Soot climate forcing via snow and ice albedos

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Plausible estimates for the effect of soot on snow and ice albedos (1.5% in the Arctic and 3% in Northern Hemisphere land areas) yield a climate forcing of +0.3 W/m² in the Northern Hemisphere. The "efficacy" of this forcing is ≈2, i.e., for a given forcing it is twice as effective as CO_2 in altering global surface air temperature. This indirect soot forcing may have contributed to global warming of the past century, including the trend toward early springs in the Northern Hemisphere, thinning Arctic sea ice, and melting land ice and permafrost. If, as we suggest, melting ice and sea level rise define the level of dangerous anthropogenic interference with the climate system, then reducing soot emissions, thus restoring snow albedos to pristine high values, would have the double benefit of reducing global warming and raising the global temperature level at which dangerous anthropogenic interference occurs. However, soot contributions to climate change do not alter the conclusion that anthropogenic greenhouse gases have been the main cause of recent global warming and will be the predominant climate forcing in the future.

aerosols | air pollution | climate change | sea level

The Intergovernmental Panel on Climate Change (1) estimates the global climate forcing by fossil fuel black carbon (BC) aerosols as 0.2 W/m². Jacobson (2) suggests that the fossil fuel BC forcing is larger, ≈0.5 W/m². J.H. and colleagues (3–5) have argued that the total anthropogenic BC forcing, including BC from fossil fuels, biofuels, and outdoor biomass burning, and also including the indirect effects of BC on snow/ice albedo, is still larger, 0.8 ± 0.4 W/m². Here we estimate the magnitude of one component of the BC climate forcing: its effect on snow/ice albedo.

Several factors complicate evaluation of the BC snow/albedo climate forcing and dictate the approach we use to estimate the forcing.

BC Amount in Snow. BC is highly variable in space and time. A classic study of Arctic sites (6) found average BC (excluding Greenland) of 30 ppbw (parts per billion by weight; equivalent to ng/g or μg/liter meltwater) in 1983–1984. In contrast, only 4 ppbw was found in 1998 (7) upwind of the drifting SHEBA (Surface Heat Budget of the Arctic Ocean) site on Arctic sea ice (≈76°N, 165°E), although 35 ppbw and more was found in a limited region downwind of the SHEBA site. Differences among measured amounts exceed estimated errors and imply real variability.

BC Optical Effects. Classic theory calculates snow/ice albedo as multiple scattering by BC and ice spheres (8, 9). Empirical data (8) revealed that BC was two to five times more effective in reducing snow albedo than the model indicated. A factor of two was accounted for by use of a more realistic density and absorption cross-section for BC. Still greater absorption per unit BC mass, by perhaps another factor of two, can be obtained with realistic shapes, voids, and degrees of internal mixing of the BC within ice particles (10, 11), but there remains uncertainty and variability of the appropriate absorption cross-section for BC dispersed in snow.

BC vs. Soot. Soot is produced by incomplete combustion of carbonaceous material, mainly fossil fuels and biomass. The carbon combustion products are usually classified as BC and organic carbon (OC), but distinctions within the complex carbonaceous mixtures can be ambiguous (12, 13). BC is commonly defined in an operational sense as the absorbing component of carbonaceous aerosols, which may result in some humic-like or other organic material contributing to estimated BC absorption. We employ this operational definition, because the practical question concerns the impact of soot on snow/ice albedo, and for this it matters little whether the carbon is elemental or in other carbonaceous aerosols.

We compile empirical data on BC amount in snow and compare calculated effects on snow albedo with field data. We then use these data to specify plausible albedo changes for climate simulations. We calculate the forcings due to specified albedo changes, carry out equilibrium climate simulations for these forcings, calculate the efficacy of snow/ice albedo forcing relative to CO_2, and carry out transient climate simulations for 1880–2000. Finally, we discuss potential implications.

Soot in Snow. Be thou as chaste as ice, as pure as snow, thou shalt not escape calumny (Shakespeare, *Hamlet*). Perceptions persist about the purity of fresh snow, but measurements tell another story. Optical and electron microscopes showed (14) the typical snow crystal in Sapporo, Japan in the 1970s to contain thousands of aerosols, including soot. It was inferred (14) that the snowflakes, falling from ≈500 m, had a collection efficiency of order unity (>0.2) in sweeping up aerosols in their path. In central Antarctica, snow crystals in 1969 were found to commonly contain 25–50 aerosols (15), two orders of magnitude less than in Japan, but again implying a collection efficiency of order unity. Mechanisms for high collection efficiency may include (15) electrostatic attraction, thermophoresis (temperature gradient between snowflake and environment), and diffusiophoresis (vapor pressure gradient between snowflake and environment).

