mainly thermodynamic) processes should be referred to as a "non-Twomey-Albrecht" indirect forcing (NTAIF), or as suggested in Hansen *et*

al. (1998), the "semi-direct" forcing. Obviously, the indirect forcing caused by BC revealed in this study must be a NTAIF since BC aerosols are not allowed to serve as CCNs in this model. This NTAIF can be defined as the difference between BC's all sky forcing and its clear sky forcing, or in other words, the cloud forcing purely caused by BC aerosols:

$$NTAIF(x,y,h,t) = \Delta F_{BC}^{allsky}(x,y,h,t) - \Delta F_{BC}^{clearsky}(x,y,h,t) \tag{3}$$

where the two forcings on the right hand side of the equation are defined in Equation 2 for all sky and clear sky, respectively. Note this formula has actually deducted the cloud forcing caused by other factors (*e.g.*, sulfate aerosols) under the external mixing assumption. The impact on surface property by BC aerosols, however, is preserved. The 20-year means of NTAIF derived from the model runs are –0.07 and –0.05 W/m^2 at the surface and TOA, respectively, for the Model 1 run, or 22% and 11% of the absolute values of the corresponding all sky forcings (**Table 4**). Interestingly, the ranges of variation of annual mean NTAIF during the 20-year integration are very large, namely 0.89 and 0.81 W/m^2 at the TOA and surface, respectively, and this implies a large variability in the thermodynamic kind of BC aerosol-cloud interaction (note the variation range of all sky forcing is much larger than that of clear sky forcing) (**Table 5**). The pattern of longitude-latitude distribution of NTAIF is quite complicated (**Figure 10**); significant NTAIF exists not only over the Northern Hemisphere as expected but also over the Southern Oceans. The results suggest that changes in atmospheric circulation, especially the meridional transport of heat and water vapor, may be caused by BC forcing. High positive NTAIF can be seen over Eastern China, South Asia, and Northern America, representing a decrease of cloud cover, or "cloud burning" effect, in these polluted areas apparently due to enhanced evaporation of clouds caused by the warming effect of BC, which is supported by a case study of a marine stratocumulus cloud system outside the Indian coast (Ackerman *et al*., 2000). This "cloud burning" pattern seems to be quite persistent through seasons, particularly over heavily polluted areas such as Eastern China. In contrast, at the northern edge of the Russian Far East, a negative forcing induced by a local increase in clouds (mainly low and middle clouds) can be seen. This enhancement of cloud coverage is due to the adiabatic cooling at the surface caused by a unique distribution of BC aerosols in the middle and upper troposphere over this region as shown in next section. The signature of surface albedo changes is hardly noticeable in the NTAIF distribution.

Table 4. 20-year averages of the annual and global means of NTAIF of BC (in W/m^2).

	TOA	Surface
Model 1	–0.05	–0.07
Model 2	–0.14	0.18
Model 3	–0.08	–0.12

Table 5. Minima/maxima of annual and global means of NTAIF of BC based on the 20-year results, all in W/m^2.

	TOA	Surface
Model 1	–0.51/0.31	–0.52/0.31
Model 2	–0.65/0.18	–0.19/0.48
Model 3	–0.43/0.40	–0.49/0.40

