



1 Impacts of absorbing aerosol deposition on snowpack and hydrologic
2 cycle in the Rocky Mountain region based on variable-resolution
3 CESM (VR-CESM) simulations

4 Chenglai Wu^{1,2}, Xiaohong Liu^{1,*}, Zhaohui Lin^{2,3}, Stefan R. Rahimi-Esfarjani¹, and
5 Zheng Lu¹

6 ¹*Department of Atmospheric Science, University of Wyoming, Laramie, Wyoming,*
7 *USA*

8 ²*International Center for Climate and Environment Sciences, Institute of Atmospheric*
9 *Physics, Chinese Academy of Sciences, Beijing, China*

10 ³*University of Chinese Academy of Sciences, Beijing, China*

11

12 **Corresponding to:*

13 Xiaohong Liu
14 Department of Atmospheric Science
15 University of Wyoming
16 Dept. 3038, 1000 East University Avenue
17 Laramie, WY 82071
18 Email: xliu6@uwyo.edu.

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21 Abstract

22 Deposition of light-absorbing aerosols (LAAs) such as black carbon (BC) and dust
 23 onto snow cover has been suggested to reduce the snow albedo, and modulate the
 24 snowpack and consequent hydrologic cycle. In this study we use the
 25 variable-resolution Community Earth System Model (VR-CESM) with a regionally
 26 refined high-resolution (0.125°) grid to quantify the impacts of LAAs in snow in the
 27 Rocky Mountain region during the period of 1981-2005. We first evaluate the model
 28 simulation of LAA concentrations both near the surface and in snow, and then
 29 investigate the snowpack and runoff changes induced by LAAs in snow. The model
 30 simulates similar magnitudes of near-surface atmospheric dust concentrations as
 31 observations. Although the model underestimates near-surface atmospheric BC
 32 concentrations, the simulated BC-in-snow concentrations are overall comparable to
 33 observations. Regional mean surface radiative effect (SRE) due to LAAs in snow
 34 reaches up to $0.6\text{--}1.7\text{ W m}^{-2}$ in spring, and dust contributes to about 21-43% of total
 35 SRE. Due to positive snow-albedo feedbacks induced by the LAAs' SRE, snow water
 36 equivalent reduces by 2-50 mm and snow cover fraction by 5-20% in the two regions
 37 around the mountains (Eastern Snake River Plain and Southwestern Wyoming),
 38 corresponding to an increase of surface air temperature by $0.9\text{--}1.1^\circ\text{C}$. During the snow
 39 melting period, LAAs accelerate the hydrologic cycle with monthly runoff increases
 40 of $0.15\text{--}1.00\text{ mm day}^{-1}$ in April-May and reductions of $0.04\text{--}0.18\text{ mm day}^{-1}$ in
 41 June-July in the mountainous regions. Of all the mountainous regions, Southern



42 Rockies experience the largest reduction of total runoff by 15% during the later stage
43 of snow melt (i.e., June and July). Our results highlight the potentially important role
44 of LAA interactions with snowpack and subsequent impacts on the hydrologic cycles
45 across the Rocky Mountains.

46

47 1. Introduction

48 Water resources are essential to human society and economic development as
49 well as ecosystems in the western United States. Most of primary water resources in
50 the inland western U.S. come from the Rocky Mountains' snowpack (Serreze et al.,
51 1999). Therefore, to develop the water resource management strategy, it is necessary
52 to have on information of snow accumulation and snowmelt timing. Climate change
53 is an important factor influencing the snowpack in the Rocky Mountain region, as has
54 been shown in many previous studies (e.g., Abatzoglou, 2011; Pederson et al., 2011;
55 Rhoades et al., 2017). Another important factor is the light-absorbing aerosols (LAAs,
56 e.g., black carbon (BC), organic carbon (OC), and dust) in snow (e.g., Flanner et al.,
57 2007; Painter et al., 2007; Qian et al., 2015; Yasunari et al., 2015). Previous studies
58 have shown that LAAs in snow can significantly reduce the surface albedo (often
59 known as snow darkening effect (SDE)), modify the surface energy budget and
60 snowmelt, and lead to the modification of hydrologic cycles (e.g., Warren and
61 Wiscombe, 1980; Hansen and Nazarenko, 2004; Flanner et al., 2007, 2009; Painter et
62 al., 2007, 2010; Qian et al., 2009, 2011; Yasunari et al., 2015). Moreover, the



63 LAAs-induced snow albedo reduction may initiate positive feedback processes, which
64 can amplify the reduction of snowpack (e.g., Flanner et al., 2009; Qian et al., 2009).

65 In past decades modeling studies have been undertaken to quantify the impacts
66 of SDE by LAAs (e.g., Flanner et al., 2007; Qian et al., 2009; Oaida et al., 2015;
67 Yasunari et al., 2015). Generally the models they developed have the ability to
68 simulate the temporal evolution of snow albedo under the influence of LAAs in snow.
69 These studies have enhanced our understanding of the spatial and temporal variations
70 of climate forcings and responses due to LAAs in snow from regional scales (e.g.,
71 Qian et al., 2009; Oaida et al., 2015) to global scales (e.g., Flanner et al., 2007;
72 Yasunari et al., 2015). For example, the impacts of LAA in snow are stronger in
73 regions with considerable snow cover and sufficient LAAs deposition (e.g., Arctic,
74 Northeast China, Tibetan Plateau, and western U.S.) ~~than in other regions~~, and they
75 are largest during the snowmelt period due to the positive snow-albedo feedback.
76 However, as also mentioned in these studies, reliable quantification of impacts of
77 LAAs in snow is hindered by the model deficiencies in simulating the snowpack and
78 aerosol cycles, with additional uncertainties induced by the parameterization of
79 snow-aerosol-radiation interactions.

80 In particular, previous studies ^{have} ~~used~~ the coarse-resolution global climate
81 models (GCMs) or high-resolution regional climate models (RCMs) to quantify the
82 impacts of LAAs in snow. However, there are weaknesses in ^{both} ~~either~~ coarse-resolution
83 GCMs ~~and~~ RCMs. Both snowfall and snow accumulation depend on ~~the~~ temperature

and in



A variable-resolution version of the Community Earth System Model (VR-CESM) has been developed (Zarzycki et al., 2014a, b). With a refined high resolution, VR-CESM has shown significant improvements of the Atlantic tropical



105 storms (Zarzycki and Jablonowski, 2014) and South America orographic precipitation
106 (Zarzycki et al., 2015). The model has also been used in the regional climate
107 simulations ~~in~~ ^{of the} western U.S., and results show ~~that~~ ^{the} VR-CESM is capable of
108 reproducing the spatial patterns and the seasonal evolution of temperature,
109 precipitation, and snowpack in ~~the~~ ^{the} Sierra Nevada (Huang et al., 2016; Rhoades et al.,
110 2016) and ~~in~~ Rocky Mountains (Wu et al., 2017). In particular, VR-CESM simulates
111 reasonably the magnitude of snow water equivalent, the timing of snow water
112 equivalent peaks, and the duration of snow cover in the Rocky Mountains, ~~by~~ ^{as shown in}
113 ~~comparisons~~ ^{comparisons} against ~~the~~ Snow Telemetry (SNOTEL) and MODIS (Moderate
114 Resolution Imaging Spectroradiometer) snow cover observations (Wu et al., 2017).

115 Following the evaluation study of Wu et al. (2017), here we use VR-CESM to
116 investigate the impacts of LAAs in snow (BC and dust) on the snowpack and
117 hydrologic cycles over the Rocky Mountains. By comparing the two VR-CESM
118 simulations with and without ~~considering the impacts of~~ LAAs in snow, we examine
119 ~~the changes in~~ ^{impacts on} surface radiative transfer, temperature, snowpack, and runoff induced
120 by LAAs in snow. To our knowledge, it is the first time that VR-CESM is applied for
121 the study of LAAs in snow. Our results will demonstrate that VR-CESM is skillful for
122 this kind of research.

123 The remainder of the paper is organized as follows. Section 2 introduces the
124 model and experimental design. Section 3 describes the observation data used for
125 validation of model simulations of aerosol fields in the surface air and in snow.



126 Section 4 presents the evaluation of aerosols fields, followed by their surface radiative
127 effect (SRE), as well as the change of surface temperature, snowpack, and runoff
128 induced by LAAs in snow. Discussion and conclusions are given in section 5.

129 2. Model and experimental design

130 The model used in this study is VR-CESM, a version of CESM (version 1.2.0)
131 with the variable-resolution capability (Zarzycki et al., 2014a, b). CESM is a
132 state-of-the-art Earth system modeling framework that allows for ~~for~~ investigation of
133 a diverse set of Earth system interactions across multiple time and space scales
134 (Hurrell et al., 2013). CESM uses the Community Atmosphere Model version 5
135 (CAM5) for the atmospheric component (Neale et al., 2010). The variable-resolution
136 capability is implemented into the Spectral Element (SE) dynamic core of CAM5.
137 The SE dynamic core uses a continuous Galerkin spectral finite-element method
138 designed for fully unstructured quadrilateral meshes, and has demonstrated
139 near-optimal (close to linear) parallel scalability on tens of thousands of cores (Dennis
140 et al., 2012). This enables the model to run efficiently on decadal to multi-decadal
141 time scales. For the land component, CESM uses the Community Land Model version
142 4 (CLM4). CLM4 can be run at the same horizontal resolutions as CAM5 and thus **can**
143 also benefit from the variable-resolution capability of CAM5.

