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Abstract

A new physically-based parameterization of black carbon (BC) in snow was developed and implemented in the Canadian Atmospheric Global Climate Model (CanAM4.2). Simulated BC snow mixing ratios and BC snow radiative forcings are in good agreement with measurements and results from other models. Simulations with the improved model yield considerable trends in regional BC concentrations in snow and BC snow radiative forcings during the time period from 1950–1959 to 2000–2009. Increases in radiative forcings for Asia and decreases for Europe and North America are found to be associated with changes in BC emissions. Additional sensitivity simulations were performed in order to study the impact of BC emission changes between 1950–1959 and 2000–2009 on surface albedo, snow cover fraction, and surface air temperature. Results from these simulations indicate that impacts of BC emission changes on snow albedos between these two decades are small and not significant. Overall, changes in BC concentrations in snow have much smaller impacts on the cryosphere than the net warming surface air temperatures during the second half of the 20th century.

1 Introduction

Snow plays an important role in the land surface radiation budget and therefore climate. Snow albedo can be strongly reduced by absorbing aerosols such as black carbon (BC), brown organic carbon, or dust which have been deposited on the snow through dry and wet deposition and cause warming of the snowpack (Warren and Wiscombe, 1980, 1985; Hansen and Nazarenko, 2004; Hadley and Kirchstetter, 2012).

BC is produced by incomplete combustion of fossil and biomass fuels and is emitted into the atmosphere from a wide range of natural and anthropogenic sources. Total emissions of BC from fossil fuel and biofuel combustion have increased strongly from 1850 to 2000 (Bond et al., 2007; Skeie et al., 2011). In addition to effects on snow

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albedo, BC can affect Earth's climate through direct radiative, semi-direct and indirect (cloud) effects.

An initial albedo reduction by BC in snow can be amplified by positive feedbacks involving snow processes, for example:

- i. a warmer snowpack causes increases in snow grain effective radius (Flanner and Zender, 2006) which thereby causes further reductions in snow albedo,
- ii. larger snow grain sizes in a warm snowpack leads to a stronger BC snow radiative forcing (Warren and Wiscombe, 1980; Hadley and Kirchstetter, 2012),
- iii. BC tends to become more highly concentrated near the surface of the snowpack when snow melts (Clarke and Noone, 1985; Doherty et al., 2010), and
- iv. an earlier snow melt onset may cause an earlier exposure of a darker underlying surface.

Different models have been used to simulate impacts of BC on snow and climate. Hansen and Nazarenko (2004) used the few available measurements of BC in snow at that time to determine snow albedo changes in the Arctic, Antarctica, Greenland, and other regions in the Northern Hemisphere. These albedo change estimates were used to calculate the radiative forcing (RF) and climate response. Jacobson (2004a) estimated BC concentrations in snow prognostically and used them to calculate the change in albedo and RF. More recently, Flanner et al. (2007, 2009), Koch et al. (2009b), Rypdal et al. (2009b), Skeie et al. (2011), and Jiao et al. (2014) used prognostic modelling methods to simulate BC in snow and interactions with climate.

Flanner et al. (2007) investigated impacts of BC in snow on snow radiative and microphysical processes. They performed simulations for strong and weak boreal fire years and suggested that both BC emissions and the treatment of snow aging represent large sources of uncertainty in the simulations of BC in snow. In a recent study of Jiao et al. (2014), BC deposition fields from 25 global aerosol models were used to investigate effects of BC transport and deposition processes on BC concentration in

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snow. The analysis reveals that these processes are more important for differences in simulated BC concentrations in snow than differences in snow melt water scavenging rates or emissions in models.

Widespread surface melting of the Greenland ice sheet in 1889 and 2012 was recently attributed to warm temperatures combined with reductions in surface albedo resulting from deposition of black carbon from Northern Hemisphere forest fires (Keegan et al., 2014). On the other hand, the extent to which absorption of visible radiation by BC in snow might have contributed to rapid warming of the mid to high latitude land surface and decline in Eurasian springtime snow cover (Déry and Brown, 2007) is presently unknown. Studies of BC in snow are complicated by the fact that anthropogenic emissions of BC are non-uniform in space and time, compounded by large variations in atmospheric transport of BC at high latitudes (Stohl et al., 2006). In addition, there is large inter-annual variability in vegetation fire emissions of BC. Variations in total BC emissions are also compounded by changes in snow cover extent (SCE) and depth owing to variations in temperatures and snowfall. Consequently, concentrations of BC in snow are highly variable and climate effects are difficult to quantify robustly.

Some of the recent studies focused on RFs during individual years (Flanner et al., 2007, 2009) while Hansen et al. (2007a), Koch et al. (2009a, 2011), and Skeie et al. (2011) analyzed differences between pre-industrial and present-day processes and climate. Some of the studies produced evidence for potentially large impacts of BC in snow on radiative forcings and climate. For instance, Flanner et al. (2009) concluded that fossil fuel and biofuel emissions of BC induce a large springtime snow cover loss over Eurasia as a consequence of strong snow-albedo feedbacks and large BC emissions from Asia, with a magnitude similar to impacts of anthropogenic carbon dioxide concentrations on SCE. According to Koch et al. (2009a), black carbon in snow has caused an increase in surface air temperature between 1890 and 1995 that more than balances a decrease in surface air temperature caused by direct radiative effects of aerosols during this time period. However, in a later study, Koch et al. (2011) found only

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small impacts of BC on snow and ice during the twentieth century. In a related study, Holland et al. (2012) concluded that the deposition and cycling of aerosols in sea ice cause the sea ice to melt and may therefore play a role in late-twentieth-century Arctic sea ice trends. On the other hand, Goldenson et al. (2012) found that deposition of aerosols have small impacts on Arctic sea ice and climate.

The purpose of this work is to study impacts of BC emissions on cryospheric and climate processes in a recent version of the Canadian Atmospheric Global Climate Model (CanAM4.2). We apply a physically based parameterization of the snow albedo in combination with new parameterizations for BC concentrations in snow and effective radius of snow grains, which are described in Sect. 2. Simulations of BC in snow, RFs of BC in snow, and changes in snow cover fraction (SCF) and surface air temperatures from 1950s to 2009s are discussed in Sect. 3.