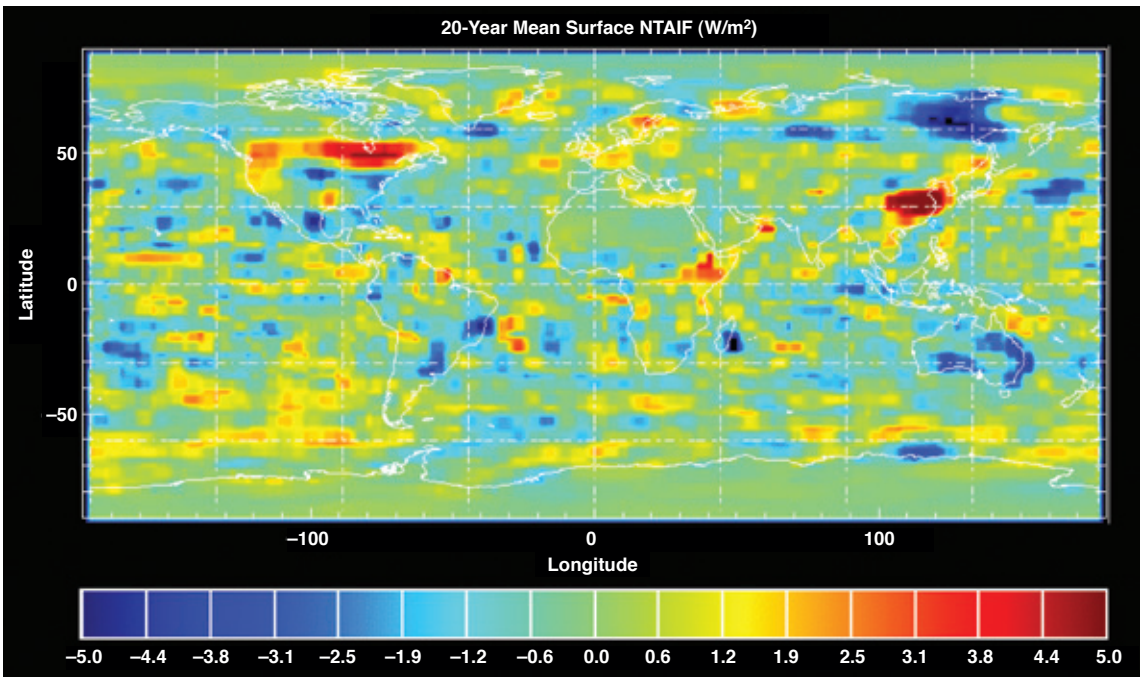

Figure 10. Twenty-year-mean spatial distribution of the non-Twomey-Albrecht indirect forcing at the surface in W/m^2, derived from annual mean data of REF and Model 1 BCRAD runs based on equation 3.

4. EFFECT OF BLACK CARBON ON SURFACE TEMPERATURE

The modeled results suggest that BC heats the atmosphere mainly between 5^o and 50^o latitude in the Northern Hemisphere, corresponding to the most polluted areas (**Figure 11a**). The heating applies through most of the troposphere. At high latitudes in the Northern Hemisphere, BC-induced atmospheric solar heating primarily occurs in the middle and upper troposphere due to horizontal transport of aerosols which most likely were first transported vertically by convection in mid-latitudes (interestingly, there is a small cooling zone close to the ground in the high latitude Northern Hemisphere). Corresponding to this atmospheric warming, there is a significant reduction in absorbed solar radiation by the Earth's surface (**Figure 11b**) that obviously can lead to a cooling of the ground. This sharp change of the thermodynamic effect caused by BC from warming to cooling across the atmosphere-land interface generates negative fluxes of both sensible and latent heat in the low- and middle-latitudes in the Northern Hemisphere to compensate for the radiative cooling at the surface caused by black carbon aerosols. Resultant changes in surface temperature are thus determined by the difference between compensating heat fluxes and the reduction in absorbed solar radiation at the surface. Note a previous study using an interactive sulfate aerosol-climate model (Roeckner *et al*., 1999) has also indicated the similar response of surface heat fluxes to the reduction of incoming solar radiation caused by sulfate aerosols. However, the situation related to BC aerosol becomes more complicated due to BC's atmospheric heating, which is opposite to the sulfate aerosols' cooling effect on atmosphere. In the low- and middle-latitudes of the Northern Hemisphere, large negative heat fluxes somewhat offset the cooling effect caused by the decrease of absorbed solar radiation at the surface.