High collection efficiency is not general, but it is believed that wet deposition (via snow and rain) is the primary removal mechanism for aerosols as small as BC (16, 17), which has a typical dimension 0.1 μm (12, 13). Dry deposition also can be significant, accounting for several tens of percent of deposition in some situations (18).

We use observed BC amounts in surface snow and precipitation, together with measured snow albedos, to estimate BC albedo effects. The distribution of BC in global aerosol transport models helps extrapolate site measurements to the global scale.

BC Concentration in Snow. Table 1 summarizes BC amounts in surface snow and precipitation based on data in Table 4, which is published as supporting information on the PNAS web site. Snow samples in the 1980s (6), including sites in Alaska, Canada, Greenland, Sweden, and Spitzbergen, and on sea ice in the central Arctic, yielded typical BC amounts of 10–50 ppbw (excluding Greenland). "Arctic haze" studies (19) showed that most of the aerosols originated in Europe and the Former Soviet Union (FSU) in winter/spring driven by circulation around the Icelandic Low and Siberian High. BC emissions from Eurasia probably declined sharply in the 1990s as, e.g., FSU BC emissions fell by a factor of

Abbreviations: BC, black carbon; DAI, dangerous anthropogenic interference; ppbw, parts per billion by weight.

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Table 1. Measured BC amount and calculated visible snow albedo change

Location	Observed A_v, %	BC amount, ppbw	Calculated ΔA_v, %			
			New snow		Old snow	
			Ext	Int	Ext	Int
Arctic, 1980s	≈90–97	10 (low)	0.8	1.5	2.5	4.5
		30 (mean)	1.9	3.2	6.0	9.5
NH land	≈88–95	20 (low)	1.5	2.5	4.5	7.7
		60 (high)	3	5	9	14
Greenland		2 (low)	0.3	0.5	0.7	1.2
		6 (high)	0.5	0.9	1.7	3.0
Antarctica		0.2 (South Pole)	.05	0.1	0.1	0.2
		2.5 (Ross Shelf)	0.3	0.5	0.8	1.5

NH, Northern Hemisphere; Ext, external mixing; Int, internal mixing.

four with the collapse of the FSU economy (20). Reduced BC emissions are not necessarily permanent in the face of possible economic recovery, increased shipping in the opening Northwest and Northeast Passages, regional hydrocarbon resource development, and increased use of diesel-powered vehicles.

BC amounts found in snow at lower latitudes in the Northern Hemisphere are highly variable, usually in the range 5–100 ppbw. Larger amounts found in the French Alps, ≈100–300 ppbw (21), may be related to the high proportion of diesel engines in European surface transportation. It seems likely that East Asia snow has large BC amounts, because China and India are now the largest sources of BC emissions, and photographs reveal a thick brown haze filled with BC that butts against the Himalayas (22), but measurements are lacking. Greenland BC measurements yield mainly 2–6 ppbw.

Pristine Antarctic regions have been found to contain 0.1–0.3 ppbw (23, 24), two orders of magnitude less than in the Arctic. BC amounts of 3 ppbw were found 1 km downwind of the South Pole station (23), where the station's power plant and aircraft operations were a suspected source. BC of 3 ppbw was reported at Siple Dome above the Ross Ice Shelf (25), an amount that might be related to the location being closer to the coast, but the amount seems large for Antarctica unless it was influenced by local pollution.

Clarke and Noone (6) found a consistent "scavenging ratio," relating BC amount in snow to BC concentration in air, in polar regions of both hemispheres. This relation allows estimates of BC in snow for other locations (see *Supporting Text*, which is published as supporting information on the PNAS web site).

