A sensitivity study on modeling black carbon in snow and its radiative forcing over the Arctic and Northern China

Yun Qian¹, Hailong Wang¹, Rudong Zhang²,¹, Mark G Flanner³ and Philip J Rasch¹

¹Atmospheric Sciences and Global Change Division, Pacific Northwest National Laboratory, Richland, WA, USA
²Key Laboratory for Semi-Arid Climate Change of the Ministry of Education, College of Atmospheric Sciences, Lanzhou University, Lanzhou 730000, Gansu, People’s Republic of China
³University of Michigan, Ann Arbor, MI, USA

E-mail: yun.qian@pnnl.gov

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Abstract
Black carbon in snow (BCS) simulated in the Community Atmosphere Model (CAM5) is evaluated against measurements over Northern China and the Arctic, and its sensitivity to atmospheric deposition and two parameters that affect post-depositional enrichment is explored. Improvements in atmospheric BC transport and deposition significantly reduce the biases (by a factor of two) in the estimation of BCS concentration over both Northern China and the Arctic. Further sensitivity simulations using the improved CAM5 indicate that the melt-water scavenging efficiency (MSE) parameter plays an important role in regulating BC concentrations in the Arctic through the post-depositional enrichment, which not only drastically changes the amplitude but also shifts the seasonal cycle of the BCS concentration and its radiative forcing in the Arctic. The impact of the snow aging scaling factor (SAF) on BCS shows more complex latitudinal and seasonal dependence, and overall impact of SAF is much smaller than that of MSE. The improvements of BC transport and deposition in CAM5 have a stronger influence on BCS than perturbations of the two snow model parameters in Northern China.

Keywords: black carbon, snow, radiative forcing, Arctic, Northern China

1. Introduction
Light absorbing aerosols (LAA, e.g., black carbon, dust) influence water and energy budgets of the atmosphere and snowpack in multiple ways (e.g., Twomey et al 1984, Albrecht 1989, Hansen et al 1997, Ramanathan et al 2001, Lau et al 2006, Qian et al 2009, 2011, Bond et al 2013). In addition to their effects associated with atmospheric heating by absorption of solar radiation and interactions with clouds, LAA deposited on snow can reduce snow reflectance (a.k.a., surface darkening), which is likely to accelerate the snow aging process and further reduces snow albedo and increases the speed of snowpack melt (Warren and Wiscombe 1980, Hansen and Nazarenko 2004, Xu et al 2009, Ming et al 2009).

Previous studies investigating the global climate impacts of BC in snow (BCS) indicate that BCS produces surface warming not only in the Arctic but also in lower latitudes across the northern hemisphere (Hansen and Nazarenko 2004, Jacobson 2004, Flanner et al 2007, 2009, Koch et al 2009, Shindell and Faluvegi 2009, Rypdal et al 2009, Huang et al 2011, Ye et al 2012, Zhang et al 2013). Modeling studies have suggested that this snow darkening mechanism has greater warming and snow-melting efficacy than any other anthropogenic agent (Hansen et al 2005, Flanner et al 2007, ...
Qian et al (2011, Skiles et al 2012). For a certain amount of atmospheric deposition, this large impact results from a series of positive feedback mechanisms: (1) as melt commences some of the LAA are not washed away with melt-water and accumulate on the surface or inside of snowpack (e.g., Conway et al 1996, Flanner et al 2007, Doherty et al 2010), (2) additional warming of snow with reduced albedo increases snow grain sizes, which further lowers the albedo (e.g., Warren and Wiscombe 1980, Hadley and Kirchstetter 2012), (3) with sufficient melt more of the darker underlying surface is exposed, leading to the well-known ‘snow albedo feedback’ (e.g., Warren and Wiscombe 1980, Hansen and Nazarenko 2004; Flanner et al 2007; Brandt et al 2011, Hadley and Kirchstetter 2012). The magnitude of positive feedback through LAA accumulation depends on the scavenging efficiency of LAA by snow-melt water. Doherty et al (2013) analyzed field measurements of the vertical distribution of BC and other LAA in snow in the Arctic during the melt season and found a significant melt-induced amplification of BCS concentration, increasing concentrations of BC by up to a factor of five. Xu et al (2012) also revealed a post-depositional enrichment of BC in snowpack by measuring the redistribution of BC along monthly snow-pits on a Tien Shan glacier in Northern China. Their results indicate that large uncertainties in estimating the effect of BC on snow and glacier melting likely exist in current climate models.