2 Modelling approach

An interactive model treatment of BC emissions, transport, and deposition is required in order to simulate the RF of BC in snow and feedbacks between aerosol, snow physical processes, and climate. The Canadian Atmospheric Global Climate Model (CanAM4.2) is similar to the earlier version CanAM4 (von Salzen et al., 2013) but uses version 3.6 of the Canadian Land Surface Scheme (CLASS) to model the energy and water balances of the soil, snow, and vegetation canopy components (Verseghy, 2012). Relative to CanAM4, there are many improvements to the simulation of snow. These include new formulations for vegetation interception of snow (Bartlett et al., 2006), for unloading of snow from vegetation (Hedstrom and Pomeroy, 1998), for the albedo of snow-covered canopies (Bartlett and Verseghy, 2015), for the limiting snow density as a function of depth (Tabler et al., 1997; Brown et al., 2006), and finally for the thermal conductivity of snow (Sturm et al., 1997). Water retention in snowpacks has also been incorporated.

Other differences between CanAM4 and CanAM4.2 that are relevant to this study are the introduction of an aerosol microphysics scheme (von Salzen, 2006; Ma et al., 2008;

Peng et al., 2012), a higher vertical resolution in the upper troposphere, a reduced solar constant (1361 W m^{-2}) and an improved treatment of the solar continuum used in the radiative transfer. However, the model retains the same T63 spectral resolution, with 49 vertical levels.

Different types of natural and anthropogenic aerosols are considered, including sulfate, black and organic carbon, sea salt, and mineral dust. Parameterizations for emissions, transport, gas-phase and aqueous-phase chemistry, and dry and wet deposition account for interactions with simulated meteorological variables for each individual grid point and time step. Hydrophobic and hydrophilic black and organic carbon aerosol are considered, with a specified lifetime, representing aerosol aging processes of hydrophobic carbonaceous aerosols (Croft et al., 2005). Fluxes and properties of the air (downwelling longwave and shortwave radiation, precipitation, air pressure and temperature, specific humidity and wind speed) are passed to CLASS which in turn computes the density, liquid water content, mass and temperature of a single layer of snow at each time step. The snow albedo calculation in CLASS does not explicitly account for snow metamorphism nor albedo reductions by BC which requires addition of new parameterizations (Sect. 2.1).

2.1 Radiative and snow microphysical processes

A parameterization for snow albedo which accounts for contributions of BC-snow mixing ratio and the effective snow grain size was developed (Cole et al., 2015). For the sake of completeness the approach is described in brief. Lookup tables for snow albedo and transmission are used, which are functions of snow water equivalent, underlying surface albedo, solar zenith angle, snow grain size and BC concentration. These tables are created for incident radiation that is purely direct beam or diffuse and for each radiation band in the CanAM4.2 radiative transfer model. Table contents are computed using the DISORT radiative transfer model (Stamnes et al., 1988). For each possible combination in the tables DISORT calculations are performed for 280 different wavelengths between 0.25 and 4.0 microns which are then averaged over each band, weighted by

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either incident direct or diffuse radiation. For each of the calculations the optical properties of snow and BC are computed using Mie calculations. The size distribution and refractive indices for BC are taken to be that in Bäumer et al. (2007) to maintain consistency between the atmosphere and snow radiative effects.

The response of snow albedo to the presence of BC, relative to pristine snow, is shown in Fig. 1 for a semi-infinite snow layer in which BC is externally mixed and illuminated by direct beam radiation at a 60° zenith angle. For comparison, we show results from the Snow, Ice, and Aerosol Radiation (SNICAR) model (Flanner et al., 2007, 2009) using the same configuration. It is clear that the Cole et al. (2015) snow albedo model produces an albedo response to BC in snow that is roughly 25 % less than SNICAR for all BC concentrations and snow grain sizes. This can be largely attributed to the choice of BC optical properties. In SNICAR the mass absorption coefficient at 550 nm is $7.5 \text{ m}^2 \text{ g}^{-1}$, following recommendations by Bond and Bergstrom (2006), while the parameterization of Bäumer et al. (2007) produces a mass absorption coefficient of $4.9 \text{ m}^2 \text{ g}^{-1}$ at 550 nm. Recreating the lookup tables in Cole et al. (2015) with BC refractive indices that are consistent with SNICAR (Flanner et al., 2012) result in similar albedo responses due to BC in snow (not shown).

We consider snow over bare ground as well as on continental ice sheets and sea-ice for the development of new parameterizations for snow grain sizes and BC concentrations in snow. The existing parameterizations for snow in CLASS (3.6) are used to simulate properties of a single layer of snow with variable depth. Since vertical profiles of snow physical properties are not resolved in CLASS and extinction of solar radiation mainly occurs near the surface of snow layers, parameterizations of snow microphysical processes need to account for unresolved processes in the snowpack. The Snow Model Intercomparison Project (SnowMIP) experiments have shown that the performance of CLASS for simulations of the seasonal evolution of snow on land is comparable to that of a multi-layer snowpack model (Brown et al., 2006), despite the use of a single layer of snow.

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2.1.1 Snow metamorphism

Snow grain size influences the snow albedo and thermal conductivity as well as chemical interaction between snow and atmosphere (Domine et al., 2008). It generally increases over time as the snowpack ages and therefore causes reductions in snow albedo, especially in the near infrared (Wiscombe and Warren, 1980). The visible albedo, however, depends more strongly on concentrations of snow impurities such as black carbon (BC) and mineral dust (Warren and Wiscombe, 1980). Larger snow grains amplify the rate of snow albedo reduction by BC in snow (Hadley and Kirchstetter, 2012; Aoki et al., 2011). This is because of deeper radiation penetration in coarser snow grains (Warren and Wiscombe, 1980). In addition, the growth of snow grains contributes to snow albedo feedbacks as the snowpack warms (Flanner and Zender, 2006). Therefore, an accurate model representation of snow grain sizes is potentially important for simulations of the effects of BC on climate.

Snow with non-spherical complex particles can be represented as a collection of spheres, which conserves the volume to surface area ratio well (error less than 5%) (Grenfell and Warren, 1999; Grenfell et al., 2005). The effective radius, R_e , the surface area-weighted mean radius of a collection of spherical particles, is given by

$$R_e = \frac{3}{\rho_i S}, \quad (1)$$

where $\rho_i = 917 \text{ kg m}^{-3}$ is the bulk ice density. The snow specific surface area (S) is defined as the ratio of the surface area per unit mass of snow in $\text{m}^2 \text{ kg}^{-1}$, which defines atmosphere-snow chemistry interactions and surface energy balance through snow albedo and turbulent fluxes.