144 CESM also includes advanced physics for CAM5 (Neale et al., 2010) and
145 CLM4 (Oleson et al., 2010). The CAM5 physics suite consists of shallow convection
146 (Park and Bretherton, 2009), deep convection (Zhang and McFarlane, 1995; Richter



147 and Rasch, 2008), cloud microphysics (Morrison and Gettelman 2008) and
148 macrophysics (Park et al. 2014), radiation (Iacono et al. 2008), and aerosols (Liu et al.,
149 2012). For aerosols, a modal aerosol module (MAM) is adopted to represent the
150 internal and external mixing of aerosol components such as BC, OC, sulfate,
151 ammonium, sea salt, and mineral dust (Liu et al., 2012). CLM4 physics includes a
152 suite of parameterizations for land-atmosphere exchange of water, energy and
153 chemical compounds. In particular, CLM4 explicitly represents the snowpack (snow
154 accumulation and melt) by a snow model and its coupling with the SNow, Ice and
155 Aerosol Radiation (SNICAR) model for snow-aerosol-climate interactions (Flanner et
156 al., 2007). SNICAR incorporates a two-stream radiative transfer solution of Toon et
157 al. (1989) to calculate the snow albedo and the vertical absorption profile from solar
158 zenith angle, albedo of the substrate underlying snow, mass concentrations of
159 atmospheric-deposited aerosols (BC and dust), and ice effective grain size (r_e). r_e is
160 simulated with a snow aging routine (Oleson et al., 2010). SNICAR is compatible
161 with the new modal aerosol module of CAM5 (Liu et al., 2012) in the treatment of
162 aerosol deposition (Flanner et al., 2012). As our knowledge of OC optical properties
163 is limited, the impact of absorbing OC on snow albedo is not included in the standard
164 CLM4 and thus not considered in this study.

165 For the high-resolution modeling, we have designed a variable-resolution grid
166 that transits from global quasi-uniform 1° resolution to a refined 0.125° resolution in
167 the Rocky Mountains (Figure 1a). The variable-resolution grid is the same as that



168 used in Wu et al. (2017), and is generated by the open-source software package called
169 SquadGen (Ullrich, 2014). A topographical dataset for this variable-resolution grid is
170 also generated accordingly by the National Center for Atmospheric Research (NCAR)
171 global model topography generation software called NCAR_Topo (v1.0) (Lauritzen et
172 al., 2015) as described in Wu et al. (2017). As shown in Figure 1b, the high-resolution
173 grids resolve well the variations of terrain in the Rocky Mountains. Note that the
174 standard CESM using the coarse 1° grids cannot resolve the fine-scale variations of
175 terrain in the Rocky Mountains (see Figure 2 of Wu et al. (2017)). In Wu et al. (2017),
176 we have shown that VR-CESM performs well in the simulation of regional climate
177 patterns, including spatial distributions and seasonal evolution of temperature,
178 precipitation, and snowpack in the Rocky Mountain region. In this study, we further
179 apply VR-CESM to simulate the SDE of LAAs and its impacts on snowpack and
180 hydrologic cycles in the Rocky Mountains.

181 VR-CESM is run in the coupled land-atmosphere mode with prescribed
182 observed monthly $1^\circ \times 1^\circ$ sea surface temperature and sea ice coverage (Hurrell et al.,
183 2008), following the Atmospheric Model Intercomparison Project (AMIP) protocols
184 (Gates, 1992). The simulation period is from 1979 to 2005, and the results for the last
185 25 years (1981-2005) are used for the analysis shown below. Historical greenhouse
186 gas concentrations, and anthropogenic aerosol and precursor gas emissions are
187 prescribed from the datasets of Lamarque et al. (2010). In particular, the BC
188 emissions consist of various sources, including domestic, energy, transportation,



189 waste, shipping, and wildfire (forest and grass fires) emissions. We note that the
190 horizontal resolution for BC emission used in this study is $1.9\times 2.5^\circ$. The relatively
191 coarse resolution of BC emission may partly explain the model's bias in the
192 simulation of BC concentrations near the surface and in snow across regions where
193 local BC sources can contribute significantly to the observed BC concentrations, as
194 will be discussed in section 4.

195 For dust aerosol, the emission flux is calculated interactively in the model at
196 each time step by a dust emission scheme (Oleson et al., 2010). The dust emission
197 flux is calculated from the friction velocity, threshold friction velocity, atmospheric
198 density, clay content in the soil, areal fraction of exposed bare soil, and source
199 erodibility (Oleson et al., 2010; Wu et al., 2016). Due to the large uncertainty in
200 modeled dust emission, the dust emission scheme also adopts a posterior tuning factor
201 (T) to simulate the reasonable dust emission amount. With the increase of model
202 resolution, VR-CESM produces much higher dust concentrations compared to the
203 observations (section 3) in North America if T used in the standard CESM with
204 quasi-uniform 1° resolution is used. Therefore, for VR-CESM, T is reduced by a
205 factor of 2.6 to produce the similar magnitudes of near-surface dust concentrations as
206 the observations, as will be shown in section 4.1. Note that such a reduction of T is
207 only applied in North America, since other continents have a resolution of
208 quasi-uniform 1° , the same as in the standard CESM.



209 In addition to the control experiment with the impacts of LAAs (BC and dust)
210 in snow included (CTL), we conduct a sensitivity experiment that turns off the impact
211 of LAAs in snow (NoSDE). Through the comparison of these two simulations (CTL
212 and NoSDE), the impacts of SDE by LAAs on the snowpack and hydrologic cycles
213 can be identified. To facilitate the analysis of SDE, we also calculate the surface
214 radiative effect (SRE) by BC (dust) in snow in the control experiment from the
215 difference of absorbed radiation with all aerosols (i.e., the standard radiation call) and
216 with all aerosols except BC (dust) as Flanner et al. (2007) (a diagnostic radiation call).

217 To quantify the impacts of LAAs in snow, we mainly focus on the five regions.
218 Three of these regions are in the high mountains: Northern Rockies, Greater
219 Yellowstone region, and Southern Rockies. The elevation is higher in Greater
220 Yellowstone region and Southern Rockies (>2250 m) than in Northern Rockies (>750
221 m). The other two regions are over the plains near the mountains: Snake River Basin
222 and Southwestern Wyoming. These two regions are selected because they are close to
223 the source regions of BC and dust and also have considerable snow cover (>50%) in
224 winter. These five regions are shown in Figure 1c.

225 3. Observations

226 We will use various observations to validate the model simulation of aerosol
227 (BC and dust) concentrations near the surface and in snow.

228 First, we use the observations of near-surface ^{atmospheric}BC and dust concentrations from
229 the Interagency Monitoring of PROtected Visual Environments (IMPROVE) network



230 (Malm et al., 1994). Observed mass concentrations of Elemental Carbon (EC) are
 231 used for the comparison with model simulation of BC concentrations. Although EC
 232 can be somewhat different from BC (Andreae and Gelencser, 2006), EC
 233 concentrations have been widely used for the validation of BC concentrations in
 234 previous studies (e.g., Koch et al., 2009; Liu et al., 2012). For dust, simulated dust
 235 concentration accounts for dust particles with diameters below 10 μm . To compare,
 236 observed mass concentrations of fine soil (FS, with the diameter $<2.5 \mu\text{m}$) and coarse
 237 mass (CM, with the diameter between 2.5 μm and 10 μm) from IMPROVE are
 238 combined, following the approach of Kavouras et al. (2007) and Wells et al. (2007).
 239 In reality, in addition to dust, CM may also contain other aerosols such as sulfate,
 240 nitrate, organic and elemental carbon, and sea salt. However, according to the study of
 241 Malm et al. (2007), who analyzed the speciation of coarse particles collected at nine
 242 selected rural IMPROVE stations in 2004, the contributions of dust to CM are above
 243 70% (74-90%) ^{at} the three stations in inland western U.S. In their study, lower
 244 contributions of dust to CM (34% and 65%) were found in the two stations near the
 245 coast. We caution that these two stations were $<150 \text{ km}$ away from the metropolitan
 246 regions indicating that urban emissions may also contribute to CM there. Additional
 247 contributions may result from sea salt or sodium nitrate resulting from reactions of
 248 nitric acid with sea salt, as mentioned in their study (Malm et al., 2007). Therefore, to
 249 minimize the contributions of other aerosols to CM, we do not use the stations in or
 250 near the metropolitan regions or near the coast for the validation of dust concentration.



251 Nonetheless, we acknowledge that there may be small contributions from other
252 aerosols to CM and the estimated dust concentration by summing FS and CM may
253 represent an upper limit of dust concentrations (with the diameter $<10\ \mu\text{m}$) from the
254 observations. Note that the observation period of IMPROVE varies with the stations;
255 ~~some stations started collecting data in the~~ 1980s, and some ~~from~~ more recently (2000s). To
256 derive a climatological dataset for model comparisons, we only select the stations
257 with more than 5 years of dust observations. ~~Totally~~ ^{In total} there are 80 and 94 stations for
258 BC and dust observations, respectively, in the western U.S. (Figure 2).