To quantify the magnitude of these positive feedbacks and their impact on snow and ice, it is critical to explore the sensitivity and uncertainty associated with the LAA deposition and snow-melting process in current earth system models. In this study, we first evaluate the simulated BCS against the measurements collected from multiple field campaigns over the Arctic (Doherty et al 2010) and Northern China (Wang et al 2013b). Then we conduct a series of sensitivity experiments to examine the impact of the snow aging factor and melt-water scavenging efficiency (MSE) on the snow-melting and radiative forcing of BC, and also compare the uncertainty resulting from the BC deposition with that related to the treatment of snow aging and melt-water scavenging for BC.

2. Model, experiment design and observation

We use the Community Atmosphere Model version 5 (CAM5; Neale et al 2010), which is the atmosphere component of the Community Earth System Model version 1 (CESM1, http://www.cesm.ucar.edu). Aerosol evolution in CAM5 is controlled by a combination of emissions, transport, aerosol microphysics, and dry and wet removal. Aerosol and cloud microphysics and their interactions in the model are described and evaluated by Gettelman et al (2008, 2010), Ghan et al (2012), and Liu et al (2012). Wang et al (2013a) have recently made improvements to the representation of aerosol convective transport and wet removal in CAM5, which substantially improves the aerosol simulation in remote regions such as the upper troposphere and high latitudes. This improvement is important for better characterizing atmospheric radiative heating and dirty snow effect induced by absorbing aerosols, especially in remote regions such as the Arctic.

The Snow, Ice, and Aerosol Radiative (SNICAR) model (Flanner and Zender 2006, Flanner et al 2007), which is coupled with the land component of CAM5/CESM to simulate snow albedo as well as the solar absorption within each snow layer, is the most detailed representation of snow metamorphism processes available for climate models. The radiative influence of BC on snow albedo simulated by SNICAR has been validated with recent laboratory and field measurements (Brandt et al 2011, Hadley and Kirchstetter 2012). The SNICAR model was originally developed for an older version of CAM (CAM3). Modifications have been made to the LAA treatment and radiative transfer codes for consistency with the new modal aerosol treatment in CAM5 (Liu et al 2012, Flanner et al 2012).

The MSE for hydrophilic BC is an important parameter for the prediction of the BCS amount in the SNICAR model (Flanner et al 2007). The MSE parameter relates the mass of BC (\(M_{BC}\)) in meltwater leaving a snow layer to the mass concentration of BC (\(C_{BC}\)) in the snow layer experiencing melt, i.e.:

\[
dM_{BC}/dt = Q_{melt} \times \text{MSE} \times C_{BC}
\]

where \(Q_{melt}\) is the flux of melt water leaving the snow layer (units of kg m\(^{-2}\) s\(^{-1}\)).

The default value for MSE is set to 0.2. However, due to a lack of \textit{in situ} measurement constraints, the uncertainty range for this parameter could be very large. Jiao et al (2014) applied an offline version of the Community Land Model (CLM) embedded with SNICAR along with BC deposition fields from different aerosol models, and found that while MSE impacts radiative forcing, it has little impact on model performance relative to observations from Doherty et al (2010) because most of their measurements occurred prior to snow melt. Here we conduct two ten-year CAM5 sensitivity simulations (SEh and SEl), in which the MSE is set to 2.0 and 0.02, respectively, to examine the impact of MSE on BCS and associated radiative forcing over both high and middle latitudes. Measurements suggest a very large range of uncertainty for MSE, varying with locations, temperature, size of BC-containing particles, snow aging status and so on. The range of MSE estimated by Doherty et al (2013) only represents the specific months and sites in the Arctic. Here we scale the default value (0.2) of MSE down and up by a factor of ten and use this arbitrary but reasonable range (0.02–2) to test model sensitivities to the parameter.

The snow aging scaling factor (SAF) is another uncertain parameter in the SNICAR model. SAF is a function of (1) a dry snow metamorphism parameterization implemented with lookup tables, (2) wet snow metamorphism, and (3) re-freezing of melt water. The series of equations used to describe these relationships are listed in the CLM4 Technical Note (http://www.cesm.ucar.edu/models/cesm1.0/clm/CLM4_Tech_Note.pdf). The SAF is simply used as a multiplier on the instantaneous rate provided by this set of equations. It has a default value of 1.0 and is used to scale the
simulated rate of snow aging, which depends on snow temperature, temperature gradient, density, and refreezing of melt water (e.g., Flanner and Zender 2006, Oleson et al. 2010). To characterize model response to ranges in this parameter, we conduct another set of sensitivity simulations, SAh and SAi, in which the SAF is set to 0.2 and 5.0 (i.e., 1/5 and 5 times of the default value, respectively), respectively. The two sets of sensitivity simulations are performed using the version of CAM5 with improved remote aerosol distributions (Wang et al. 2013a). The base case (IMPRV), with default values for both MSE and SAF, is conducted for comparison to the sensitivity simulations. Another simulation using the standard CAM5 configuration (CAM5std, same as IMPRV but without the improved remote aerosol distributions by Wang et al. 2013a) is also used to demonstrate the impact of the aerosol process improvements. Table 1 summarizes the model configuration of all simulations and the relevant parameters used in each simulation.