BC Effect on Snow Albedo. Theory relating snow albedo to BC amount (8, 9), which combines Mie scattering and a multiple scattering approximation, accounts well for measured albedo as a function of ice crystal size at wavelengths where absorption by ice is dominant ($\lambda > 0.7$ μm). The presentation of this theory in figure 2 of ref. 9 accounts for externally mixed soot and ice particles but leaves open the possible need to increase the BC effective absorption by perhaps a factor of two for best agreement with empirical data. A critique (11) concludes that there are inherent uncertainties in the effect of BC on snow albedo; specifically, (*i*) if BC particles are within ice crystals, rather than externally mixed, the absorption power of BC particles increases by a factor 1.4 or more; (*ii*) the soot particle shape affects its absorption power, with randomly oriented needles or disks being more absorbing then spheres of the same mass, by factors as much as two or more; (*iii*) voids in the soot particles increase the absorption power per unit mass nearly in proportion to the void fraction; and (*iv*) there is large uncertainty in the optical constants of soot, as much as a factor of two or more in the imaginary part of the refractive index.

Given the uncertainty in the effect of a given BC mass on snow albedo, we consider two cases in Table 1. External mixing is the standard relation (figure 2 of ref. 9), whereas internal mixing increases the BC absorption coefficient by a factor of two, for better agreement with empirical data. Table 1 gives results for both fine-grained fresh snow and snow in which the ice particles are larger because of aging or melting effects. Snow grain size tends to be larger in late winter and spring, when there is enough sunlight that BC absorption is most important. Average conditions in nature should fall between fresh and old snow, and empirical data suggests that internal mixing is a better approximation than external mixing, but substantial uncertainty must be admitted.

Specified Albedo Changes. We carry out climate simulations for specified snow/ice albedo changes of Table 2. The multiple cases help determine the contributions of different geographical regions. The response is linear for these small forcings, so results for alternative forcings with the same geographical distribution are implicit. In cases 1 and 2, the albedo is decreased by the specified amount only at $\lambda < 0.77$ μm and half that amount at 0.77 μm $< \lambda <$ 0.86 μm. The spectrally integrated albedo change is ≈60% of that at $\lambda < 0.77$ μm. Cases 3 and 4 change the albedo individually for Arctic sea ice and Northern Hemisphere land, with the change made at all wavelengths solely to assure good signal/noise response.

The snow albedo at $\lambda < 0.77$ μm is decreased 2.5% (1.5% spectrally averaged) in the Arctic, a realistic and perhaps conservative estimate of BC effects in the 1980s, by 1% at $\lambda < 0.77$ μm in Greenland, which contributes negligibly to the hemispheric mean response, and by 5% at $\lambda < 0.77$ μm (3% spectrally averaged) in other snow-covered land areas in the Northern Hemisphere, because BC in the air and in precipitation is about twice as large in these regions compared to mean Arctic conditions in the limited measurements available and in tracer transport models. All cases have no albedo change in Antarctica. Case 1 has 1% albedo change in snow regions of the Southern Hemisphere other than Antarctica. We consider case 1 to be the most realistic.

Climate Model. Our simulations use the current Goddard Institute for Space Studies climate model. Model physics is similar to that in the SI2000 model (5), but with the current "model E" version of the code as modularized, modified for parallel computations, documented, and otherwise improved by G. Schmidt and R. Ruedy.

Table 2. Specified snow and ice albedo changes

Experiment	Arctic, %	NH land, %	Antarctica, %	Rest of SH, %
Case 1	2.5 (vis λ)	5 (vis λ)	0	1 (vis λ)
Case 2	2.5 (vis λ)	5 (vis λ)	0	0
Case 3	0	5 (all λ)	0	0
Case 4	2.5 (all λ)	0	0	0

NH, northern hemisphere; SH, southern hemisphere.