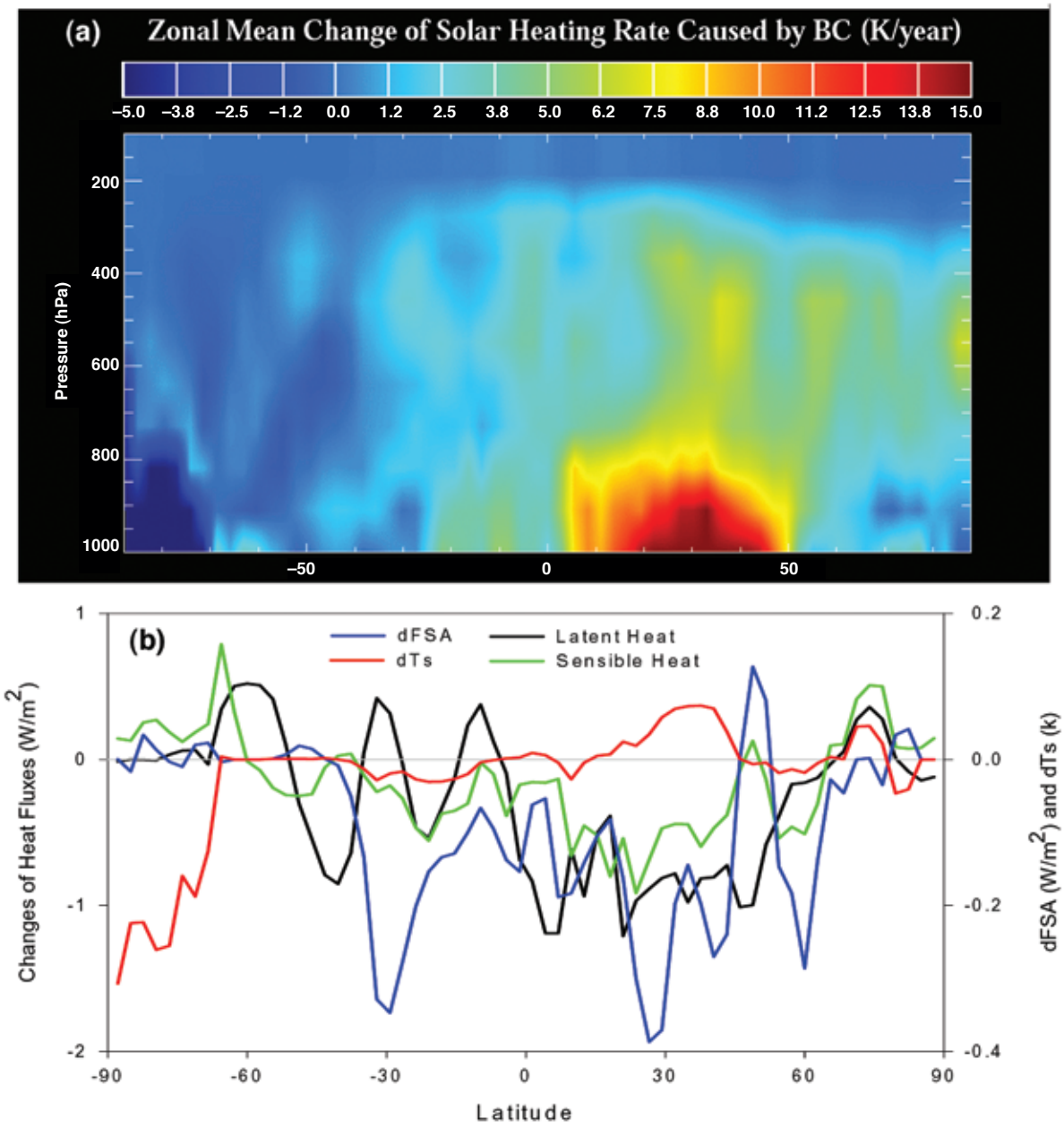

Figure 11. Zonal mean changes of: **(a)** atmospheric solar heating rate in K/year; and **(b)** solar radiation absorbed by the Earth's land surface (dFSA in W/m^2), latent and sensible heat fluxes (in W/m^2), and land surface temperature (dTs in K) caused by including the black carbon aerosol radiative effect in the model. All data are 20-year averages.

Overall, the surface temperature has slightly increased in these areas. However, in the Southern Hemisphere mid-latitudes (note that only relatively small land areas are involved), the cooling effect caused by reduced surface absorption has not been compensated by the sensible and latent heat fluxes and this leads to a decrease of surface temperature. A cooling can be seen in the South Polar Region primarily due to changes in meridional sensible heat transport apparently caused by the BC forcing (not shown).