The size of the snow grains depends on the macrophysical state of the snowpack which is determined by its temperature, type, density, and amount of liquid water. We consider three mechanisms for the temporal evolution of snow grain size in a layer at the top of the snowpack: dry aging, wet aging and snow deposition. Other processes

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such as wind and sintering that can also affect snow grain size are not considered owing to a lack of available parameterizations and simplicity of the CLASS snow model.

For dry aging of snow grains, we use a physically-based approach by Flanner and Zender (2006) which predicts the temporal evolution of R_e (m), of dry snow due to water vapour transport in the snowpack, which is driven by differences in snow grain curvature and temperature gradient. The calculation depends on snowpack temperature, vertical temperature gradient, and snow density. It is formulated as

$$\frac{dR_e}{dt} = \frac{dR_e}{dt} \Big|_0 \left(\frac{\tau}{R_e - R_{e_0} + \tau} \right)^{\frac{1}{\kappa}}, \quad (2)$$

where $\frac{dR_e}{dt} \Big|_0$ (ms^{-1}) is the initial rate of change of the effective radius, and τ and κ are empirical parameters which account for the spatial distribution of grain sizes and irregularity in grain spacing and are specified according to the original version of the parameterization. R_{e_0} is the effective radius of fresh snow and is set to $54.53 \mu\text{m}$ corresponding to $\text{SSA}_0 = 60 \text{ m}^2 \text{ kg}^{-1}$ (Flanner et al., 2007) based on observational data of Legagneux et al. (2004).

The parameterization of wet aging accounts for the increase in snow grain size by rounding due to presence of liquid water within the snowpack. This process is modeled using an empirical equation expressed in terms of R_e (μm),

$$\frac{dR_e}{dt} = \frac{10^{18} C_1 f_{\text{liq}}^3}{4\pi R_e^2}, \quad (3)$$

based on experimental data, (Brun, 1989; Oleson et al., 2010). Here $C_1 = 4.22 \times 10^{-13}$ and f_{liq} is the mass liquid water fraction given by $\frac{W_{\text{liq}}}{W_{\text{liq}} + W_{\text{ice}}}$ where W_{liq} and W_{ice} are the masses of water and ice in (kg m^{-2}), respectively. Flanner et al. (2007) also use this parameterization for wet aging.

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We parametrize the effects of snowfall on mean snow effective radius (m) in a shallow layer at the surface of the snowpack according to

$$\frac{dR_e}{dt} = -\frac{1}{\tau_d}(R_e - R_{e_0}), \quad (4)$$

where τ_d is an adjustment time scale which is given by

$$\tau_d = \frac{\rho_s z_s}{F_s} \quad (5)$$

for $F_s > 0$. Here, z_s (m), ρ_s (kg m^{-3}) and F_s ($\text{kg m}^{-2} \text{s}^{-1}$) are instantaneous snow depth, density of snow layer, and snowfall flux, respectively. If there is no snowfall, $F_s = 0$, then the effective radius remains constant. We use $z_s = 25$ cm as the maximum depth of the surface snow layer in parameterizations of snow grain size, corresponding to a typical depth of solar penetration in snow (France et al., 2011). If the depth of snowpack is shallower than 5 cm, the effective radius is kept constant. We found a weak sensitivity of results to the assumed depth of the snow layer between 15 and 25 cm, which does not affect the conclusions presented in the following.

There are few field measurements for effective radius of snow with vertical profiles available which can be used to validate parameterizations for snow grain sizes. The measurements which we use are from six locations and are summarized in Table 1. The averaged simulated effective radius corresponding to the location and month of the measurements is 0.29 ± 0.18 mm. This is comparable to an averaged observed effective radius of 0.22 ± 0.11 mm.

2.1.2 BC concentration in snow

The parameterization of BC concentrations in snow is based on a BC mass budget for a shallow homogeneous layer of snow at the top of the snowpack, with the additional simplifying assumption that there is no transfer of BC between the layer and underlying snow except for removal of BC by meltwater scavenging. The BC concentration

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is determined by accumulation rates of BC and snow through deposition processes. In addition, partial scavenging of BC by melting snow water will act to increase the BC concentration in the top layer. Processes such as the advection of BC and snow by wind or sea ice and effects of sublimation of snow on BC concentrations are not considered.

The parameterization has the following components:

- BC deposition: the BC concentration (kg m^{-3}) in the snow layer increases by BC deposition and decreases by snowfall. If changes in snow density are omitted and the layer depth (z_s) is constant,

$$\frac{dC}{dt} = \frac{F_b}{z_s} - \frac{CF_s}{\rho_s z_s}, \quad (6)$$

where F_b is the total black carbon deposition flux ($\text{kg m}^{-2} \text{s}^{-1}$), resulting from dry deposition, convective and large-scale wet deposition.

The BC snow mass budget can also be expressed as an adjustment of the concentration towards an effective surface BC concentration,

$$\frac{dC}{dt} = -\frac{1}{\tau_d}(C - C_0), \quad (7)$$

where the adjustment time scale (τ_d) is given by Eq. (5). The effective surface concentration (C_0) is given by $C_0 = \rho_s \frac{F_b}{F_s}$. If there is no snowfall, then the change in BC concentration is given by $\lim_{F_s \rightarrow 0} \frac{dC}{dt} = \frac{F_b}{z_s}$.

- Scavenging: Snow meltwater can scavenge hydrophilic BC out of the snow layer. However, BC may accumulate near the surface during the melt as observed in Conway et al. (1996), Xu et al. (2006), and Doherty et al. (2010) and enhance the RF due to incomplete scavenging (Flanner et al., 2007). We parametrized this

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process by assuming the BC concentration change in time to be proportional to the meltwater flux and concentration of BC in the snow surface layer,

$$\frac{dC}{dt} = (1 - s) \frac{F_w}{\rho_s z_s} C. \quad (8)$$

Here, F_w ($\text{kg m}^{-2} \text{s}^{-1}$) is the meltwater flux out of the snow layer and $0 \leq s \leq 1$ is the scavenging factor. For $s = 1$, in case of a perfectly efficient scavenging efficiency, the concentration of BC on top of the snow layer is not changed by melting of snow. In the other limiting case of $s = 0$, BC is not removed by snow meltwater and therefore tends to accumulate in the snow surface layer over time. We choose $s = 0.1$ for which the simulated BC are in good agreement with measured BC (Doherty et al., 2010). Rypdal et al. (2009b) and Skeie et al. (2011) assume that all BC remains at the snow surface as snow melts while Flanner et al. (2007, 2009) and Jiao et al. (2014) apply a scavenging efficiency of 0.2 for hydrophilic BC and 0.03 for hydrophobic BC, based on Conway et al. (1996). In contrast, Hansen et al. (2005) and Koch et al. (2009b) did not include any contribution of snow melt to BC snow concentrations.