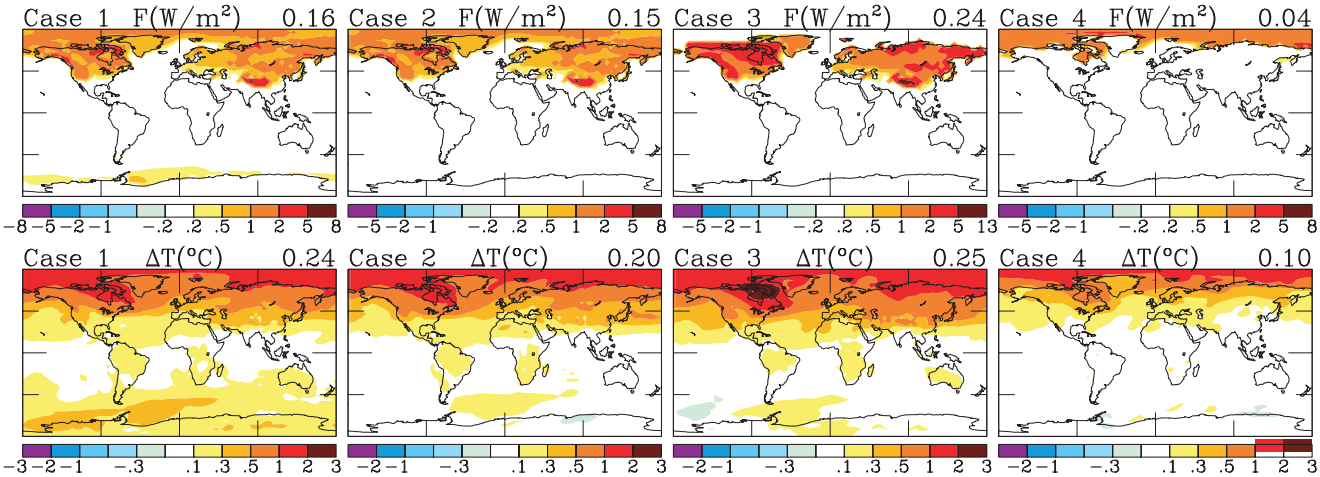

Fig. 1. Climate forcing in W/m² (*Upper*) and equilibrium annual-mean T_s response in °C (*Lower*) for changes of snow and ice albedos specified in Table 2. Numbers on the upper right are global means.

Specifically, we use model version E037 that existed in the summer of 2003, which we dub SI2003. Resolution is 4° × 5° with 18 layers, the added 6 layers, compared with the 12-layer SI2000 model, being in the stratosphere. The model top is raised to 0.1 hPa and stratospheric drag is reduced, compared with SI2000, to allow more realistic stratospheric climatology (J. Perlwitz and J.H., unpublished work). The climate sensitivity of the SI2003 model is ≈2.6°C for doubled CO_2, somewhat less than the ≈3°C sensitivity of the SI2000 model.

Climate Forcing. Climate forcing caused by the specified snow/ice albedo changes (Table 2) is shown in Fig. 1. This adjusted forcing, F_a, is obtained as the flux change at the top of the atmosphere after the stratospheric temperature has adjusted to the presence of the perturbation with the troposphere and surface temperatures fixed. F_a is a commonly used forcing (1, 26). An alternative forcing definition is compared below.

The climate forcing due to snow/ice albedo change is of the order of 1 W/m² at middle- and high-latitude land areas in the Northern Hemisphere and over the Arctic Ocean for cases 1 and 2, which we suggest have a realistic magnitude for the soot effect (3% spectral-mean snow albedo change over land, 1.5% over the Arctic, and 0.6% over Greenland). This compares with a global mean forcing by present anthropogenic CO_2 (compared to preindustrial times) of ≈1.5 W/m², which is relatively uniform over the globe (1, 26). The mean Northern Hemisphere forcing is ≈0.3 W/m² in the more realistic cases 1 and 2.

Equilibrium Response. We calculate the equilibrium response to these forcings in the Goddard Institute for Space Studies climate model with a Q-flux ocean in which horizontal heat transports are specified (27). The annual mean response of surface air temperature (T_s) is shown in Fig. 1. The calculated warming is unexpectedly large, relative to the forcings, if one uses the sensitivity to standard forcings such as CO_2 to judge the expected response. The reason for the greater "efficacy" of the surface albedo forcing is discussed below.

Fig. 2 shows the calculated equilibrium T_s response as a function of month and latitude. This, and the map of temperature change (Fig. 1), can be compared with observed temperature change since 1880 (Fig. 3). The model results are an equilibrium response, but transient results are similar, as shown below. The observed change is based on an analysis (28) that screens and adjusts meteorological station data from the Global Historical Climatology Network for urban and other inhomogeneities and uses sea surface temperature data (29, 30) for ocean regions. It is apparent that warming due to snow/ice albedo change potentially could account for a significant fraction of observed warming. The location and season of observed warming in the Northern Hemisphere, with warming at middle and high latitudes that is large in the winter, extends well into the spring, and is minimal in summer, is consistent with a significant fraction of the warming being due to a decreased snow/ice albedo.