259 Second, we ~~collected~~ ^{use} field measurements of BC mass concentrations in snow
260 (C_{BC}) from previously published studies. Although field observations of C_{BC} in snow
261 extended back to 1980s, they were made mostly in high-latitudes, ~~Alps Mountains,~~ ^{in the}
262 Cascade Mountains, eastern Canada, and West Texas/New Mexico (see Qian et al.
263 (2015) and references therein). Recently, Doherty et al. (2014) made valuable
264 measurements of the vertical profiles of LAAs in seasonal snow from January to
265 ~~March of 2013.~~ ^{in the western U.S.} They used an Integrating Sphere integrating SandWich (ISSW)
266 Spectrophotometer to estimate the BC mixing ratios in snow over 67 sites in North
267 America (including 17 sites in the Rocky Mountain region). Observed C_{BC} by
268 Doherty et al. (2014) was recorded on a single day. Doherty et al. (2016) further
269 provided the temporal variations of C_{BC} at four stations, three ~~in~~ ⁱⁿ Idaho (January to
270 March of 2014) and one ~~in~~ ⁱⁿ Utah (February to March of 2013 and 2014). Doherty et al.
271 (2016) also calibrated the ISSW measurements using an incandescence technique (the



272 Single Particle Soot Photometer, SP2) in a subset of the observations, which was
273 supposed to capture C_{BC} more accurately, and derived a ratio of C_{BC} by ISSW to C_{BC}
274 by SP2 based on their linear relationship for the estimation of real C_{BC} . This
275 calibration is applied to the dataset of Doherty et al. (2014) in our study, and thus the
276 observations of Doherty et al. (2014) and Doherty et al. (2016) are comparable ~~and~~
277 ~~used here~~ used ere

278 In addition, Skiles and Painter (2016b) made daily measurements of
279 BC-in-snow with ~~an~~ SP2 in the Senator Beck Basin Study Area (SBBSA) in the San
280 Juan Mountains during a period of two ~~month~~ months (late March to middle May) in 2013.
281 The locations and sample dates as well as the measurements for these stations are
282 given in Table 1. If measurements of C_{BC} on multiple days were made, the means and
283 standard deviations of C_{BC} are given. As our simulation period does not encompass
284 the years 2013 and 2014, we will compare the monthly mean results for 25 years
285 (1981-2005) by focusing more on the general magnitudes and spatial distributions of
286 C_{BC} . At each station, the mean results for the month (or months) when the
287 observations were made, as well as the maximum and minimum C_{BC} , are derived
288 from the 25-year simulation and compared to the observations.

289 ~~There, for dust, there~~ There are few observations of dust mass mixing ratio in snow
290 (C_{dust}) in the Rocky Mountain region. To our best knowledge, the only published
291 observations were conducted in the Senator Beck Basin Study Area (SBBSA) in the
292 San Juan Mountains (in Southern Rockies) ~~with~~ ; these are



(Painter et al., 2007, 2012; Skiles and Painter, 2016a, 2016b). Snow samples at a depth of 30 cm were collected at irregular time intervals for each dust event from March to May. We will compare simulated C_{dust} for the whole snow columns with the observations (for 0-30 cm depth), but acknowledge that C_{dust} may vary in the snow underneath 0-30 cm snow layer. Another consideration is that observed C_{dust} contains all the dust particles while simulated C_{dust} only accounts for the dust particles with diameters smaller than 10 μm . According to the observations of Reynolds et al. (2016), the concentration of total suspended particles (TSP) in the atmosphere is mainly contributed from particles with diameters larger than 10 μm . This will affect the model comparison with the observations, which will be discussed in section 4.

4. Results

4.1 Spatial patterns of near-surface aerosol concentrations

Before we examine the impacts of aerosol deposition onto snow, we will first evaluate the aerosol simulations by the model. Figure 2 shows the spatial patterns of cold season (winter and spring) mean emission fluxes and near-surface concentrations of BC and dust in the western U.S. from the VR-CESM simulation. The IMPROVE stations are also denoted by cycles with larger cycle sizes indicating higher observed near-surface BC/dust concentrations. In the model, the BC emission flux is prescribed and is largest in the Pacific Coast and southern Arizona. BC emission fluxes are relatively large in central–northern Colorado and Northwestern Utah, where large metropolises are located. Corresponding to the patterns of BC emission flux,



314 simulated near-surface BC concentrations ($>100 \text{ ng m}^{-3}$) are also higher in these
315 regions. A band with relatively high near-surface BC concentrations around 50-100
316 ng m^{-3} is also found in southern Idaho, to the west of the Greater Yellowstone region
317 and to the south of Northern Rockies, indicating the transportation of BC around the
318 mountains. Near-surface BC concentrations decrease at higher elevations. The spatial
319 patterns simulated by the model are generally consistent with observations, e.g.,
320 higher BC concentrations in the source regions and lower in the mountains.

321 Dust sources are located in the dry regions with exposed bare soils, such as the
322 southwestern U.S. (southern California, western Arizona, and southern New Mexico),
323 the northern Mexico, the Great Basin, and the Colorado Plateau. Dust emissions are
324 also found in the Great Plains, although it is much weaker. In the Great Plains
325 agricultural activities can disturb the soil, making it vulnerable to wind erosion
326 (Ginoux et al., 2012). Simulated cold season mean dust concentrations are higher
327 ($10\text{-}500 \text{ } \mu\text{g m}^{-3}$) in the source regions, but decrease dramatically ($0.1\text{-}5 \text{ } \mu\text{g m}^{-3}$) to the
328 mountains. Compared to the observations, the model reproduces the spatial patterns of
329 near-surface dust concentrations with higher concentrations in the southwestern part
330 of US. However, the model tends to overestimate the dust concentrations in Utah,
331 indicating that dust emission may be overestimated there.

332 Comparisons of modeled and observed near-surface BC/dust concentrations at
333 the IMPROVE stations are further shown in Figure 3. Overall, the model captures the
334 magnitudes of observed near-surface BC and dust concentrations with the differences



335 between observations and simulations within a factor of 5 for most of the stations.

336 The model simulates similar magnitudes of observed near-surface BC concentrations

337 at the stations along the West Coast and in the Southwestern U.S. However, the model

338 tends to underestimate observed near-surface BC concentrations in Utah–Nevada

339 regions, the Rocky Mountains, and the Great Plains, where the stations are located ~~the~~

340 ~~the~~ downwind of source regions (Figure 2b). In particular, observed near-surface BC

341 concentrations are underestimated by about a factor of two in the Rocky Mountains.

342 The underestimation of near-surface BC concentrations in these regions may suggest

343 that transport of BC in our simulations is too weak. This deficiency may also be

344 ascribed to ~~the~~ local BC sources (e.g., Doherty et al., 2014) not resolved by the

345 prescribed BC emission in the model (e.g., at $1.9\times 2.5^\circ$ resolution). For dust, although

346 the model overestimates near-surface dust concentrations for most of the stations near

347 the dust sources (southwestern U.S., Utah, and Nevada), the model simulates well the

348 magnitude of near-surface dust concentrations in the Rocky Mountains. This may also

349 be associated with underestimated transport in the model, ~~as indicated in the bias of~~ **consistent with the low bias in**

350 near-surface BC concentrations in ~~the~~ downwind regions.

351 Note that although only the BC and dust emission fluxes over the western U.S.

352 are shown in Figure 2, long-range transport of these aerosols from other regions (e.g.,

353 Asia and Africa) can also contribute to BC (e.g., Zhang et al., 2015) and dust (Wells

354 et al., 2007) concentrations in the western U.S. **In addition,** ~~Meanwhile,~~ there are ~~also~~ substantial

355 variations of aerosol emission in the western U.S. As mentioned in section 2, although



we adopt VR-CESM with a refined high resolution (0.125°) in the Rocky Mountains,
we use a coarse resolution gridded emission dataset (i.e., $1.9^\circ \times 2.5^\circ$) for BC. For dust,
the small-scale variations of dust ~~emission~~ ^{emissions} can be represented in the model as it is
calculated online in the model. However, dust emission depends on many variables
such as near-surface winds, soil moisture, vegetation cover, and soil texture (Oleson
et al., 2010; Wu et al., 2016), which may ~~be~~ ^{themselves} biased. In particular, in Utah and Nevada,
simulated near-surface dust concentrations are about 2-3 times ~~larger than~~ ^{larger than}
~~observations~~ ^{ed}, indicating ~~the~~ ^{emissions} significant overestimation of dust ~~emission~~ in the region.
Despite the aforementioned biases, the model does reasonably well in the simulation
of spatial variations of near-surface BC and dust concentrations in western U.S.