2.2 Simulation setup

In order to investigate BC in snow and effects on cryospheric and climate processes, two simulations were performed from 1950–2009 with one year of model spin-up for 1949. In the control simulation (CTRL) snow aging and BC concentration in the snow are calculated prognostically using the parameterizations described in the previous sections. A sensitivity experiment (EXP) based on the control simulation is performed by applying specified monthly climatological BC concentrations in snow rather than time varying prognostic values. The climatological BC concentrations in this simulation were obtained by temporal averaging of BC concentration in snow over the first decade of the control simulation, 1950–1959, for times when snow is present. Consequently,

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differences between the two simulations can be attributed to differences in simulated BC radiative forcing in snow evolving over time relative to the 1950s.

Simulations were performed following specifications outlined by Taylor et al. (2011). Monthly mean sea surface temperature (SST) and sea-ice boundary conditions were used for the model lower boundary (Hurrell et al., 2008). These are the same boundary condition data that were required for the fifth Coupled Model Intercomparison Project CMIP5. Time-varying monthly averaged concentrations for CO₂, CH₄, O₃, N₂O, effective CFC11 and CFC12 were those of the historical period used in the CMIP5 project and extended to 2010 using the Representative Concentration Pathway (RCP4.5) scenario (Moss et al., 2010).

Emissions of BC and other aerosol compounds were provided using a combination of linearly interpolated monthly emission fluxes from different data sets in an attempt to include the most recent improvements in newly available emission data sets. Anthropogenic emissions in 2005 and subsequent years were specified according to the Evaluating the Climate and Air Quality Impacts of Short-Lived Pollutants (ECLIPSE) inventory, version 4a (Klimont et al., 2015a, b). Monthly varying emissions for forest and grass fires from GFED, version 3.1, were used for 1997 and subsequent years (van der Werf et al., 2010). Aircraft and shipping emissions and emissions before the first year of these data sets were taken from the CMIP5/RCP data base (Lamarque, 2010; Moss et al., 2010), with linear interpolation of the data sets in time between available start and end years.

With this approach, total global mean emissions of BC in CanAM4.2 are 4.87 Tg year⁻¹ during 1950–1959 and 8.82 Tg year⁻¹ during 2000–2009. The latter agrees well with the CMIP5/RCP data emissions during this time period (available at <http://tntcat.iiasa.ac.at:8787/RcpDb/dsd?Action=htmlpage&page=compare>). However, differences in regional emission patterns exist owing to improved temporal resolution, addition of new emission sectors and other revisions in the updated data set.

By using the observation-based SSTs and sea ice extent according to the CMIP5 approach, simulated climate conditions and BC concentrations are directly comparable

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to observations. However, this implies that feedbacks between BC in snow, oceanic heat transport, and sea ice are not accounted for in the simulations. Implications of this approach will be discussed in detail in the following.

Ensemble mean results are used to account for natural atmospheric variability in the simulations. We performed simulations with 8 ensemble members for each of the control and experiment simulations. Note that ocean and sea ice conditions are identical for each ensemble member. This implies that differences in ensemble mean results will represent only the portion of variability of the natural climate system which is associated with variations in atmospheric and land surface processes, disregarding ocean and sea ice processes.

2.3 Simulated and observed BC in snow and ice

2.3.1 Deposition of BC

To simulate the BC concentration in snow accurately, we require accurate BC deposition fluxes. Simulated BC deposition fluxes were compared to ice core measurements at four locations in Greenland (McConnell et al., 2007; McConnell, 2010; McConnell and Edwards, 2008) during the time period of the simulations. The ice cores were collected in 2003 in Greenland at Act2 (66.00° N, 45.2° W), D4 (71.4° N, 43.9° W), Humboldt (78.5° N, 56.8° W) and Summit (72.6° N, 38.3° W). BC contents in the ice cores were measured with a laser soot photometer (SP2) and were used to derive the deposition flux of BC from 1772 to 2003. Annually averaged BC deposition fluxes from these measurements and CanAM4.2 are shown in Fig. 2. CanAM4.2 reproduces the declining trend in the observed BC deposition flux for Greenland, which may be explained by a decrease in emissions and transport of BC from North America. The model results are in particularly good agreement with measurements at Summit and D4. However, the model underestimates deposition rates at the south-westerly locations and the trends are also weaker compared to the measurements. Note that the observed inter-annual variability is particularly large for these two sites which is likely associated

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Siberia causes localized model underestimates for Russia. Concentrations of BC in North America are well reproduced at latitudes above about 50° N and systematically underestimated at lower latitudes. The mean simulated and measured BC snow concentrations averaged over all samples temporally and spatially are 35 and 108 ng g⁻¹, respectively and the correlation between model results and observations is 0.84. Note that biases for Chinese sites contribute significantly to the differences in sample mean concentrations, which incompletely reflect conditions in the Northern Hemisphere. Despite the unprecedented size of the observational data set employed here, a rigorous validation of BC concentrations in snow in global models will require measurements over longer time periods and from additional regions. Note that the large variability in simulated BC concentrations in snow over Arctic Ocean sea ice compared to observation in Fig. 3 can likely be attributed to differences in BC mass budgets in snow over sea ice. This may be related to the omission of advection of sea ice and other processes for snow on sea ice in the model.

Biases in simulated BC snow concentrations are partly related to emissions. Models generally tend to substantially underestimate absorption aerosol optical depth for North American and Asian sites, which has been attributed to a lack of BC emissions in these regions (Koch et al., 2009b; Bond et al., 2013). This is broadly consistent with underestimates of absorption aerosol optical depth and near-surface atmospheric concentrations of BC in CanAM4.2 in these regions (not shown). Similarly, Kopacz et al. (2011) found that BC concentrations in snow were underestimated in their simulations at Chinese sites. On the other hand, there are no obvious systematic regional biases in BC concentration in snow in the Arctic in models (Bond et al., 2013). Koch et al. (2009a) and Skeie et al. (2011) extended their comparison to include measurements from Forsström et al. (2009) as well.