These geographical and seasonal features do not confirm the existence of a snow/ice albedo forcing. High-latitude and winter amplifications of the climate response are a natural response of the climate system and climate models to most forcings. Careful study of the transient climate response to all forcings is needed, and this requires information on the time-dependence of forcings, some of which may be difficult to establish.

Nevertheless, it is worth noting that the anthropogenic soot snow/ice albedo forcing and response must be mainly in the Northern Hemisphere. This is the same region in which the negative direct and indirect forcings by anthropogenic sulfate and nitrate aerosols are concentrated. If these negative forcings are as large as has been estimated (1, 3), it raises the question of why observed warming is as large in the Northern Hemisphere as in the Southern

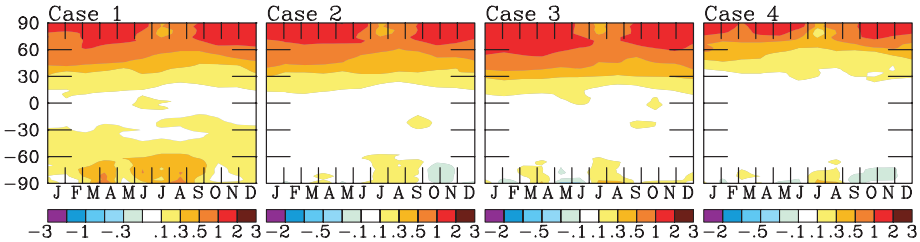

Fig. 2. Equilibrium T_s (°C) response to the snow/ice albedo forcings of Fig. 1 as a function of month and latitude.

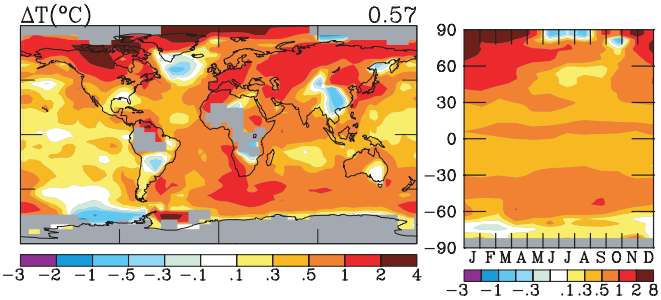

Fig. 3. Observed 1880–2002 T_s change based on adjusted meteorological station data over land (28) and sea T_s data for the ocean (29, 30).

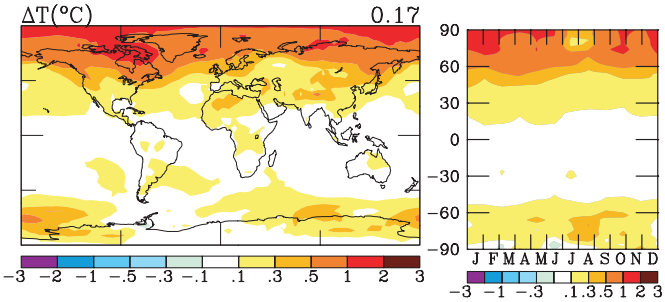

Fig. 4. Simulated 1880–2002 T_s change for the transient BC snow/ice albedo forcing that peaks in the 1990s with 83% of the case 1 forcing of Fig. 1.

Hemisphere. A suggested explanation (5) is that there are large positive air pollution forcings in the same region, primarily BC and tropospheric O_3. Figs. 2 and 3 suggest that the BC indirect snow/ice albedo forcing could provide a substantial part of the reason for large Northern Hemisphere warming.

Efficacy. When climate forcings are used to compare different climate change mechanisms, there is an implicit assumption that the climate response is expected to be proportional to the forcing. We (26) tested this assumption for a large number of forcings and found that it was good for many forcings, in the sense that the global mean T_s response was consistent among these forcings within ≈20%. Absorbing aerosols and O_3 were identified as exceptions, for which the climate response depended sensitively on the altitude at which the forcing was introduced. We did not explicitly test the effectiveness of changes in snow albedo, but we showed that an arbitrary ("ghost") forcing at the surface was twice as effective at high latitudes as at low latitudes. Reasons for the large response to high-latitude surface forcings were identified as (*i*) the ability of a high-latitude forcing to stimulate snow/ice albedo feedbacks and (*ii*) the relative stability of the atmospheric temperature profile at high latitudes, which tends to confine the thermal response to the surface.