4.2 Aerosol-in-snow concentrations

Figure 4 shows the spatial distributions of BC and dust mass mixing ratios in
snow in winter and spring from VR-CESM simulations. BC-in-snow mass mixing
ratio in the Rocky Mountains ranges from $2\text{--}50\text{ ng g}^{-1}$, which is consistent with a
previous study (Qian et al., 2009). The dust-in-snow mass mixing ratio ($0.1\text{--}50\text{ }\mu\text{g g}^{-1}$)
is about 2-3 orders of magnitude higher than that of BC-in-snow. The spatial pattern
of BC-in-snow mixing ratios is consistent with that of near-surface atmospheric BC
concentration, which features higher values in northern Utah and southern Idaho and
lower values in the higher mountains (Figure 2b). Dust-in-snow mixing ratios are
higher in Utah and downwind regions (western Colorado and southern Idaho), which
is consistent with the distribution of near-surface atmospheric dust concentrations.



377 Dust-in-snow mixing ratio is also higher in the northern Great Plains, where dust
378 emission is also evident (Figure 2c). In addition, BC and dust mixing ratios are larger
379 ($10\text{--}100\text{ ng g}^{-1}$ and $2\text{--}50\text{ }\mu\text{g g}^{-1}$, respectively) in the Southern Rockies than in Northern
380 Rockies and Greater Yellowstone region. BC and dust mass mixing ratios are smaller
381 ~~in~~^{the} Greater Yellowstone region with ranges from $10\text{--}50\text{ ng g}^{-1}$ and from $0.2\text{--}2\text{ }\mu\text{g g}^{-1}$,
382 respectively, and are smallest in the Northern Rockies with the values below 20 ng g^{-1}
383 and below $2\text{ }\mu\text{g g}^{-1}$, respectively. BC and dust mixing ratios in snow are larger in
384 spring than in winter in most of the Rocky Mountain region. This indicates that BC
385 and dust accumulate within the snow column during the snow accumulation and the
386 early portion of the snow melting season.

387 The comparison of BC mass mixing ratios in the snow column at the 17 sites
388 from VR-CESM simulations and observations is shown in Figure 5. As observations
389 only sampled the snow in one day or tens of days, simulated mean and standard
390 deviations of BC-in-snow mass mixing ratios in the same month are given for the
391 comparison. The months of observations occurred between January and May.
392 Observed BC-in-snow mass mixing ratios range from 5.6 ng g^{-1} to 33.6 ng g^{-1} at the
393 17 sites. The model reproduces reasonably the magnitude of observed BC-in-snow
394 mass mixing ratios at most of the stations. The exception is that at the two stations
395 (sites #15 and #16) where the model significantly underestimates observed
396 BC-in-snow mass mixing ratios by a factor of more than 5. Site #15 is close to the
397 sites #13 and #14 and they are all located on the southwestern flanks of Northern



Rockiest (Figure 4). Observed BC mass mixing ratios at site #15 (14.9 ng g^{-1}) is higher than those at sites #14 (13.3 ng g^{-1}) and #13 (9.8 ng/g), which indicates the northward transport of BC (Doherty et al., 2016). Simulated BC-in-snow mass mixing ratios are around 8.8 ng g^{-1} and 10.9 ng g^{-1} at sites #13 and #14, respectively, which are comparable to the observations. However, the simulated value is only 2.5 ng g^{-1} at site #15. This indicates that the model may lack the ability to simulate the transport and/or deposition of BC in this region. Site 16 is located in northeastern Utah, where snow depths are smaller compared to other mountains regions. When snow is melted completely, BC-in-snow mixing ratio will be zero, but the model will average the simulation results at different time steps to derive the mean result. In the observations, only BC-in-snow samples are accounted for. The different sample strategy may partly explain the difference between the simulations and observations. Another reason may be that observations show a large interannual variability of BC-in-snow mass mixing ratios at site #16 and the observation period was not long enough for the derivation of a climatological mean as in the simulation.

Note that although near-surface BC concentrations in the atmosphere are underestimated in the Rocky Mountains in the model (section 4.1), BC mass mixing ratios in the snow are not overall underestimated. In our study, we calibrate the observational data of Doherty et al. (2014) by using the correction factor based on the comparison of ISSW and SP2, assuming that SP2 can more accurately measure the mass mixing ratio of BC compared to ISSW (Doherty et al., 2016). However,



419 although SP2 can provide a direct measurement of BC, SP2 may underestimate the
420 real amount of BC-in-snow mass when BC is attached to larger particles (e.g., dust
421 and sea salt) or aggregates to large sizes in snow due to the size range (e.g., $\sim 0.08\text{--}0.7$
422 μm) limitations in SP2 (Qian et al., 2015). Because of this, the real amount of
423 BC-in-snow mass may be higher than that measured by SP2. Therefore, VR-CESM
424 may also underestimate the BC-in-snow mass mixing ratios, although the exact degree
425 of underestimation is unknown. Another reason for the inconsistency of BC mass
426 mixing ratios in snow and near-surface BC concentrations in the atmosphere may be
427 related to the snow aging/melting and BC-in-snow accumulation and flushing-out,
428 which are associated with large uncertainties (Flanner et al., 2007; Qian et al., 2014).

429 For dust in snow, the simulated mean dust mass mixing ratio in snow in spring
430 is $19.6 \mu\text{g g}^{-1}$ in the San Juan Mountains. The simulated standard deviation, minimum,
431 and maximum of dust mass mixing ratios for 1981-2005 are $22.4 \mu\text{g g}^{-1}$, $5.3 \mu\text{g g}^{-1}$,
432 and $118.7 \mu\text{g g}^{-1}$, respectively. This value is one to two orders of magnitude smaller
433 ~~than the observed mixing ratios from~~ compared to the observations of Skiles and Painter et al. (2016a), which showed that,
434 at the end of the snow season, the total dust-in-snow mass mixing ratios range from
435 0.2 to 4.8 mg g^{-1} . Much smaller dust-in-snow mass mixing ratios in the simulations
436 may be ascribed to the fact that the model only accounts for dust particles with
437 diameters smaller than $10 \mu\text{m}$, while the observations include all the sizes of dust
438 ~~particles in the snow. Actually, an observation made~~ **Observations** by Reynolds et al. (2016) in the
439 Utah-Colorado region showed that concentrations of total suspended particles (TSP)



440 in the atmosphere are ~~mainly contributed~~ ^{mostly} from larger particles with diameters larger
441 than 10 μm . Therefore, the model may underestimate the impacts of dust deposition
442 into snow, ~~and the dust impacts~~ ^{. Dust calculated} in this study, which will be discussed below, ~~can be~~ ^{should}
443 regarded as those from the dust particles with diameters smaller than 10 μm .

444 4.3 Surface radiative effect (SRE) by aerosol-in-snow

445 Figure 6 shows the spatial distribution of instantaneous surface radiative effect
446 (SRE) due to BC- and dust-induced snow albedo change, respectively, in winter
447 (December-January-February) and spring (March-April-May). Due to the decrease of
448 surface albedo, surface net shortwave radiation is increased. The spatial patterns of
449 SRE are determined ~~collectively~~ ^{distribution} by both the amount of aerosol in snow and the
450 ~~distribution of snowpack~~ (snow depth and snow cover fraction). Finer-scale structures
451 of SRE in the Rocky Mountains and the adjacent regions are simulated by VR-CESM
452 with a higher horizontal resolution compared to previous simulations by
453 coarse-resolution GCMs (e.g., Flanner et al., 2009; Yasunari et al., 2015). The SRE is
454 generally above 0.2 W m^{-2} over the mountains especially in the Greater Yellowstone
455 region and Southern Rockies. SRE can reach similar magnitudes on the southern
456 periphery of Northern Rockies and west side of the Greater Yellowstone region,
457 where higher near-surface atmospheric BC/dust concentrations and BC/dust-in-snow
458 mass mixing ratios are simulated (Figures 2 and 4). SRE is stronger in spring than in
459 winter for both BC and dust, which is consistent with previous studies (Flanner et al.,
460 2009; Yasunari et al., 2015). This is because of the stronger solar insolation and larger



461 albedo reduction due to snow aging and aerosol accumulation within snow in spring.
462 ~~There are more dust emission and consequent~~ **Dust emissions, and consequently** dust transport and deposition **are higher** in spring
463 than in winter, which may also partly contribute to the larger dust-induced SRE in
464 spring than in winter. ~~For different aerosols,~~ BC-induced SRE is somewhat larger than
465 dust-induced SRE in both winter and spring. BC-induced SRE is mostly below 1 W
466 m⁻² in winter, but reaches up to 2-5 W m⁻² in spring. Dust-induced SRE is mostly
467 below 0.5 W m⁻² in winter and increases to 1-5 W m⁻² in spring.

468 Compared to ~~those in~~ the Greater Yellowstone region and Southern Rockies,
469 SRE in the Northern Rockies is much smaller (mostly below 0.05 and 0.5 W m⁻² in
470 winter and spring), because of smaller aerosol-in-snow **mixing ratios** ~~mass values~~ in this region
471 (Figure 4). Note that BC-induced SRE is still significant (mostly around 0.2-2 W m⁻²
472 and 2-5 W m⁻² in some local regions) in the northern Great Plains, eastern U.S.,
473 southern Canada, and eastern Canada. This was also shown in previous studies using
474 coarse-resolution GCMs (e.g., Flanner et al., 2009; Yasunari et al., 2015). In addition,
475 the model also simulates non-negligible dust-induced SRE (mostly around 0.05-0.2 W
476 m⁻² and up to 0.2-0.5 W m⁻² in some local regions) near the dust sources in southern
477 Canada and the northern Great Plains.