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stantial (Fig. 7, bottom panel). Significant reductions in sea ice fraction and snow cover fraction (SCF) over Europe and Northern Asia lead to significant reductions in surface albedos, as will be further discussed in the following. There is no obvious relationship between changes in surface albedo and BC in snow for most of the regions. However, for the Himalayas and Tibetan Plateau, the changes in surface albedos are similar to changes in surface albedos induced by BC in snow. This indicates that cryospheric changes in these regions during the time period of the simulation may have been substantially affected by changes in BC in snow. However, confidence in these results is low given the low resolution of the model.

Although impacts of changes in BC in snow on surface albedos are not statistically significant and therefore unlikely to affect local climate processes systematically, we analyzed changes in SCF and surface air temperature for consistency and completeness.

As expected, there are no statistically robust changes in annual SCF that are associated with changes of BC in snow (Fig. 8, middle panel). Reductions of SCF in Eurasia and increases of SCF in North America (NA) appear to be broadly consistent with BC emissions increase in Eurasia and reduction in NA but they are not statistically significant. However, the simulated increase in SCF in northern Russia is counter-intuitive because BC concentration has increased from 1950s to 2000s. The reductions in SCF in the Himalayas and Tibetan Plateau are consistent with an accumulation of BC in snow in these regions. In comparison, significant reductions in SCF in the control run between 1950s and 2000s in the Northern Hemisphere, especially near the boundaries of the Arctic ocean, Eurasia and Northern Asia, can be largely attributed to the specified reductions in sea ice fraction and increasing SSTs between these decades (Fig. 8, bottom panel). These results are broadly consistent with observation in Déry and Brown (2007) which show that SCF in Eurasia and North America during spring 1979–2008 has decreased by 13.5 and 6 %, respectively, since 1968–1978.

Our results appear to contradict conclusions by Flanner et al. (2009) according to which emissions of BC induce a large springtime snow cover loss over Eurasia. This is

forcing of BC in snow are omitted for the latter, although they are likely to be very small (Sect. 3.1).

Similar to results for SCF discussed earlier, our simulated changes in near-surface air temperature do not agree with warming trends due to BC in snow according to Flanner et al. (2007, 2009) and Koch et al. (2009a). The simulated Arctic (global) ensemble mean temperature response to BC in snow in our simulations is -0.007 (-0.001) $^{\circ}\text{C}$ in the annual mean and -0.027 (-0.019) $^{\circ}\text{C}$ for MAM. This yields a temperature change per unit of emission flux close to $0^{\circ}\text{C year Tg}^{-1}$. In comparison, the Arctic (global) annual mean equilibrium temperature response from total BC in snow was estimated as 1.61 (0.15) and 0.50 (0.10) $^{\circ}\text{C}$ for 1998 and 2001, respectively (Flanner et al., 2007, 2009), corresponding to a global mean temperature change per unit of emission flux of 0.015 – $0.017^{\circ}\text{C year Tg}^{-1}$. According to Koch et al. (2009a), BC in snow has caused an increase in temperature of 0.5 (0.2) $^{\circ}\text{C}$ in the Arctic (global) over the twentieth century based on equilibrium climate simulations. However, Koch et al. (2011) simulated transient Arctic (global) temperature changes of BC in snow of $+0.20$ (-0.03) and -0.17 (-0.02) $^{\circ}\text{C}$ for 1890s–1990s and 1970s–1990s, respectively. Clearly, the magnitude of the simulated temperatures response depends on the choice of time period and associated changes in emissions in these studies. Temperature responses to changes in BC in snow are affected by substantial variability in climate and are poorly constrained in models.

The net impact of changes in concentrations of BC in snow on snow albedos depends on the magnitude of snow albedo feedbacks, which determine how strongly surface warming is amplified through a reduction in surface albedo resulting from warming or melting snow (Qu and Hall, 2007). Fletcher et al. (2015) showed that climate models generally reproduce observation-based estimates of snow albedo feedbacks well but tend to underestimate the temperature sensitive component of the total snow albedo feedback. This implies that where snow persists, its albedo does not reduce fast enough when the temperature warms, including warming from absorption of solar radiation by BC in snow. A very recent and preliminary analysis shows that the

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snow albedo feedbacks in models potentially contributes to differences in responses in climate to changes in BC concentrations in snow. Consequently, more rigorous model comparisons with focus on snow albedos and snow albedo feedback processes would be useful in the future.

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Table 1. Observed and simulated effective radius of snow in surface snow.

Region	Sampling period	Mean obs. (mm)	Mean model (mm)	Reference
Fairbanks, Alaska (64.87° N, 147.73° W)	Nov 2003 to Apr 2004	0.36	0.42	Taillandier et al. (2006)
Dome C, Antarctic plateau (75.1° S, 123.33° E)	Nov 2008 to Jan 2009	0.15	0.23	Gallet et al. (2012)
Barrow, Alaska (71.33° N, 203.34° E)	Mar to Apr 2009	0.13	0.58	Domine et al. (2012)
Col de Porte, France (45.30° N, 5.77° E)	Jan to Apr 2010	0.32	0.11	Morin et al. (2013)
Churchill, Canada (58.74° N, 66.18° E)	Jan to Apr 2010	0.23	0.18	Derksen et al. (2012)
Montmorency, Canada (47.33° N, 288.87° E)	Jan to Mar 2012	0.10	0.19	Unpublished data by F. Domine

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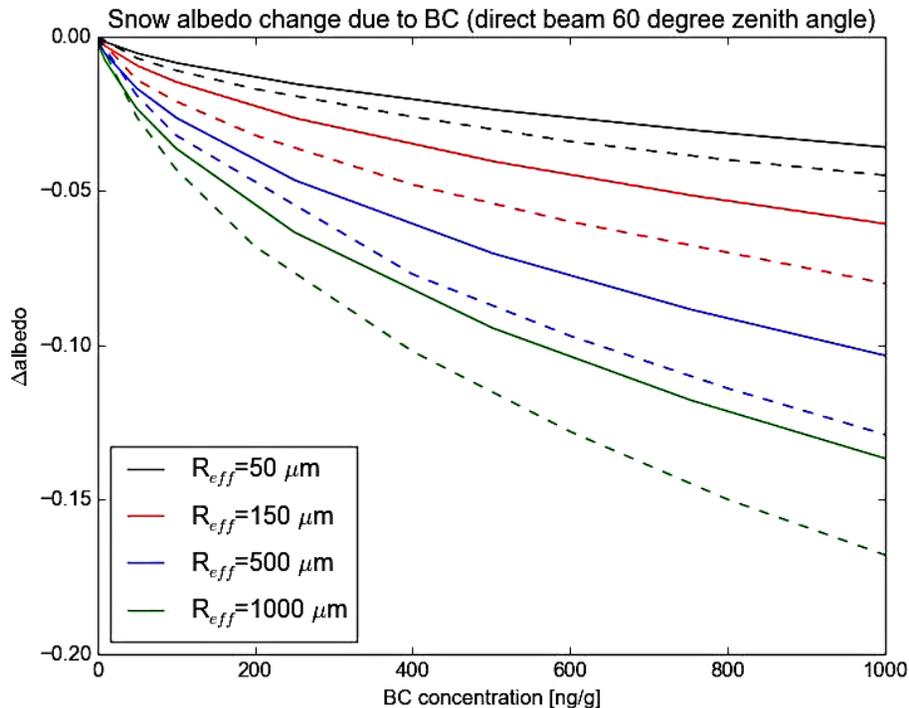