We define the "efficacy" of a climate forcing, E, as the ratio of the equilibrium T_s change for that forcing to the temperature change for the same magnitude of forcing by CO_2. Comparison with CO_2 has practical merit, because CO_2 is expected to be the largest anthropogenic climate forcing in the foreseeable future.

Table 3 gives the efficacies for the snow/ice albedo forcings of Fig. 1 for two definitions of climate forcing. F_i and F_a are the standard "instantaneous" and "adjusted" climate forcings (1, 26). For these forcings, the efficacy of the snow/ice albedo forcing is ≈2, i.e., the equilibrium T_s response to the forcing is about twice as large as for a CO_2 forcing of the same magnitude. (We did not run the climate model for small CO_2 forcings equivalent to the ice/snow albedo forcings but rather scaled the doubled CO_2 response in proportion to the forcing. For positive forcings within the present range that is a good approximation. For larger forcings, and especially for negative forcings, high accuracy requires that the climate model be run for several CO_2 changes to define the nonlinear relationship.)

The variation of efficacy from one case to another is due mainly to the latitude of the forcings. For the reasons mentioned above, the response to polar forcings is larger than the response to lower latitude forcings. The efficacy for changes of Arctic sea ice albedo is >3. In additional runs not shown here, we found that the efficacy of albedo changes in Antarctica is also >3.

We define the effective forcing as $F_e = F_a E_a$. F_e provides a better measure of expected climate impact of a climate forcing mechanism than does either F_i or F_a. The effective forcing for the assigned snow albedo change in the most realistic cases 1 and 2 is $F_e \sim 0.6$ W/m^2 in the Northern Hemisphere or $F_e \sim 0.3$ W/m^2 globally.

Transient Simulations. The above results suggest that a significant fraction of global warming in the industrial era might be due to soot's effect on snow albedo. Thorough investigation of this issue requires knowledge of soot deposition on snow as a function of geography and time, and the study should include other known forcings, so that results can be compared with observations.

Here we make transient calculations for 1880–2000 by using a simple soot snow/ice albedo forcing, with the limited aim of obtaining a better indication of the possible magnitude of global warming due to soot snow/ice albedo forcing. The time dependence of the forcing is taken proportional to global mean BC amount in the transport model of Koch (17) with the BC emissions history including a "technology factor" (20). The technology factor accounts for reductions in BC emissions due to increasing use of low-emission power plants and improved diesel technology. Despite the technology factor, global BC emissions increase with time, as shown by figure 3 of ref. 20.

We take the geographical distribution of the forcing as in case 1. In the Koch/Novakov BC scenario, 83% of the 1850–2000 BC increase occurs during 1880–2000. So the forcing in the transient simulation increases by $0.83 \times 0.16 \sim 0.13$ W/m^2 between 1880 and 2000. The fixed pattern of the forcing is probably not a serious flaw for our purposes, because we only examine the total change from 1880 to the present, which tends to average over fluctuations of sources.

Fig. 4 is the five-member ensemble mean T_s change, obtained

Table 3. Climate forcings and their efficacies

Experiment	Forcing, W/m²		Response,	Efficacy		
	F_i	F_a	ΔT, °C	E_i	E_a	F_c, W/m²
2× CO_2	4.05	3.63	2.57	1.00	1.00	3.63
Case 1	0.17	0.16	0.24	2.22	2.12	0.34
Case 2	0.16	0.15	0.20	1.97	1.88	0.28
Case 3	0.23	0.24	0.25	1.71	1.47	0.35
Case 4	0.04	0.04	0.10	3.94	3.53	0.14

with the Q-flux ocean with diffusive mixing of heat into the 4-km deep ocean (5). The exact pattern of simulated warming has limited meaning, given the crude spatial distribution of the forcing and the simple representation of the ocean. However, the mean warming provides an indication of what should be expected for the assumed albedo change. This global warming, 0.17°C, is a substantial fraction of observed warming (Fig. 3). Even if our snow/ice albedo forcing were too large by a factor two, the impact is not negligible.

The simulated warming is largest at high latitudes in winter, because of reduced sea ice cover induced by the warming, and least in summer. These spatial and temporal characteristics are consistent with observations. This does not provide confirmation of the soot albedo effect, because other forcings may have a similar impact on the pattern of climate change. All forcings need to be investigated one-by-one in systematic simulations and compared with observations.