The CAM5 simulations were conducted at 1.9 degree (latitude) and 2.5 degree (longitude) horizontal grids, with 30 vertical layers. A standard set of present-day (year-2000) climatological conditions, including meteorology, aerosol and trace gases, was used to initiate model simulations and run for 11 years, of which the first year was treated as model spin up time and the final ten years were analyzed. Year-2000 monthly merged Hadley-NOAA/OI sea surface temperatures and sea ice concentrations were prescribed. Emissions of all aerosol species (including BC) were from year-2000 monthly mean inventories (Lamarque et al. 2010), which are also repeated used for 11 years.

The observations of BCS were made in certain months of a specific year (Doherty et al. 2010, Wang et al. 2013b). When making the observation-model comparison of BCS, we did match the months but had to use model climatology (ten-year mean), since the same emission data were used for all years. Spatially, measurements at points falling into the same model grid cells were averaged first and then compared with model climatological mean in the corresponding grids. The same method was also used in some previous studies (e.g., Goldenson et al. 2012, Lee et al. 2013, Jiao et al. 2014). Observations used for our model evaluation are estimates of the true mass of BC in snow samples using the wavelength-dependence of the measured absorption from the integrating sandwich/integrating sphere spectrophotometer (Grenfell et al. 2011). The spectrally-resolved total light absorption is divided into BC and non-BC fractions based on the absorption Angstrom exponent of the material on the filter, and by assigning different absorption Angstrom exponents to BC and non-BC LAA. BC is assumed to have a mid-visible (550 nm) mass absorption efficiency (MAE) of 7.5 m² g⁻¹ in the model, based on a review of BC properties (Bond and Bergstrom 2006), whereas measured BC masses are an equivalent mass assuming a 550 nm MAE of 6.3 m² g⁻¹ in Northern China (Wang et al. 2013b) and 6.0 m² g⁻¹ in the Arctic (Doherty et al. 2010). To facilitate a consistent comparison between the measured and modeled BCS concentrations, we divide the measured BCS provided in Wang et al. (2013b) by a factor of 1.19 and those provided in Doherty et al. (2010) by a factor of 1.25 to obtain the BC concentration that would be given by their optical measurements if the MAE of BC was, in fact, 7.5 m² g⁻¹. Evaluating the BCS in both high and middle latitudes provides insight into the long-range aerosol transport in the model and snow-darkening effect in different underlying surface and climate regimes.

### 3. Results

#### 3.1. Evaluation

We first evaluate the simulated BCS concentration against the measurements collected from multiple field campaigns over the Arctic (Doherty et al. 2010) and Northern China (Wang et al. 2013b). Figure 1(a) compares the simulated (IMPRV) and observed BCS concentration as a function of latitude. The observed BCS concentration ranges between 1 and 200 ng g⁻¹ in the Arctic and around 50 to 2000 ng g⁻¹ in Northern China. The model simulates reasonably well the magnitude of BCS concentration between the middle latitudes (Northern China) and high latitudes (Arctic) is also captured in the IMPRV.

The BCS concentration is significantly improved in the IMPRV compared to the standard CAM5 (CAM5std). The mean and median values of BC in snow are overestimated in Northern China but underestimated in the Arctic by a factor of about two in the CAM5std compared to the observations and IMPRV (figure 1(b)). Previous studies (e.g., Law and Stohl 2007) have shown that BC in the Arctic mostly originates from lower latitudes. The comparison here suggests that BC is deposited too efficiently in the standard CAM5 in the lower latitudes before reaching the high latitudes. With the improvement in aerosol transport and wet removal in the atmosphere (Wang et al. 2013a), the BCS is also much better simulated (IMPRV versus CAM5std in figure 1), providing us with more confidence in this version of CAM5 to estimate the radiative forcing induced by BC in snow.