478 Figure 7 shows the monthly variations of SRE induced by BC and dust SDE in
479 the five regions (Northern Rockies, Greater Yellowstone region, Southern Rockies,
480 Eastern Snake River Plain, and Southwestern Wyoming). Table 2 gives the regional
481 averaged winter and spring SRE in these five regions. Consistent with the spatial



distributions shown in Figure 6, aerosol-induced SRE averaged in Northern Rockies is
about half to one-fourth of ~~those in the~~ ^{that in the} Greater Yellowstone region and Southern Rockies.
Compared to that in winter, SRE is much larger in spring, which is a result of aerosol
accumulation in snow and relatively strong solar insolation. ~~The peaks of~~ ^{Maxima in the monthly} SRE occur
in April-May in the three mountainous regions (Northern Rockies, Greater
Yellowstone region, and Southern Rockies), which corresponds to the onset of
snowmelt, ~~after the peaks of~~ ^{consistent with observed peaks in} snow water equivalent (early to middle April; see figure
11 of Wu et al. (2017)). In the Eastern Snake River Basin and Southwestern Wyoming,
~~the peaks of~~ ^{maxima in monthly} SRE occur in March, which is different from the three mountainous
regions because the snowmelt period begins earlier (in February to March) in these
two regions (section 4.4). Regional mean total SRE in spring induced by BC and dust
can reach up to 1.58-1.7 W m⁻² with peaks around 2.0 W m⁻² in Greater Yellowstone
region and Southern Rockies. In ~~the~~ ^{the} Eastern Snake River Plain and Southwestern
Wyoming, ~~regional~~ ^{regions} mean total SRE in winter and spring is around 0.4-0.6 W m⁻² and
0.7-0.8 W m⁻², respectively. For the contribution of different aerosols, BC-induced
springtime SRE is larger than dust-induced SRE in the five regions. Despite this,
dust-induced springtime SRE can still contribute to about 20-30% of total springtime
SRE in the northern part of Rocky Mountains (Northern Rockies, Greater
Yellowstone region and Eastern Snake River Plain). In the southern part of Rocky
Mountains (Southwestern Wyoming and Southern Rockies), dust-induced springtime
SRE contributes more significantly (about 30-40%) to total springtime SRE.



503 4.4 Impacts of aerosol SDE on the surface temperature and snowpack

504 Figure 8 shows surface air temperature, snow water equivalent, and snow
505 cover fraction changes due to the aerosol SDE in winter and spring, respectively.
506 Snow water equivalent is defined as the amount of water contained within the
507 snowpack, measured ~~by~~ ^{by} kg m^{-2} which is equivalent to mm after dividing ~~the~~ ^{by} the density
508 of water (1000 kg m^{-3}). Snow cover fraction is defined as the fraction of surface area
509 covered by snow. These changes are derived from the difference between ~~two~~ ^{the}
510 simulations (CTL and NoSDE). The crosses in the figure denote the regions where
511 changes are statistically significant at the 0.1 level. Although SRE is largest over the
512 mountains, surface air temperature change is largest around the mountains, such as
513 over the Eastern Snake River Plain, Northern Utah, and Central-Southwestern
514 Wyoming, where surface air temperatures are increased by around $0.5\text{--}2^\circ\text{C}$ due to the
515 aerosol SDE. The large surface air temperature increase corresponds well to the
516 significant reductions of snow water equivalent (by 2–50 mm) and snow cover
517 fraction (by 5–20%) in these regions. This indicates a pronounced positive feedback
518 between snow albedo, radiation, and surface temperature around the mountains,
519 where snow water equivalent values are relatively lower and snow cover fractions are
520 smaller than those over the mountains. The positive feedback amplifies the surface
521 warming and snow melting, as was also found in a previous study using the Weather
522 Research and Forecasting (WRF) model (Qian et al., 2009). We note, however, both
523 snow water equivalent and snow cover fraction are larger over the mountains. For



524 example, winter and spring snow cover fraction is mostly above 70% on the high
525 mountains (see Figure 7 of Wu et al. (2017)). This suggests that snow on the high
526 mountains is less susceptible to the aerosol SDE. Another reason for the smaller
527 change of surface air temperature over the mountains is that snow water equivalent
528 and snow cover fraction are increased (especially in the Northern Rockies and Greater
529 Yellowstone region) due to the increase of snowfall in these regions (Figure 9f),
530 which cancels out the reduction of snow water equivalent resulting from aerosol SDE.
531 The increase of snowfall is due to enhanced water vapor transport from the Pacific
532 Ocean (Figure not shown), which is likely related to the large-scale circulation change
533 due to aerosol SDE. Note that increases of snow water equivalent and snow cover
534 fraction in the Northern Rockies and Greater Yellowstone region due to aerosol SDE
535 do not pass the significant test at 0.1 level because of the large interannual variability
536 in these regions.

537 Table 2 gives the winter and spring surface air temperature changes due to
538 LAAs in snow averaged over the five regions. Seasonal mean surface air temperature
539 change is around 0.9-1.1 °C in the Eastern Snake River Plain in winter and spring,
540 while this change is around 0.003-0.17 °C (winter) and around 0.3-0.5 °C (spring) in
541 the mountainous regions (Northern Rockies, Greater Yellowstone region, and
542 Southern Rockies). In Table 2, we also show the efficacy of snow albedo forcing,
543 which is defined as the ratio of surface air temperature change to SRE. The efficacy is
544 mostly around 0.1-0.5 in the three mountainous regions, but it is 1.3-2.2 in the Eastern



Snake River Plain and Southwestern Wyoming. This indicates that stronger snow albedo feedbacks exist in the latter two regions.

Figures 10-11 show the monthly evolution of regional mean surface air temperature, snow water equivalent, and snow cover fraction, and their changes due to aerosol SDE in the Eastern Snake River Plain and Southwestern Wyoming, respectively. Monthly variations of surface air temperature, snow water equivalent, and snow cover fraction are similar between the two regions: lowest surface air temperature and largest snow cover fraction in January, and highest snow water equivalent in February-March. Significant changes of these variables from the aerosol SDE occur in both regions. The largest surface air temperature increase is 1.5 °C in the Eastern Snake River Plain and 1.6 °C in Southwestern Wyoming, occurring in April and December, respectively. In the Eastern Snake River Plain (Southwestern Wyoming), aerosol SDE leads to the reduction of snow water equivalent by 6-28 mm (6-16 mm) and snow cover fraction by 5-15% (6-19%) from December to March, which corresponds to a fraction of 18-33% (20-37%) and 7-28% (8-28%), respectively, with respect to the NoSDE simulation results. In April (late snowmelt period) when snow water equivalent and snow cover fraction are both relatively small, the aerosol SDE is more significant, which reduces snow water equivalent values and snow cover fractions by about half.

4.5 Runoff change induced by aerosol SDE



565 In the model, the runoff includes the surface runoff and sub-surface runoff.
566 Runoff is mainly from ~~the~~ rainfall and snowmelt. The change of rainfall is shown in
567 Figures 9c-9d, and the snowmelt change is shown in Figure 12. Aerosol SDE
568 increases the snowmelt by 0.1-2 mm/day in the mountains during the snow
569 accumulation and early snowmelt period (in autumn, winter and spring). In the late
570 snowmelt period, aerosol SDE reduces the snowmelt due to ~~to~~ ^{there being} less snowpack available
571 for melting in the plains (in spring) and in the mountains (in summer). Note that
572 snowmelt is slightly reduced by the aerosol SDE in autumn in the Southern Rockies,
573 which is a result of less snowpack ^{available} ~~for~~ melting due to the reduced snowfall in this
574 region (Figure not shown).

575 Because of the change in rainfall and snowmelt due to aerosol SDE, surface
576 runoff changes too. Figure 13 shows the runoff change induced by aerosol SDE in
577 four seasons. In winter, runoff is barely modified by the aerosol SDE in the Rocky
578 Mountains, except in the Northern Rockies where runoff is increased by 0.1-2 mm
579 day⁻¹, associated with increased rainfall (Figure 9c) and increased snowmelt (Figure
580 12a). In spring, runoff changes the most compared to all of the other seasons, with the
581 runoff increased by up to 0.5-2 mm day⁻¹ in the mountainous regions. This is mainly
582 due to the increase of snowmelt resulting from surface warming (Figure 12b) as well
583 as due to more snow available for melt resulting from snowfall increase (Figure 9f).
584 The changes in runoff are statistically significant at 0.1 level in most of the
585 mountainous regions in spring. Absolute runoff increases are stronger in the Northern



586 Rockies and Greater Yellowstone regions than in the Southern Rockies in terms of the
587 area and magnitude, probably due to the smaller snow water equivalent in Southern
588 Rockies (Wu et al., 2017). As more snowmelt occurs in spring, less snowpack is
589 available for melt in summer and thus surface runoff is reduced by about 0.1-1 mm
590 day⁻¹. There is little runoff change in autumn, as there is less runoff generated from
591 rainfall and snowmelt than in other seasons. Overall, BC and dust residing in snow
592 accelerate the hydrologic cycles by increasing the runoff in spring and reducing the
593 runoff in summer. Surface warming also increases the ratio of rainfall to total
594 precipitation, which can accelerate the generation of runoff. Note that in some regions
595 of the plains, such as the central-eastern Montana, Southwestern Wyoming, ~~and~~ ^{the} Snake
596 River Basin, the snowmelt changes by 0.1-1 mm/day due to the aerosol SDE, but the
597 runoff changes little. This is because the water generated from snowmelt is mainly
598 stored in soil or transformed into evapotranspiration in these regions. Also note that
599 there are statistically significant increases of runoff in the southern Great Plains in
600 spring, but the change is small (around 0.05-0.1 mm/day; Figure 13b). This change is
601 a result of slight increases in both rainfall (Figure 9d) and snowmelt (Figure 12b).