Figure 1. Response of snow albedo to the presence of BC, relative to pristine snow, for different snow grain effective radius. Full lines refer to simulations based on the approach by Cole et al. (2015) in CanAM4.2; dashed lines refer to results from simulations with SNICAR (Flanner et al., 2007, 2009).

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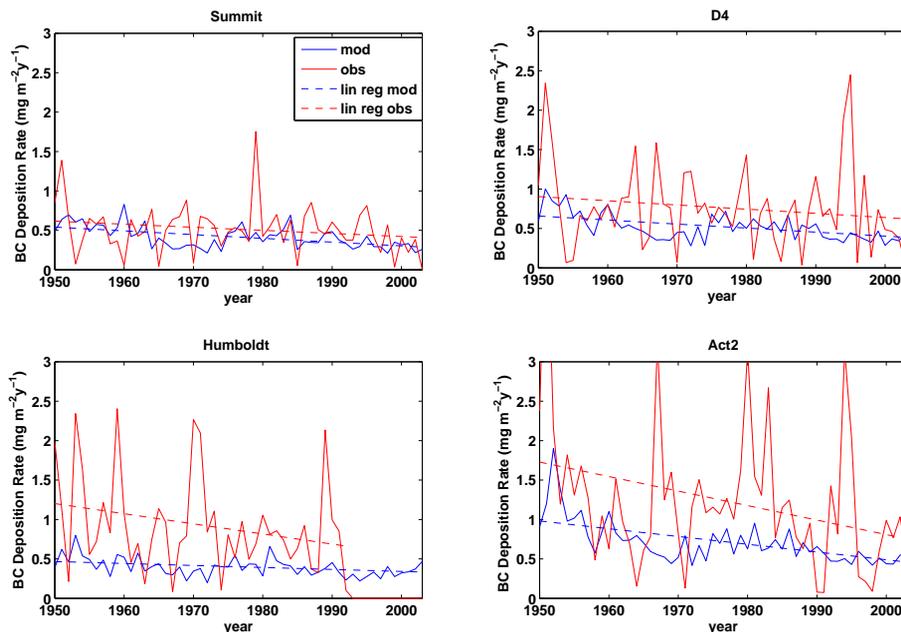


Figure 2. Measured and modelled black carbon (BC) deposition flux in ice cores in Greenland at Summit (72.6° N, 38.3° W) (top left), D4 (71.4° N, 43.9° W) (top right), Humboldt (78.5° N, 56.8° W) (bottom left) and Act2 (66.0° N, 45.2° W) (bottom right) from 1950–2009. The measurements, solid line, and their linear regression, dashed line, are in red and the model results, solid line, and their linear regression, dashed line, are in blue.

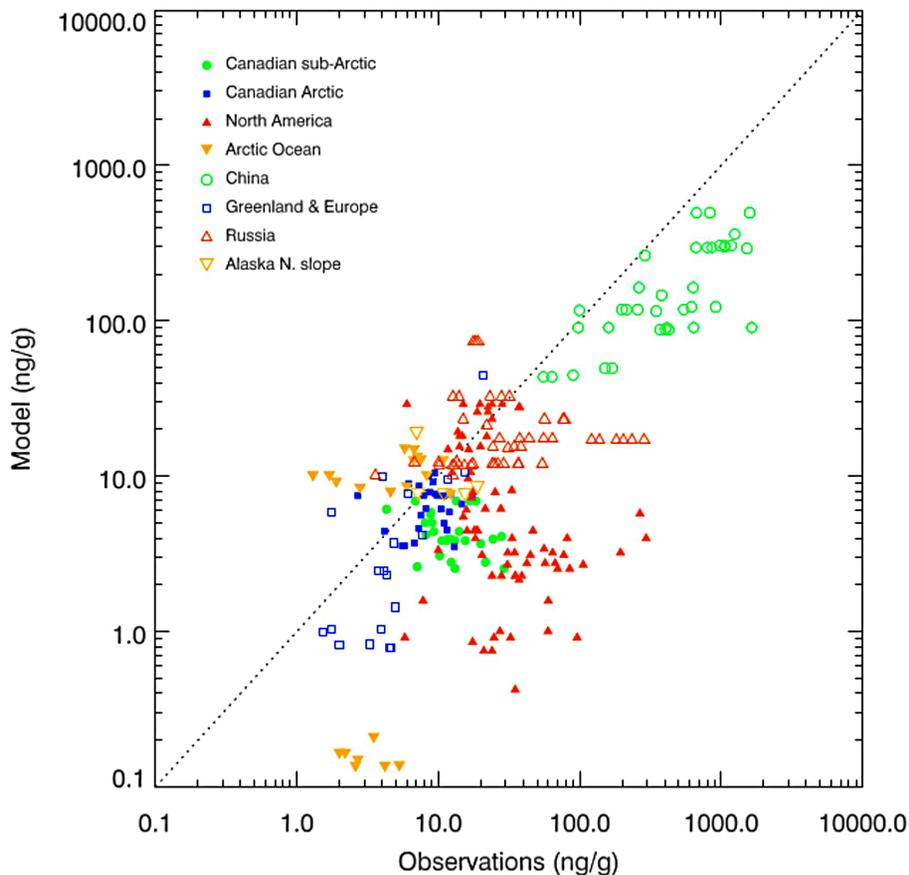


Figure 3. Comparison of measured and modelled BC in snow in log-scale scatter plot. The measurements are vertically averaged over depth for the top 25 cm and compared to simulated results over the same grid-box and month.