Discussion

"It's just good clean soot," says Dick Van Dyke as the chimney-sweep in *Mary Poppins*. Environmentalists and climatologists are not as sanguine about soot, but they have devoted greater attention to sulfates. We suggest that soot is a more all-around "bad actor" than has been appreciated.

Global Warming. Soot snow/ice albedo climate forcing is not included in Intergovernmental Panel on Climate Change evaluations. This forcing is unusually effective, causing twice as much global warming as a CO_2 forcing of the same magnitude. This high efficacy is a straight-forward consequence of positive albedo feedbacks and atmospheric stability at high latitudes.

Our estimate for the mean soot effect on spectrally integrated albedos in the Arctic (1.5%) and Northern Hemisphere land areas (3%) yields a Northern Hemisphere forcing of 0.3 W/m^2 or an effective hemispheric forcing of 0.6 W/m^2. The calculated global warming in an 1880–2000 simulation is about one quarter of observed global warming.

Our partly subjective estimate of the uncertainty in the soot albedo forcing is a factor of two, mainly because of the sparseness of accurate snow albedo and soot content measurements. In addition, if a large fraction of BC were natural, that would reduce the estimated anthropogenic forcing. However, fossil fuels and biomass burning are believed to each contribute about half of global atmospheric BC, with all of the former and much of the latter being anthropogenic. Anthropogenic soot is predominant in most of the industrial Northern Hemisphere. Greenland may be an exception, because it is downwind of Canadian forests and does not have large nearby human pollution sources. Ice core data (25, 31) indicate that the preindustrial Greenland BC amount, except for occasional large forest fires, was ≈2 ppbw, half as much as some of the 20th century measurements.

On the other hand, our calculations exclude a factor that magnifies the soot warming effect. Melting snow tends to retain aerosols, darkening the surface more in the late winter and spring when the sun is high in the sky and most effective, thus increasing absorption and lengthening the melt season (6, 32). Although laboratory experiments show that fine BC particles can escape with meltwater more readily than larger aerosols, there is a tendency for even the finest aerosols to be retained and enhance absorption in the melting season (33). This omitted factor should tend to balance the fact that a fraction of BC is natural.

Melting Ice. We suggest that soot contributes to near worldwide melting of ice that is usually attributed solely to global warming. Measurements in the Alps reveal BC concentrations as large as 100 ppbw (34, 35), enough to reduce the visible albedo by ≈10% and double absorption of sunlight (21). However, much smaller BC amounts perturb snowmelt because of positive feedbacks.

The ultimate flux perturbation at the snow surface due to BC is larger than the product of incident solar flux and the direct BC-induced snow albedo change. The BC-caused warming of the snow speeds snow "aging," i.e., the growth of grain size. More important, the warmer air causes the melt season, with its much lower albedo, to begin earlier and to last longer on sea ice and land ice. In our climate model, as illustrated in Fig. 5, which is published as supporting information on the PNAS web site, these feedbacks more than double the direct soot flux perturbation in permafrost, sea ice, and ice sheet regions, where there is little or no shielding of snow by vegetation. This effect may be important on glaciers and the lower reaches of ice sheets, where the added meltwater not only reduces the albedo but also lubricates nonlinear dynamic processes of glacier disintegration (36).

The soot albedo effect operates in concert with regional warming in most of the world, hindering empirical distinction of climate and soot contributions. However, there has been little warming in China, including Tibet, over the past 120 years (Fig. 3), yet glaciers there are retreating rapidly (37). The brown haze over India (22), heavy with fossil fuel and biofuel soot, reaches to the Himalayas. If prevailing winds deposit even a fraction of this soot on glaciers, the snow BC content could be comparable to that in the Alps. Measurements of glacial soot are needed.

Sea ice drafts measured by submarines in the Arctic suggest a thinning of the ice by about a meter during 1958–1997 (38). The Arctic albedo changes (Table 1 and Fig. 6, which is published as supporting information on the PNAS web site) suggested by soot measurements in the 1980s (6) cause surface flux changes (Fig. 5) that are a substantial part of those needed to account for observed sea ice thinning (38).

Spring snowmelt on tundra in Siberia, Alaska, Canada, and Scandinavia has trended earlier, by 2–5 weeks, in recent decades (32). The magnitude of this shift exceeds that in climate model simulations with realistic global warming (5), suggesting that other mechanisms contribute to the early snow disappearance.