602 Figure 14 shows the monthly evolution of runoff and its change due to the
603 aerosol SDE in the three mountainous regions (Northern Rockies, Greater
604 Yellowstone region, and Southern Rockies). In the three regions, runoff peaks in the
605 late spring and early summer (in May in Northern Rockies and Southern Rockies, and
606 in June in Greater Yellowstone region) when snow melting progresses after the peak



607 of snow water equivalent in early-to-middle April (Wu et al., 2017). This indicates the
608 significant contribution of snowmelt to runoff. Overall, runoff changes are larger in
609 the Northern Rockies and Greater Yellowstone region than in Southern Rockies,
610 which is consistent with the spatial distribution of runoff changes shown in Figure 13.
611 Runoff is significantly increased in spring and decreased in June and July, indicating
612 the acceleration of the local hydrologic cycle by aerosol SDE. In the Northern
613 Rockies, runoff is also increased from October to March but in much smaller
614 magnitudes (below 0.2 mm day^{-1}) compared to April and May. In April (May), runoff
615 is increased by 0.39 (0.56), 0.22 (1.00), and 0.17 (0.15) mm day^{-1} in the the Northern
616 Rockies, Greater Yellowstone region, and Southern Rockies, respectively. This
617 increase contributes to 26% (13%), 42% (27%), and 29% (7%) of the runoff from the
618 NoSDE simulation in April (May) for the three regions, respectively. The reduction of
619 runoff in June is relatively small (0.06 and 0.11 mm day^{-1} , respectively) in the
620 Northern Rockies and Greater Yellowstone, only accounting for 2% of runoff from
621 the NoSDE simulation. However, it reaches up to 0.18 mm day^{-1} in the the Southern
622 Rockies, which accounts for 15% of runoff. In addition, due to the reduction of snow
623 available for melting in later ~~month~~ ^{months (e.g.,} July), the runoff is further reduced. With
624 respect to the ~~relative~~ ^{versus} smaller runoff in July ~~than~~ in previous months, aerosol SDE is
625 more significant, which can reduce the runoff by 0.04 (8%), 0.17 (23%), and 0.06 mm
626 day^{-1} (16%) in the three regions, respectively. Note that due to increase of



precipitation, the annual mean runoff is increased by 0.12 (12%), 0.09 (10%), and
 0.01 mm day⁻¹ (2%) in these three regions, respectively.

5. Conclusions

In this study, we use VR-CESM to quantify the impacts of LAA (BC and
 dust) deposition ~~on~~ ^{to} the snowpack and hydrologic cycles ^{and to surface air temperatures} in the Rocky Mountains. Our
 previous study has shown that VR-CESM reproduces reasonably the spatial
 distributions and seasonal evolution of snowpack in the Rocky Mountains (Wu et al.,
 2017). Here we show that the model also reproduces observed spatial distributions of
 near-surface ^{atmospheric} BC and dust concentrations in the western U.S. compared to IMPROVE
 observations. The magnitude of simulated near-surface dust concentrations is
 comparable to observations, while that of simulated near-surface BC concentrations ^{are} ~~is~~
 mostly underestimated, especially in the Rocky Mountain region. The
 underestimation of near-surface BC concentrations may be due to the absence of local
 sources ^{in the} ~~from~~ BC emissions ^{dataset} ~~used~~ in the model. Simulated aerosol-in-snow
 concentrations are closely related to the distributions of both snowpack and
 near-surface atmospheric aerosol concentrations. Simulated BC-in-snow
 concentrations are mostly comparable to the observations with the magnitude between
 2 and 30 ng g⁻¹, although significant underestimations (by as much as a factor of 10)
 were found at two stations.

Due to the deposition of LAAs ^{to} ~~on~~ snow ~~cover~~, surface net shortwave radiation
^{Regionally- and monthly-} ~~Regional~~ averaged SRE induced by LAAs in snow is 0.1-0.5 W m⁻² in



648 winter in the three mountainous regions (Northern Rockies, Greater Yellowstone
649 region, and Southern Rockies) and $0.4\text{--}0.6\text{ W m}^{-2}$ in the two regions around the
650 mountains (Eastern Snake River Plain and Southwestern Wyoming). **Monthly average** SRE is much
651 larger in spring and reaches up to $0.6\text{--}1.7\text{ W m}^{-2}$ in these five regions (Table 2). Dust
652 contributes 21–43% to the total SRE induced by LAAs in snow in spring, indicating
653 the important role of dust residing in snow. Of the five regions, dust contributes the
654 most (43%) to the total SRE in the Southern Rockies. This is not unexpected as this
655 region is close to dust sources in the Colorado Plateau.

656 As a result of SRE induced by LAAs in snow, surface air temperature
657 increases in most of the Rocky Mountain region. The surface air temperature increase
658 is largest over the Eastern Snake River Plain and Southwestern Wyoming, with winter
659 and spring surface air temperature increased by $0.9\text{--}1.1^{\circ}\text{C}$. Significant reductions of
660 snow water equivalent (by 2–50 mm) and snow cover fraction (by 5–20%) occur in
661 these two regions, indicating a strong positive snow-albedo feedback there.

662 Aerosol SDE accelerates the hydrologic cycle in the mountainous regions. In
663 April and May, monthly mean runoff is increased by 7%–42% in the three
664 mountainous regions (Northern Rockies, Greater Yellowstone region, Southern
665 Rockies). This is because of the accelerated snowmelt resulting from surface warming
666 as well as the increased snowfall resulting from enhanced water vapor transport from
667 the Pacific Ocean. The enhanced water vapor transport may be related to large-scale
668 circulation changes. In the later stage of snowmelt, monthly runoff is reduced by 2–15%



669 in June and 8-23% in July in the three mountainous regions. In particular, aerosol
670 SDE leads to a reduction of total runoff by about 15% in June and July in the
671 Southern Rockies. This highlights the important role of aerosol SDE in modulating
672 ~~the~~ ^{these} the hydrologic cycle in ~~the~~ mountainous regions.

673 We note that VR-CESM still underestimates the near-surface BC
674 concentrations, however, reproduces observed magnitudes of BC-in-snow
675 concentrations for most stations. For dust in snow, the model used in this study only
676 accounts for dust particles smaller than 10 μm , while observations made by Reynolds
677 et al. (2016) suggest that most airborne dust concentrations are characterized by dust
678 particles with diameters larger than 10 μm in the Utah-Colorado region. Therefore,
679 our simulations may significantly underestimate the impacts of dust in snow. Actually,
680 our simulations suggest SRE induced by dust-in-snow can reach up to 2-5 W m^{-2} in
681 the Southern Rockies, which is nearly an order of magnitude smaller than values ~~in~~ ^{given by}
682 Painter et al. (2007) based on observed dust-in-snow particles. Future observations of
683 LAAs in snow, particularly for the temporal evolution of LAAs in different snow
684 layers, as well as detailed size distribution measurements of dust particles in snow
685 will help reduce the uncertainties in the model quantification of the impacts of LAAs
686 in snow.

687 Although ~~the~~ uncertainties still exist, our results show LAAs in snow can
688 significantly affect the snowpack and consequent hydrologic cycle in the Rocky
689 Mountains. Previous studies have demonstrated that snowpack on the Rocky



690 Mountains has declined significantly in the second half of 20th century (e.g.,
691 Pederson et al., 2011). The role of LAAs in this decrease of snowpack is still
692 unknown. It would be interesting to investigate the role of LAAs and compare it with
693 those of other climate factors (such as natural climate variability and greenhouse gas
694 concentrations). Moreover, BC and dust emissions may also be subject to changes in
695 the future. Therefore, for better projections of future changes in Rocky Mountain
696 snowpack, the impacts of LAAs in snow under future emissions scenarios ~~also~~ need to
697 be taken into account.

698

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Table 1. Observations of BC mass concentration in snow column (C_{BC} , ng g^{-1} , i.e., ng gram BC per g snow) in the Rocky Mountain region compiled from previously published literature.

