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Publication Supplements

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Following are an additional table (4) and two figures (5 and 6), plus explanatory text.

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Table 4

Measured BC amounts in surface snow and precipitation, with the corresponding change of snow albedo at 470 nm (ΔA_v) based on figure 2 of ref. 1. (Note: ppbw = parts per billion by weight.)

Location	Description	BC amount, ppbw	Calculated ΔA_v, %, external/internal
Arctic region, 1980s (sea ice, Alaska, Canada, Sweden, Spitzbergen) (2)	Measurements from a range of fresh, nonfresh, and windblown snow, firn, and snow-pit	10 (low)	0.8/1.5 (new snow) 2.5/4.5 (old snow)
		30 (mean)	1.9/3.2 (new snow) 6.0/9.5 (old snow)
		50 (high)	2.7/4.2 (new snow) 8.2/11.5 (old snow)
SHEBA, 1998 (central Arctic) (3)	Upwind of camp	4	
	Downwind of camp	35	
Greenland (2)	Snow-pits	2 (low) (2)	0.3/0.5 (new snow) 0.7/1.2 (old snow)
		9 (high) (2)	0.75/1.3 (new snow) 2.3/4.2 (old snow)
	Fresh wind-blown	6.2 (single measurement) (2)	0.5/0.9 (new snow) 1.7/3 (old snow)
Antarctica	South Pole (4) - Pristine area	0.2	0.05/0.1 (new snow) 0.1/0.2 (old snow)
	- Downwind of station	3	0.3/0.5 (new snow) 0.8/1.5 (old snow)
	Ross Ice Shelf (5)	2.5	
Cascade Mountains (Washington) (6)	Melting snow, rounded needles, ≈0.05-mm radius observed A_v ~ 92%	22	1.5/2.5 (new snow) 4.5/7.7 (old snow)
	5 cm new wet snow over granular observed A_v ~ 90%	59	3/5 (new snow) 9/14 (old snow)
Alps, 1.8-3.3 km (7-9)	A_v ~ 94.5% for BC = 120 ppbw	120 Radius ~ 0.15 μm	4.8/7 (new snow)
	A_v ~ 93% for BC = 180 ppbw	180 Radius ~ 0.15 μm	6/8.5 (new snow)
	A_v ~ 84% for BC = 280 ppbw	280 Radius ~ 0.4 μm	7.5/11 (new snow)
Lithuania (10)	Fourteen rural sites		

		100 (mean) 150 (cold season)	
Paris, France (11)	Urban, rain	333	
Mace Head, Ireland (11)	Marine site, rain	31	
Sweden (12)	Rural	100	
Michigan (13)	Rural	72	
Detroit (14)	Urban	160	

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Figure 5

Download Fig. 5 (168 kB PDF): Simulated 1880-2002 change of net shortwave radiation at the surface for the BC snow/ice albedo forcing, as in Fig. 4.

Figure 6

Download Fig. 6 (110 KB PDF): Simulated 1880-2002 surface albedo change for the BC snow/ice albedo forcing, as in Fig. 4.

Supplementary Text

Table 1 summarizes measurements of black carbon (BC) amount for the Arctic, Greenland, other Northern Hemisphere land, and Antarctica. More extensive observations of BC amounts in snow or cloud particles are provided in Table 4. Estimates of the impact of these BC amounts on visible snow albedos, ΔA_v, based on figure 2 of ref. 1, for external mixing are included in the right column. New snow has grain radius 0.1 mm and old snow has grain radius 1 mm. The estimate for internal mixing increases the BC absorption coefficient, or effective amount, by a factor of two.

Scavenging Ratio. Measurements of soot content in snow are not as widespread as measurements of the concentration of soot in air. Clarke and Noone (2) compared their widespread measurements of BC content in snow (ng/g) in the Arctic with measurements of BC (ng/m^3 at sea level pressure) in Arctic air (3), finding an average scavenging ratio, SR, of ≈160. This was consistent with SR values of 100-200 found at Dye 3 in Greenland (4). Later simultaneous measurements of BC concentration in air and snow for six Arctic snowstorms (5) yielded SR of 60-170 with mean SR ~ 100.

Measurements of BC concentration in Arctic air during times of Arctic haze pollution (winter/spring) are typically 250-500 ng/m^3 and much less in other seasons (6). Ambient BC concentrations at lower latitudes over Northern Hemisphere nonurban land areas tend to be larger than the mean Arctic values (7-9). Thus, the atmospheric measurements of BC, together with the scavenging ratio measurements (2, 3), offer support for use of a larger snow albedo effect at lower latitudes than in the Arctic.

Global Climate Model Calculations. The albedo of snow and sea ice in the current Goddard Institute for Space Studies climate model is defined in six spectral intervals, one visible (0.3-0.77 μm) and five in the near infrared (0.77-4.0 μm) with the extreme values for fresh snow, melting snow, bare sea ice, and meltwater ponds fit to observations (10). Snow is assumed to partially cover sea ice for gridbox mean snow depths <10 cm (proportional to snow depth) and to completely cover sea ice for greater depths. Albedos for dry and wet snow are set separately (11), and there is a small snow aging effect (12). Snow becomes wet in response to surface melting or rainfall. Sea ice albedo depends on sea ice thickness and fraction of melt ponds (11, 13). Snow albedo over land has a snow age effect to simulate the effect of increasing grain size (14) and a snow fraction based on a three-layer snow model and topographic roughness (15).

Fig. 6 shows the change in surface albedo in our transient climate simulation for 1880-2002 for March to April and May to June. The specified spectral-mean snow albedo changes were 0.83 × 1.5% ~ 1.25% for snow/ice in the Arctic Ocean and 0.83 × 3% ~ 2.5% for snow over Northern Hemisphere land. The surface albedo changes are diminished over land by vegetation masking of the snow (14), which is very effective in forested areas but ineffective in tundra regions with little vegetation. The warming of the surface and air due to the soot albedo effect enhances the surface albedo changes over both land and ocean because of enhanced snow aging (increased grain size) and especially because of the earlier onset of spring melt. Thus, the soot-lowered albedo and increased temperatures initiate positive feedbacks via earlier snowmelt and rainfall.

Fig. 5 shows the change in the net short-wave heating of the surface for the same periods as in Fig. 6. Note that the regional flux changes at the surface are as much as 10 W/m^2 and more. This compares with a typical annual-mean forcing of <1 W/m^2 (Fig. 1). Clarke and Noone (2) calculated a maximum May-to-June surface flux perturbation of ≈ 6 W/m^2 for an assumed snow albedo perturbation of 2%. Given that our soot-imposed albedo change for the period 1880-2002 was only ≈ 1.25% (over sea ice), this means that we find a surface heating about twice as large as that calculated by Clarke and Noone (2). The cause of this difference is the positive feedbacks in our climate model that reduce the albedo further, especially the acceleration of the summer melt season. Our results are not in disagreement with those of Clarke and Noone (2), because they mentioned in their concluding remarks the likelihood that such positive feedbacks would enhance the soot albedo effect.

These positive feedbacks, especially the acceleration of the melt season, have practical implications. The perturbations of surface fluxes are largest in the regions of sea ice, permafrost, glaciers, and the low altitude portion of the Greenland ice sheet that is subject to summer melt. This suggests that soot may contribute to thinning of sea ice (16), melting permafrost, glacier retreat, and accelerating movement of Greenland ice (17).

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