Dangerous Anthropogenic Interference. The United Nations Framework Convention on Climate Change states the goal of stabilizing atmospheric composition at a level that avoids "dangerous anthropogenic interference" (DAI) with the climate system, but it does not define the level of climate change constituting DAI. We argue (J.H. in *Can We Defuse the Global Warming Time Bomb?*, a presentation to the Council on Environmental Quality, www.giss.nasa.gov/research/forcings/ceq_presentation.pdf) that the level of DAI will be set by the need to preserve global coastlines and that this implies the need to keep additional global warming less than ≈1°C. Others have suggested that the limit on global warming required for ice sheet stability is 2°C or larger (39). Regardless of the accuracy of our estimate, the soot snow albedo effect enters the DAI discussion in two ways.

First, it raises global temperature closer to the level of DAI. Our estimate for the equilibrium global warming of current soot levels is ≈0.2°C, most of which is already achieved. If snow albedos were restored to their pristine values, a future cooling tendency of that magnitude would be introduced, partially countering opposing warming forces and thus helping to keep us from reaching the level of DAI.

Second, it lowers the global temperature at which DAI occurs, because soot is another factor, besides climate change, that affects ice melting and sea level rise. Soot deposition increases surface melt on ice masses, and the meltwater spurs multiple radiative and dynamical feedback processes that accelerate ice disintegration. Soot-caused albedo change lengthens the melt season and increases melt rates, thus further reducing albedo and spurring nonlinear feedbacks by increasing water flow through crevasses and moulins, which speeds freeze–thaw ice break-up and lubrication of the ice sheet base (36). An adequate model of these processes does not exist, but the sense of the net effect is unambiguous. It is surely large on glaciers with BC amounts of 10–100 ppbw and probably small

on Antarctica. Soot amounts as little as 2 ppbw on Greenland may affect visible albedo as much as 1%, which is not a negligible contribution compared with the balance of fluxes that determine ice sheet mass balance.

Soot's Other Roles. As brilliantly as Dick Van Dyke plays both the decrepit banker and the chimney-sweep in *Mary Poppins*, so ignominiously does soot play multiple roles in the environment. Topics mentioned below are expanded in a workshop report (40).

Soot particles are like tiny sponges that soak up toxic organic material and metals in fossil fuel and biomass burning. Because they are so small, they penetrate human tissue deeply when breathed into the lungs. Soot is suspected of being a major contributor to approximately one million premature deaths per year that are blamed on particulate air pollution (40). Soot is the aerosol most responsible for reducing atmospheric transparency and visibility, by so much in India and China that agricultural productivity is reduced an estimated 10–20% (41) with additional loss from soot deposited on plant leaves (42). Soot is also esthetically displeasing because it is responsible for a brown appearance of urban hazes and soiling of buildings.

Soot may affect regional climate, as well as global climate (22, 43). It has been suggested that the heavy concentration of soot over China and India may be responsible for a trend toward increased flooding in the south and drought in the north (43).

Summary. The soot effect on snow albedo may be responsible for a quarter of observed global warming. Restoration of snow albedos to something approaching pristine preanthropogenic values would have the double benefit of reducing global warming and raising the global temperature threshold at which dangerous anthropogenic interference with climate occurs.

Future trends of the soot climate forcing could be positive or negative. The SHEBA (Surface Heat Budget of the Arctic Ocean) finding of less BC in the Arctic in 1998 than existed in the 1980s is qualitatively consistent with a reduction of Russian BC emissions (20). However, the generality of the SHEBA is dubious, because the measurement was local for a brief period and, for reasons mentioned above, future BC emissions in the Arctic could increase. There is a need for more complete measurements of BC in Arctic snow and better quantification of the links to snow albedo, melting ice, and climate change.

On the optimistic side, technology is within reach that could greatly reduce soot, restoring snow albedo to near pristine values, while having multiple other benefits for climate, human health, agricultural productivity, and environmental esthetics. Already, soot emissions from coal are decreasing in many regions with transition from small users to power plants with scrubbers. The largest source of soot in developed countries is now diesel fuel, and in developing countries biofuels are also important. Much cleaner diesel engines and biofuel technologies are possible (40). There are opportunities for scientific and technologic cooperation among developing and developed countries with mutual benefits.

The substantial role inferred for soot in global climate does not alter the fact that greenhouse gases are the primary cause of global warming in the past century and are expected to be the largest climate forcing the rest of this century.

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