No.	Latitude (°N)	Longitude (°W)	Elevation (m)	Date sampled	C_{BC} (ng g^{-1}) ^a	Source
1	40.9014	115.8910	1949	2/1/13	7.6	Site 8 of Doherty et al. (2014)
2	42.2767	116.0115	1772	2/1/13	5.6	Site 9 of Doherty et al. (2014)
3	43.3495	115.3968	1538	2/3/13	6.0	Site 10 of Doherty et al. (2014)
4	43.5927	113.5894	1942	2/3/13	5.8	Site 11 of Doherty et al. (2014)
5	43.4010	111.2053	1727	2/4/13	6.8	Site 12 of Doherty et al. (2014)
6	42.9357	109.8576	2274	2/4/13	6.3	Site 13 of Doherty et al. (2014)
7	41.7297	109.3668	2223	2/5/13	29.1	Site 14 of Doherty et al. (2014)
8	40.7464	109.4776	2583	2/5/13	9.3	Site 15 of Doherty et al. (2014)
9	40.1316	109.4711	1538	2/7/13	14.3	Site 16 of Doherty et al. (2014)
10	40.4929	107.8994	1962	2/8/13	9.5	Site 17 of Doherty et al. (2014)
11	40.6695	106.4158	2512	2/9/13	11.4	Site 18 of Doherty et al. (2014)
12	48.2318	105.0949	648	2/17/13	10.9	Site 24 of Doherty et al. (2014)
13	44.9475	116.0813	1528	January to March, 2014	9.8 (5.4)	Site McCall of Doherty et al. (2016)
14	44.4224	115.9899	1450	February to March, 2014	13.3 (9.5)	Site Cascade Valley of Doherty et al. (2016)
15	44.0949	115.9771	960	February to March, 2014	14.9 (8.9)	Site Garden Valley of Doherty et al. (2016)
16	40.143	109.467	1620	February to March, 2013 and 2014	33.6 (25.4)	Site Vernal of Doherty et al. (2016)
17	37.9069	107.7113	3368	March to May, 2013	9.1 (7.9)	At Senator Beck Basin Study Area (SBBSA) (Skiles and Painter, 2016b)

^a: If multi-measurements of C_{BC} are made during the observation period, the mean C_{BC} is given with the standard deviation of C_{BC} shown in parenthesis next to the mean C_{BC} .



964

965 **Table 2.** Winter (December-January-February) and spring (March-April-May) mean
 966 surface shortwave radiative effect (SRE; W m^{-2}) due to BC alone, dust alone and BC
 967 and dust together in snow, as well as surface air temperature (SAT; $^{\circ}\text{C}$) change and
 968 the efficacy of SRE in SAT change in the five regions (see Figure 1c). Note that SRE
 969 induced by BC and dust together is slightly larger than the sum of SRE induced by
 970 BC and SRE by dust separately.

971

Season	SRE by BC ^a	SRE by dust ^a	SRE by BC & dust	SAT change	Efficacy ^b
Northern Rockies					
Winter	0.13 (92%)	0.01 (8%)	0.14	0.08	0.57
Spring	0.42 (79%)	0.11 (21%)	0.57	0.32	0.56
Greater Yellowstone region					
Winter	0.24 (88%)	0.03 (12%)	0.28	0.004	0.014
Spring	1.11 (71%)	0.45 (29%)	1.70	0.50	0.29
Southern Rockies					
Winter	0.36 (77%)	0.11 (23%)	0.50	0.17	0.34
Spring	0.79 (58%)	0.58 (42%)	1.58	0.30	0.19
Eastern Snake River Plain					
Winter	0.50 (84%)	0.09 (16%)	0.62	0.93	1.5
Spring	0.54 (73%)	0.20 (27%)	0.80	1.13	1.41
Southwestern Wyoming					
Winter	0.33 (81%)	0.08 (19%)	0.43	0.93	2.16
Spring	0.43 (67%)	0.22 (33%)	0.70	0.90	1.29

972 ^a: The fraction of SRE by BC (dust) to the sum of SRE by BC and SRE by dust is given in
 973 parenthesis next to SRE by BC (dust).

974 ^b: Efficacy of snow/ice albedo forcing ($^{\circ}\text{C}$ increase per 1 W m^{-2}) is defined as the ratio of SAT
 975 change to SRE.

976



977 **Figure captions:**

978

979 **Figure 1.** (a) Model meshes for variable resolution (uniform 1° with refined 0.125° in
 980 the Rocky Mountains) used in VR-CESM. Note that each element shown contains
 981 additional 3×3 collocation gridcells. (b) Terrain height (m) in the western US with the
 982 refined region at a resolution of 0.125° surrounded by dashed lines. (c) Five regions
 983 identified for the analysis in this study, including three mountainous region (1,
 984 Northern Rockies; 2, Greater Yellowstone region; 3, Southern Rockies) and two
 985 regions in the plains around the mountains (4, Eastern Snake River Plain; 5,
 986 Southwestern Wyoming).

987 **Figure 2.** Spatial distribution of cold season (winter and spring) mean (a) BC
 988 emission flux and (b) near-surface BC concentration from the VR-CESM simulation;
 989 (c) and (d) for dust emission flux and near-surface dust concentration, respectively.
 990 Also shown are the IMPROVE stations (blue open circle) selected for model
 991 validation, with the size of the circles from small to large indicating the magnitude of
 992 observed near-surface BC/dust concentrations. The black rectangles in (b) and (d)
 993 denotes the five regions (A, West Coast; B, Rocky Mountains; C, Utah and Nevada;
 994 D, Southwestern US; E, Great Plain), which will be used to classify the stations in
 995 Figure 3. Note that units for BC and dust concentrations are ng m^{-3} and $\mu\text{g m}^{-3}$,
 996 respectively.

997 **Figure 3.** Comparison of cold season mean near-surface (a) BC and (b) dust
 998 concentrations at IMPROVE stations from VR-CESM simulation and IMPROVE
 999 observations. Also given are the mean results at all the stations from simulation and
 1000 observations and their correlation coefficient (R). The 1:1 (solid) and 1:5/5:1 (dash)
 1001 lines are plotted for reference.

1002 **Figure 4.** Winter (December-January-February (DJF); left) and spring
 1003 (March-April-May (MAM); right) mean BC (upper row) and dust (bottom row) mass
 1004 mixing ratios in snow column. Also shown are the stations for observations of BC
 1005 mass in snow column (a, b) and for observations of dust mass in snow column at
 1006 Senator Beck Basin Study Area (SBBBSA) in the San Juan Mountains (c, d). Note that
 1007 the units for BC and dust mass mixing ratios are given in different units, i.e., ng g^{-1}
 1008 and $\mu\text{g g}^{-1}$, respectively.

1009 **Figure 5.** Comparison of BC mass concentrations in the snow column (C_{BC}) at the 17
 1010 sites (see Table 1) from VR-CESM simulations and observations with the error bars



denoting the corresponding standard deviations. The observations are compiled from the previously published studies (Table 1). If multiple observations are recorded at a certain site, the observed standard deviations are calculated from these multiple observations (section 3). Simulated BC mass concentration in the snow column and its standard deviation are calculated from the 25-year mean and standard deviation of simulation in the same month as the observations (section 3).

Figure 6. Winter (December-January-February (DJF), left) and spring (March-April-May (MAM), right) mean surface shortwave radiative effect (SRE, W m^{-2}) induced by BC (top) and dust (bottom).

Figure 7. Monthly variations of surface radiative effect (SRE; W m^{-2}) during the water year (October 1st to September 30th) averaged over the Northern Rockies, Greater Yellowstone region, Southern Rockies, Eastern Snake River Basin, and Southwestern Wyoming, respectively.

Figure 8. Changes in surface air temperature (upper row; $^{\circ}\text{C}$), snow water equivalent (bottom row; mm), and snow cover fraction (bottom row; %) in winter (left) and spring (right) induced by BC- and dust-in-snow. The crosses denote the regions where changes are statistically significant at 0.1 level.

Figure 9. As Figure 8, but for total precipitation change (top), rainfall change (center), and snowfall change (bottom). The unit is mm day^{-1} .

Figure 10. Seasonal evolution of (a) surface air temperature, (c) snow water equivalent, and (e) snow cover fraction and their changes due to SDE (b, d, and f) averaged over the Eastern Snake River Plain.

Figure 11. As Figure 10, but for Southwestern Wyoming.

Figure 12. Snowmelt change (mm day^{-1}) due to SDE of BC and dust in four seasons: (a) December-January-February (DJF), (b) March-April-May (MAM), (c) June-July-August (JJA), and (d) September-October-November (SON). The crosses denote the regions where changes induced by SDE are statistically significant at 0.1 level.

Figure 13. As Figure 12, but for runoff change (mm day^{-1}).

Figure 14. Seasonal evolution of runoff (left) and their change (right) in the Northern Rockies (top), the Greater Yellowstone region (center), and Southern Rockies (bottom). The unit is mm day^{-1} .