Figure 2(a) shows the spatial distribution of CAM5 (IMPRV) simulated mean BC concentration in April in the top snow layer over land. There is a clear meridional gradient of BCS concentration decreasing with latitude from 50°N to the Arctic, as well as zonal variation. The meridional maximum BCS at 50°N is largely determined by the BC.
deposition from the atmosphere (Wang et al. 2013a). The maximum BCS concentrations (larger than 500 ng g$^{-1}$) are located at 50°N–60°N zonal area of Europe, which could be determined by many factors such as the distance to major emission sources, deposition rate and snow amount. Maximum BCS concentration larger than 200 ng g$^{-1}$ can also be seen over isolated areas in Northern China. The maximum BCS concentration in North America is around 200–500 ng g$^{-1}$, also located near 50°N. The BCS concentration in the Arctic region (north to 66.5°N, indicated by the black circle) is mostly less than 100 ng g$^{-1}$. The corresponding surface radiative forcing induced by BC in the snow column is shown in figure 2(b). Instantaneous radiative forcing is calculated by differencing surface net solar fluxes in parallel radiative transfer calculations that include and exclude BC in snow, identical to the approach used by Flanner et al. (2007). Although global mean forcing from BCS is small (Hansen and Nazarenko 2004), the regional forcing is strong enough to

**Figure 1.** Evaluation of CAM5 simulated black carbon (BC) concentration (in ng g$^{-1}$) in the top snow layer against observations in the Arctic (Doherty et al. 2010) and Northern China (Wang et al. 2013b). The top snow layer ranges in thickness from 1 to 3 cm. Configuration of the two CAM5 simulations (CAM5std and IMPRV) is summarized in table 1. Panel (a) shows the comparisons at different latitudes. The box and whisker plot in panel (b) shows the minimum and maximum value with the bar, the 25th and 75th percentiles with the box, the 50th percentile (i.e., median) by the bar within the box, and the mean value with the dot. Panels (c) and (d) are scatter plots for observations versus CAM5std and IMPRV, respectively.
change the local heat budget and snow melt. Consistent with
the spatial distribution of BCS, the larger forcing is located in
a region at 50°N–60°N and 30°E–90°E and a few isolated
regions in the Northern China, with maximum values
approximately in the range of 5–10 W m$^{-2}$. The BCS-induced
forcing is smaller than 2 W m$^{-2}$ in most other regions. Note
that differences between the spatial pattern of BCS in the top-
snow layer (figure 2(a)) and forcing due to BC in the column
(figure 2(b)) are likely due to other factors (e.g., BC in sub-
surface snow layers, the amount of other light-absorbing
aerosols, snow grain size, surface insolation affected by
clouds and vegetation in the environment, etc) that affect the
forcing calculation as well.

3.2. MSE for BC

To quantify the sensitivity of BCS to MSE, we compare the
two simulations in which the MSE is set to 0.1 and 10 times
of its default value, respectively (see table 1). Figure 3(a)
shows the difference of the monthly (April) mean BC
concentration in the top snow layer between a high and low
scavenging efficiency (SEh and SEl respectively). BCS differs
significantly over the majority of areas covered by snow,
with the maximum decrease (SEh—SEl) over 500–1000
ng g$^{-1}$ in northern Europe and western Russia. The BCS
change due to MSE is relatively small at high latitudes
including Greenland, largely because the melting is not very
active in April as the temperature is still generally below the
melting point. Consistent with the changes of BCS concen-
tration, radiative forcing also decreases over most areas
where snow is present, with a maximum decrease of 2–3
W m$^{-2}$ over western Russia (figure 3(c)). The decrease of
surface radiative forcing due to the increase of MSE is rela-
tively small in the Arctic because of the small BCS change
and the low solar radiation flux.

Figure 4(a) shows the monthly mean BCS concentration
in the Arctic averaged when snow is present. Note that BCS
forcing is averaged spatially and temporally only over the
snow-covered portion of the gridcell, i.e., it represents the
mean forcing experienced by the snow surface within the
gridcell. In the control simulation IMPRV, the BCS builds up
during the cold and early warm season with a peak value of
about 40 ng g$^{-1}$ in July, and becomes smaller than 10 ng g$^{-1}$
after the main melting season, showing a strong seasonal
cycle. The seasonal variation of BCS in IMPRV, especially
the dramatic jump from spring to early summer, is primarily
driven by the seasonal cycle of BC deposition from the
atmosphere (see supplemental figure S2). The post-deposi-
tional enrichment effect (determined by the MSE parameter)
becomes important when melting starts, as indicated by the
departure of the BCS concentration in the SEl and SEh away
from that in the IMPRV case. BCS in SEl (minimal scaven-
ging by melt-water or the maximal enrichment) has a similar
seasonal variation but the magnitude is almost doubled from
that in IMPRV. When the MSE is at its maximal value (SEh,
minimal enrichment), not only is the amplitude of the sea-
sonal cycle drastically reduced but also the peak is shifted
from July to April, indicating an important role of post-
depositional enrichment in regulating the seasonal cycle of
BC in Arctic snowpack.