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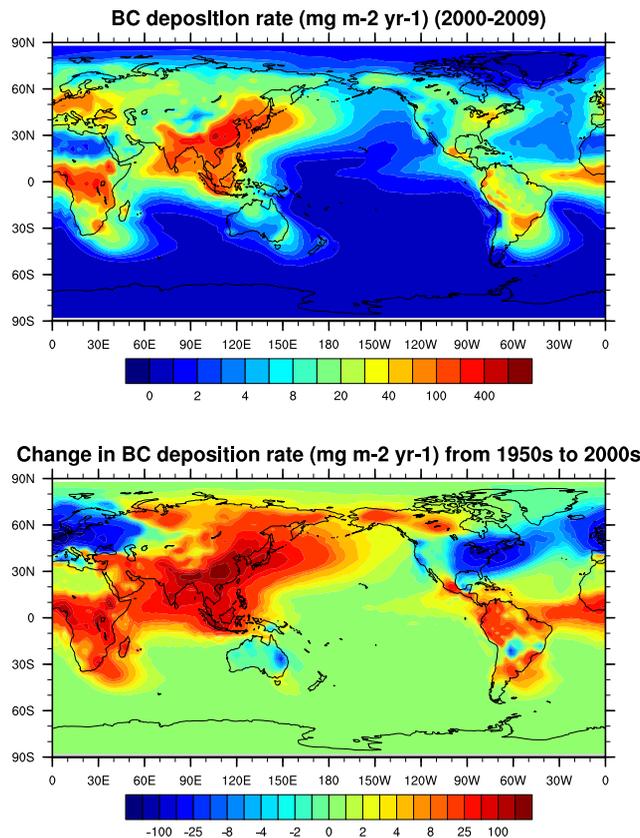


Figure 4. Annual mean BC deposition flux ($\text{mg m}^{-2} \text{yr}^{-1}$) during 2000–2009 (top) and annual mean BC deposition flux change 1950s to 2000s (bottom).

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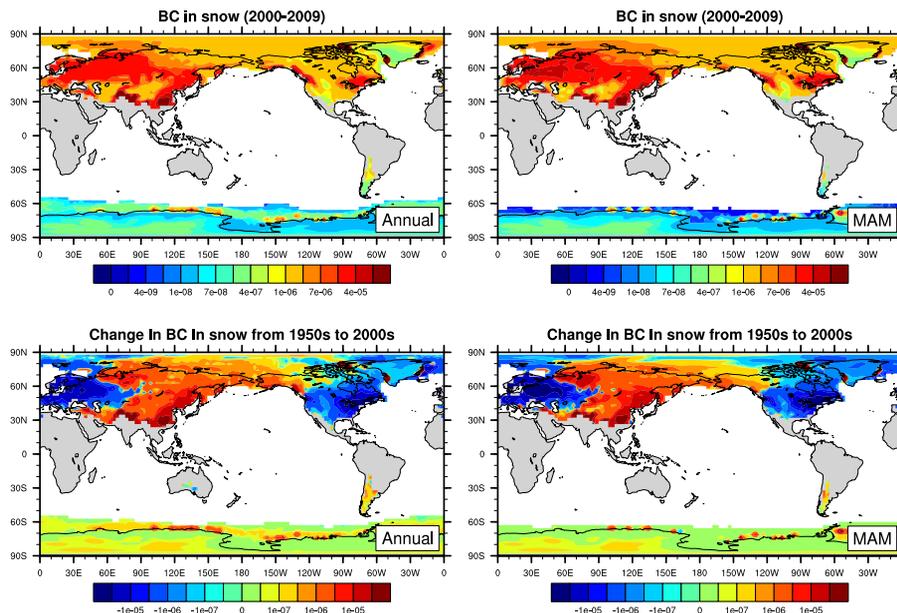


Figure 5. Annual mean (top left) and spring mean (top right) BC concentration in snow (kg m^{-3}) during 2000–2009, annual mean (bottom left) and spring mean (bottom right) change in BC concentration in snow (kg m^{-3}) from 1950s to 2000s.

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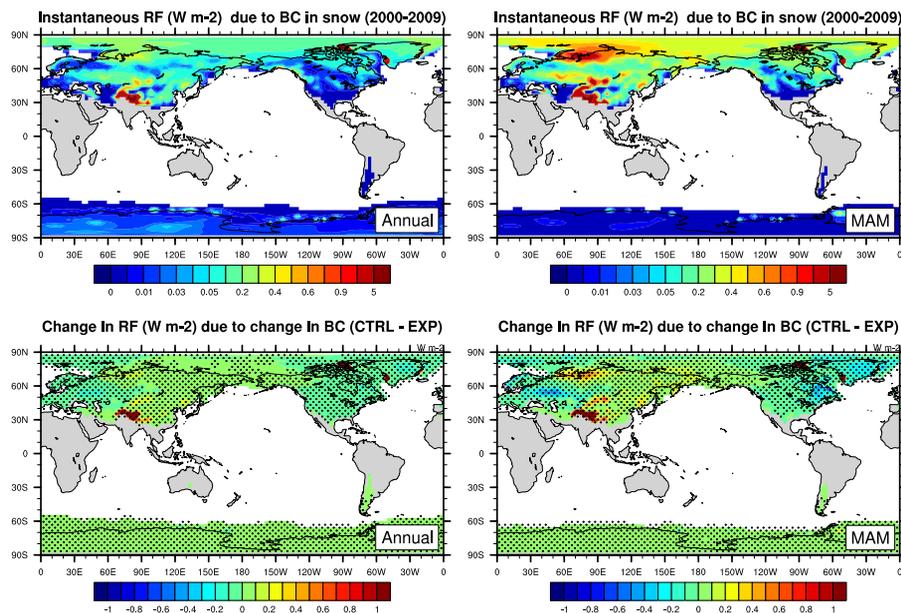


Figure 6. Annual mean (top left) and spring mean (top right) all-sky instantaneous radiative forcing of BC in snow (W m^{-2}) during 2000–2009, annual mean (bottom left) and spring mean (bottom right) radiative forcing changes (W m^{-2}) due to BC in snow from 1950s to 2000s. Hatching indicates regions where the results are statistically significant ($p = 0.05$).