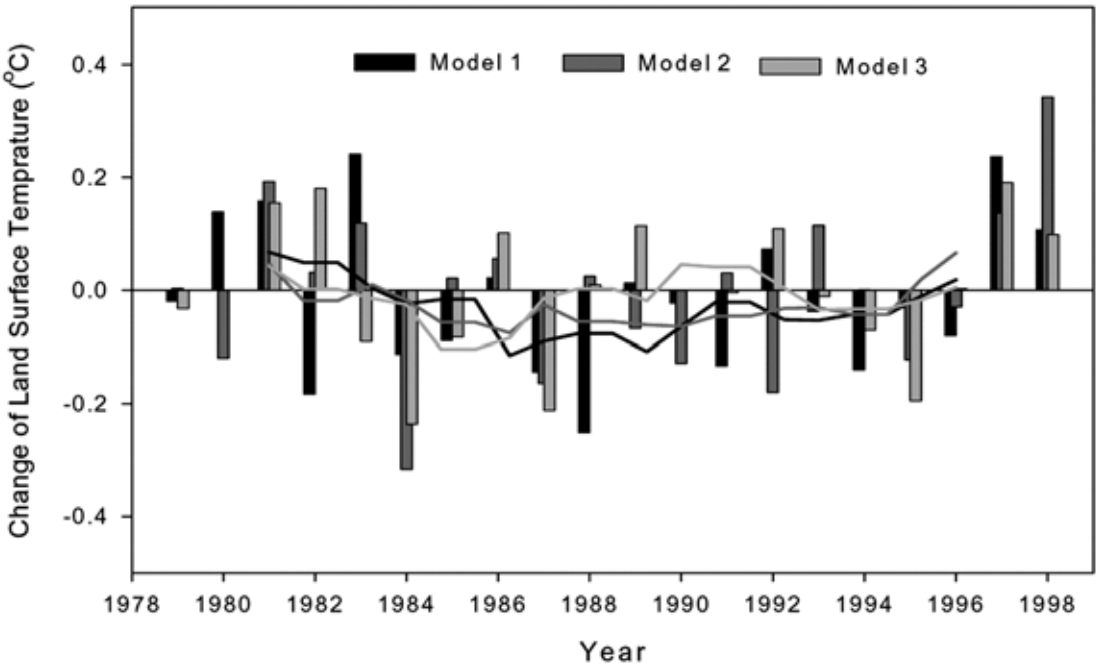

Figure 12. Forced changes of annual and global mean land-surface temperature by black carbon aerosols, derived by subtracting the temperatures of the REF run from those of the BCRAD runs. The lines show the five-year running averages. Global mean values are area-weighted averages.

There is no monotonic warming or cooling "trend" that can be derived from the modeled annual mean land surface temperature using all three different sets of optical parameters during the 20-year model integration period (**Figure 12**). Under the radiative forcing of BC aerosols, the forced change of land-surface temperature experiences a short warming period (1979 to 1982) then a long but slight cooling period (1984 to 1996; note Model 3 showed a weak warming trend between 1987 and 1992) and finally swings upward again in the last two years. Therefore, it cannot be concluded that black carbon at its emissions levels used in this study will cause a significant global-scale increase of Earth's land-surface temperature.

5. CONCLUSION AND DISCUSSION

As the results have demonstrated, responses of the modeled climate system to the BC aerosol radiative forcings are complicated and in many occasions and locations the resultant changes in the atmosphere or at the Earth's surface compensate each other. With an absolute amount 3 times higher than that of the TOA forcing, the surface forcing of BC is an extremely important factor in analyzing the climate impact of BC. Results from this study and many previous ones (see Haywood and Boucher 2000; Penner *et al*., 2001; Ramaswamy *et al*., 2001; Ramanathan *et al*., 2001) challenge the practice of using only the TOA radiative forcing by BC to project the possible climate impact of BC. In addition, the result of this study does not suggest that BC aerosols contribute to a significant increase in land-surface temperature with annual emissions of ~14 TgC (note that similar numbers have been used in several related studies). The calculated surface temperature change is determined by a subtle balance among changes in surface energy sources and sinks as well as changes in the hydrological cycle, all caused by BC radiative forcing.

This study reveals that BC aerosols can cause changes in regional climate by producing "cloud burning" in polluted regions while "cloud enhancing" in some high-latitude sites, as well as by altering precipitation and snow cover (~ ±20% over many high- or even mid-latitude regions in the Northern Hemisphere) then surface albedo, and therefore to form a "non-Twomey-Albrecht" indirect forcing. This change in forcing has yet to be directly related to the global scale climatic changes in this study. However, the potential importance of this type of indirect climate forcing to climatic and environmental changes is clear from this study. Further evaluations of the relationship between NTAIF and climatic and environmental changes are needed. This study indicates several important feedbacks between BC radiative effect and climate dynamics. In particular, it has been found that BC-caused change in surface albedo and in cloud cover is responsible for the significant interannual variations of BC's clear sky and all sky radiative forcings, respectively. It suggests the importance of using interactive aerosol-climate models to address the issues related to the climate impacts of aerosols.

It is of course true that current descriptions of many above-mentioned processes even in the most sophisticated climate models still have room for improvement. In particular, simulations of aerosol behaviors in the GCMs like the ones used in this study are quite crude in some aspects, especially those related to size-dependent processes or that deal with microphysical conversions. This suggests great uncertainty in estimating possible changes in climate, especially in surface temperature, caused by black carbon aerosols. In addition, high-resolution mesoscale models with more detailed physics and chemistry are needed to enhance our knowledge of the climate effects of absorbing aerosols and to provide possible improvements of climate models.

Two major simplifications made in this study, *i.e.*, using prescribed SST and assuming external mixing status of black carbon aerosols, could quantitatively influence the final results. The former assumption might lead to errors in analyzing climate response to aerosol forcings (*e.g.*, Kiehl, 1992). The latter assumption in particular could potentially lead to an underestimate of the forcing strength of BC (Chýlek *et al.*, 1995; Fuller *et al.*, 1995). However, neither of these simplifications should qualitatively change the major conclusions other than enhancing them. They will be improved in future study using coupled atmosphere-ocean model and a more comprehensive aerosol module.

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