The uncertain MSE parameter can significantly change
the vertical distribution of BCS and snow surface albedo, and
thus alter BCS radiative forcing. As shown in figure 4(b),
seasonal variation of the radiative forcing is similar to that of
BCS except that the forcing is also determined by the
downwelling solar radiation at the surface and other factors. A
reasonably large MSE (in SEh) can reduce the BCS-induced
forcing by a factor of 2–3 in the warm season (e.g., May–July
in the Arctic, which is a strong indication of the potential feedback via snow melting (i.e., higher MSE →
more efficiently BC being scavenged → smaller BCS-
concentration in snow → larger snow albedo → smaller BCS-
induced forcing).
Figures 4(c) and (d) show the monthly mean BCS in top snow layer and surface radiative forcing induced by BCS (in snow column) in Northern China averaged when snow is present. The BCS peaks five months earlier and is about 7 times larger in magnitude than in the Arctic. Although the top-layer BCS concentration is much larger in Northern China, the peak radiative forcing induced by BCS (in snow column) is around 1.4 W m$^{-2}$, which is smaller than in the Arctic, partly determined by the seasonal variation of insolation and the thickness of snow. The MSE parameter also affects the BCS and the induced radiative forcing in Northern China, but its impact in the peak season is not as strong as in the Arctic likely because the lifetime of snowpack in lower latitudes is short.

Results from the figure 4 can be used to compare the impact of model improvements (IMPRV versus CAM5std) on BCS concentration and associated radiative forcing with the sensitivity of these quantities to the MSE parameter used in the SNICAR model (IMPRV versus SEh and SEl). In the Arctic, the 30% underestimation of annual mean BCS concentration and the induced forcing in CAM5std is much smaller compared to the sensitivity induced by the perturbation of MSE. At the lower latitudes (e.g., Northern China), the overestimation of annual mean top-layer BCS by 50% and the induced forcing by 30% is much larger than the sensitivity to

Figure 3. The CAM5 simulated difference of ten-year monthly mean (April) (a), (b) BC concentration in top snow layer (in ng g$^{-1}$) and (c), (d) forcing induced by BC in snow column (in W m$^{-2}$) averaged when snow is present between simulations SEh and SEl for (a), (c) and between SAh and SAI for (b), (d). Stippled areas denote the statistically significant difference with significance level at 0.1 using Student’s t test. Description of the simulations is in table 1.
the MSE parameter. The model improvement of atmospheric BC deposition in IMPRV reduces the high bias in BCS but does not shift the phase of BCS seasonal cycle (IMPRV versus CAM5std in figure 4(c)). However, changes in the MSE parameter influence BCS most significantly during the melting season (March–June) because of the key role of MSE in the post-depositional enrichment.

3.3. SAF

The impact of SAF (a simple multiplier applied to the snow aging rate) on BCS and induced forcing shows more complex spatial variation and seasonal dependence. Figure 3(b) shows the difference of the monthly mean BCS in April between SAh and SAi, which apply the maximum and minimum value of the SAF parameter, respectively. When SAF is changed from low to high, BCS has an overall increase at high latitudes (north to 50°N) except for a few isolated regions in Siberia and Canada. Increase of SAF reduces snow albedo and accelerates the snow melting over most of the high latitudes (e.g., Arctic) in the melting season, leading to an increase in the BCS concentration because of the accumulation of BC in snow and shrinkage of snowpack in both thickness and cover fraction (i.e., decrease of snow water equivalent). At lower latitudes, figure 3(b) shows an overall decrease for BCS but changes are not statistically significant (at 0.1 significance level) when SAF is switched from low to high.
high, probably because the snowpack has partly disappeared in April in the SAh over many areas at lower latitudes. The meridional dependence of SAF-induced impact on the surface radiative forcing is more distinct than on the BCS concentration, i.e. larger radiative forcing in higher latitudes but smaller in the lower latitudes in SAh than in SAI (figure 3(d)). Note that, unlike the BC scavenging efficiency parameter, the SAF does not directly affect BCS but through feedback from changes in snow properties, which means stronger non-linearity in the relationship between top-layer BCS and forcing in the snow column for the SAF sensitivity tests. This likely explains the differences in the spatial patterns of BCS and forcing between the two groups of tests (figures 3(a), (c) versus figures 3(b), (d)).