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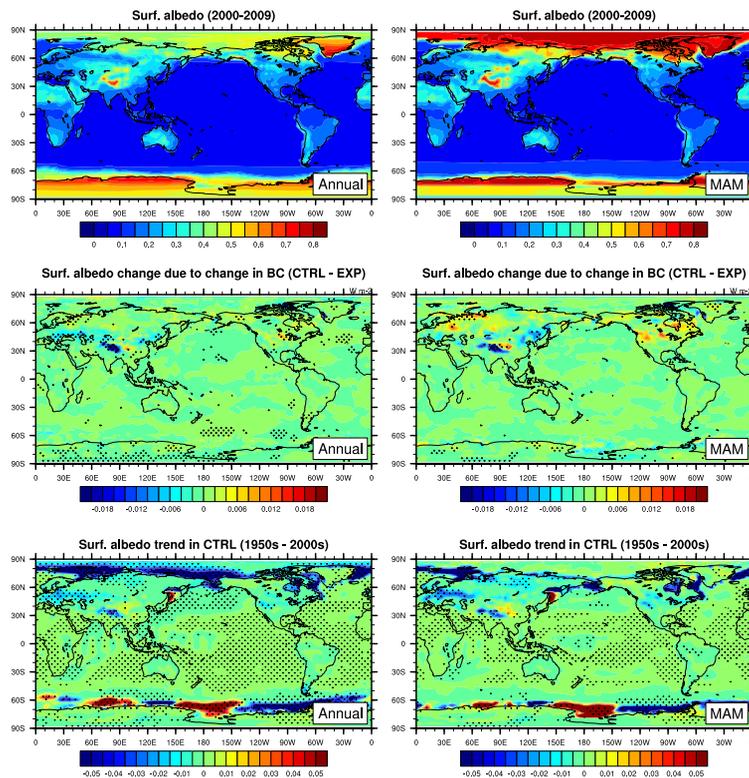


Figure 7. Annual mean (top left) and spring mean (top right) surface albedo during 2000–2009, annual mean (middle left) and spring mean (middle right) surface albedo changes due to BC in snow from 1950s to 2000s and annual mean (bottom left) and spring mean (bottom right) surface albedo changes due to all forcing from 1950s to 2000s. Hatching indicates regions where the results are statistically significant ($p = 0.05$).

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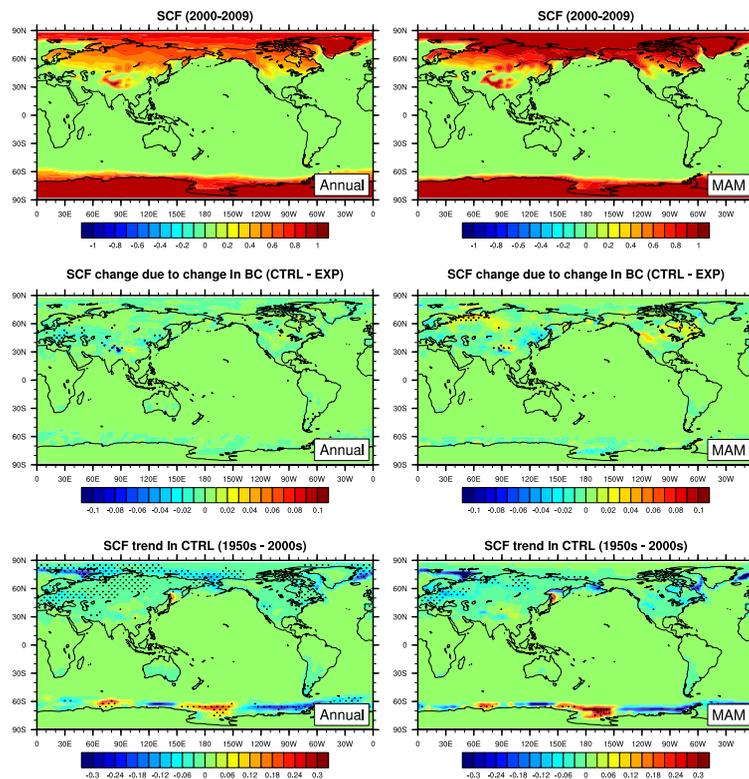


Figure 8. Annual mean (top left) and spring mean (top right) snow cover fraction during 2000–2009, annual mean (middle left) and spring mean (middle right) snow cover fraction changes due to BC in snow from 1950s to 2000s and annual mean (bottom left) and spring mean (bottom right) snow cover fraction changes due to all forcing from 1950s to 2000s. Hatching indicates regions where the results are statistically significant ($p = 0.05$).

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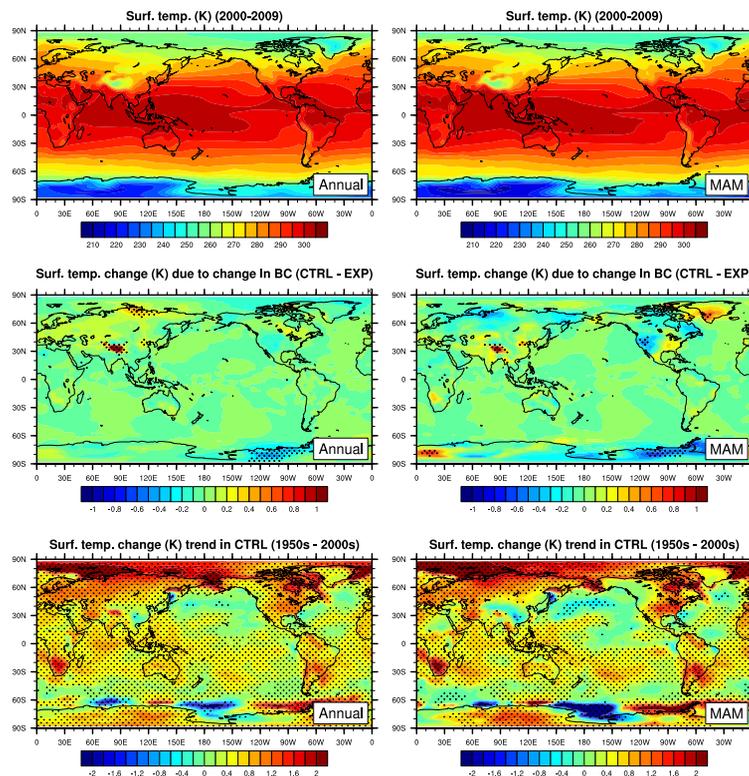


Figure 9. Annual mean (top left) and spring mean (top right) near-surface air temperature during 2000–2009, annual mean (middle left) and spring mean (middle right) near-surface air temperature changes due to BC in snow from 1950s to 2000s and annual mean (bottom left) and spring mean (bottom right) near-surface air temperature changes due to all forcing from 1950s to 2000s. Hatching indicates regions where the results are statistically significant ($p = 0.05$).