Figure 5 show the monthly mean top-layer BCS and the radiative forcing induced by BC in snow column in the Arctic and Northern China averaged when snow is present. The seasonal variation of BCS is similar in the three simulations (i.e. SAh, SAI and IMPRV) over the Arctic with a peak in July, but the magnitude of BCS in SAh is 20–30% larger than in SAI and IMPRV during the melting season. The most dramatic increase from late spring to summer occurs in SAh, suggesting a positive impact of snow aging on the post-depositional enrichment of BC in snowpack by accelerating the melting. The monthly mean difference between SAI and IMPRV for both BCS and induced radiative forcing is very small in the Arctic. In Northern China (figure 5(c)), the difference of BCS induced by SAF is very small. The uncertainty in terms of the magnitude associated with the
representation of BC transport and deposition processes (IMPRV versus CAM5std) is much larger than that associated with the SAF parameter (IMPRV versus SAh and SAi).

4. Summary

In this study we evaluate the CAM5 simulated BC in snow (BCS) against measurements over the Arctic and Northern China, and explore the sensitivity to two important model parameters. Observed BCS is 10 times higher at middle latitudes (Northern China) than at high latitudes. The BCS concentration is overestimated (underestimated) by approximately a factor of two in Northern China (Arctic) in the standard CAM5 (CAM5std), compared to the observations, suggesting that BC is removed from the atmosphere too efficiently in the CAM5std at mid-latitudes during the transport from mid-latitude sources to the high latitudes. BCS concentrations match observation more closely in the simulations using improved treatments for aerosol wet removal and convective transport. Results from the improved model treatments suggest that the maximum BCS in spring occurs in the major continents along the 50°N–60°N latitudinal band and a few isolated areas in Northern China. The maximum BCS mass concentration is larger than 1000 ng g$^{-1}$, and the induced local radiative forcing could be over 10 W m$^{-2}$ in April.

Model results indicate that the meltwater scavenging efficiency (controlled by parameter MSE) plays an important role in regulating BC concentrations in the Arctic through the post-depositional enrichment, which not only drastically changes the amplitude (by a factor of 2–7) but also shifts the seasonal cycle of the BCS concentration for the parameter range explored in this study, having significant impacts on radiative forcing. The seasonal variation of BCS in Northern China (peak in February) is different from that in the Arctic (peak in July) with the peak value 7 times higher in Northern China; however, the peak BCS radiative forcing is of a similar magnitude. The MSE parameter can even alter the BCS seasonal cycle.

The impact of the SAF on BCS shows more complex spatial and seasonal dependence in part because, unlike MSE, the SAF directly affect snow properties which then have feedback on BCS. When SAF is switched from low to high, BCS has an overall statistically significant increase at high latitudes except for a few isolated regions in Siberia and Canada, and has an overall decrease at lower latitudes but changes are not statistically significant. The increase of BCS is most dramatic during late spring to early summer in the Arctic in the simulation with a higher SAF, suggesting a positive impact of snow aging on the post-depositional enrichment of BC in snowpack. SAF has a stronger influence on the magnitude of BCS seasonal cycle over the Arctic than over Northern China. The model improvement in the representation of BC transport and deposition has a stronger influence on BCS than variations in MSE and SAF in Northern China. However, changes in the MSE parameter influence BCS most significantly during the melting season (March–June) because of the key role of MSE in the post-depositional enrichment.

Although the model simulated BCS shows good agreement with the limited set of observations, it is worth noting a number of limitations in the model simulations of BCS, including but not limited to the bias of snowfall prediction in the coarse-resolution climate model, large uncertainties in BC emission, and uncertainties associated with the treatment of BC deposition processes in climate models. While the sensitivities associated with the two snow model parameters in estimating the BCS effect are characterized, no optimal values of those uncertain parameters are identified. More comprehensive studies exploring these limitations, and including in situ observational and laboratory studies, with a focus on in-snow processes such as snow aging and melt-water scavenging of BC, will be helpful to reduce uncertainty in quantifying the climatic and hydrological effects of BC in snow/ice.

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