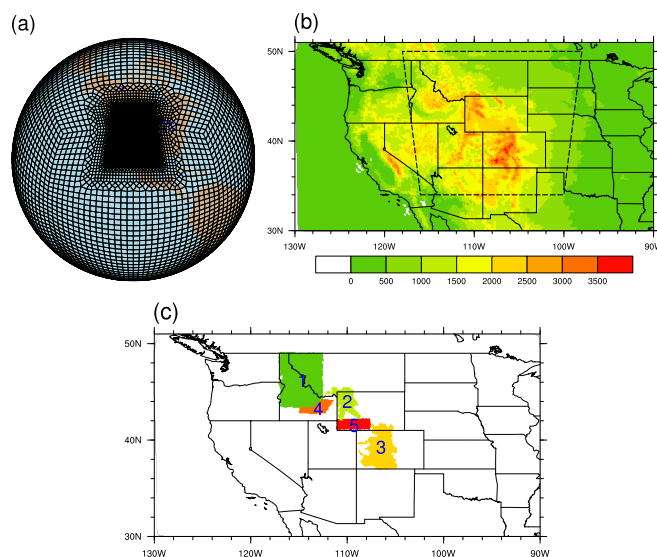
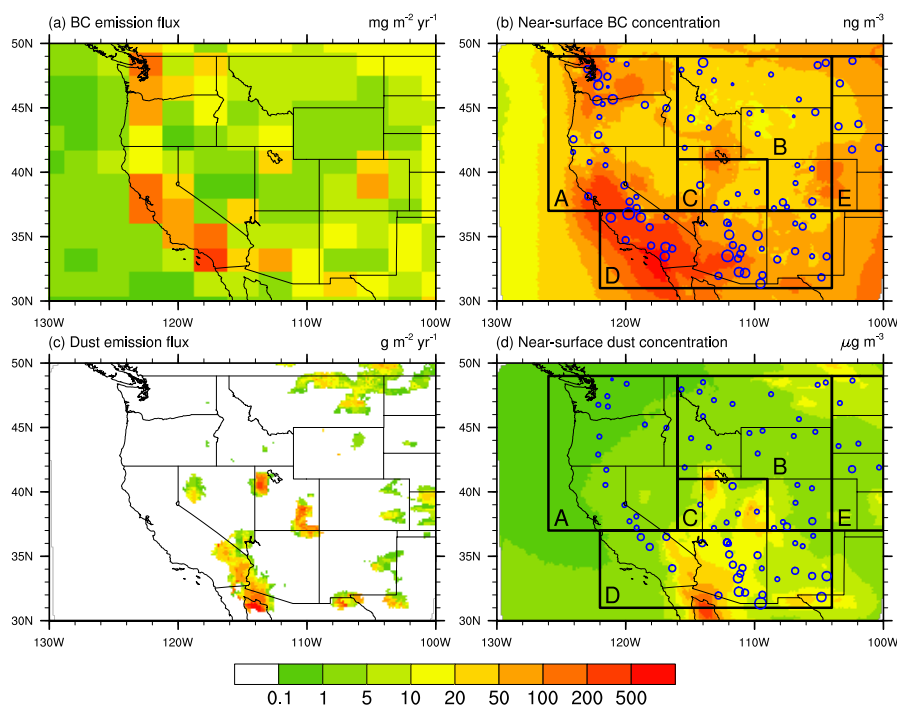


Figure 1. (a) Model meshes for variable resolution (uniform 1° with refined 0.125° in the Rocky Mountains) used in VR-CESM. Note that each element shown contains additional 3×3 collocation gridcells. (b) Terrain height (m) in the western US with the refined region at a resolution of 0.125° surrounded by dashed lines. (c) Five regions identified for the analysis in this study, including three mountainous region (1, Northern Rockies; 2, Greater Yellowstone region; 3, Southern Rockies) and two regions in the plains around the mountains (4, Eastern Snake River Plain; 5, Southwestern Wyoming).



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1061 **Figure 2.** Spatial distribution of cold season (winter and spring) mean (a) BC
 1062 emission flux and (b) near-surface BC concentration from the VR-CESM simulation;
 1063 (c) and (d) for dust emission flux and near-surface dust concentration, respectively.
 1064 Also shown are the IMPROVE stations (blue open circle) selected for model
 1065 validation, with the size of the circles from small to large indicating the magnitude of
 1066 observed near-surface BC/dust concentrations. The black rectangles in (b) and (d)
 1067 denotes the five regions (A, West Coast; B, Rocky Mountains; C, Utah and Nevada;
 1068 D, Southwestern US; E, Great Plain), which will be used to classify the stations in
 1069 Figure 3. Note that units for BC and dust concentrations are ng m^{-3} and $\mu\text{g m}^{-3}$,
 1070 respectively.

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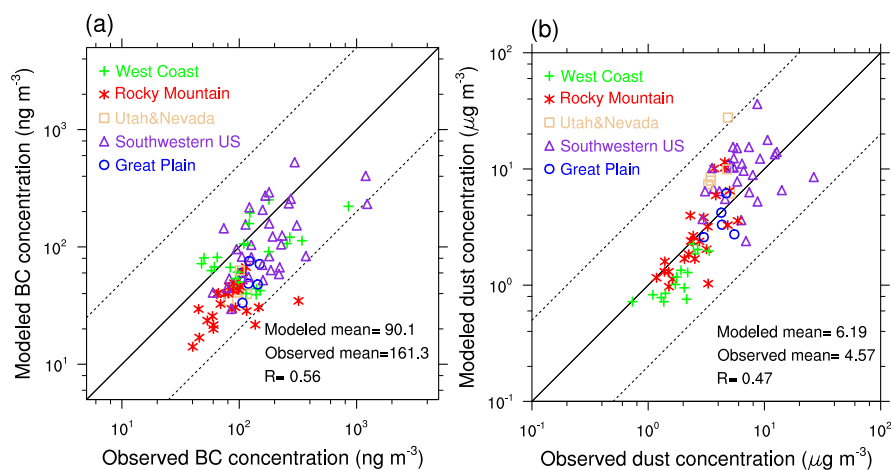
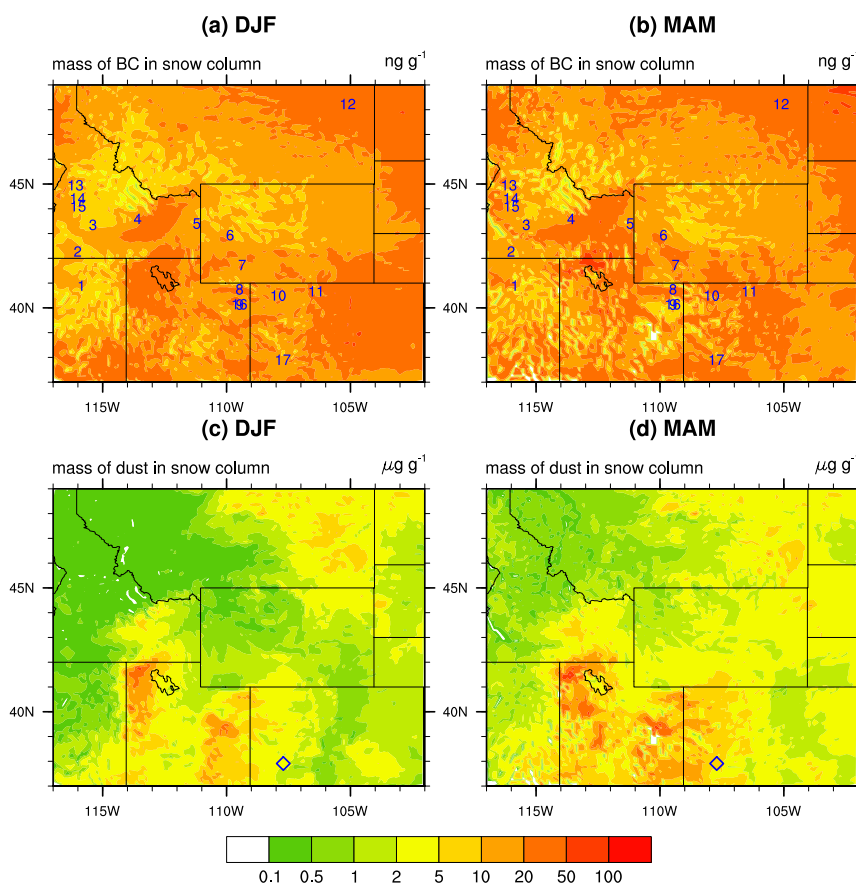


Figure 3. Comparison of cold season mean near-surface (a) BC and (b) dust concentrations at IMPROVE stations from VR-CESM simulation and IMPROVE observations. Also given are the mean results at all the stations from simulation and observations and their correlation coefficient (R). The 1:1 (solid) and 1:5/5:1 (dash) lines are plotted for reference.



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1084 **Figure 4.** Winter (December-January-February (DJF); left) and spring

1085 (March-April-May (MAM); right) mean BC (upper row) and dust (bottom row) mass

1086 mixing ratios in snow column. Also shown are the stations for observations of BC

1087 mass in snow column (a, b) and for observations of dust mass in snow column at

1088 Senator Beck Basin Study Area (SBBSA) in the San Juan Mountains (c, d). Note that

1089 the units for BC and dust mass mixing ratios are given in different units, i.e., ng g^{-1} 1090 and $\mu\text{g g}^{-1}$, respectively.

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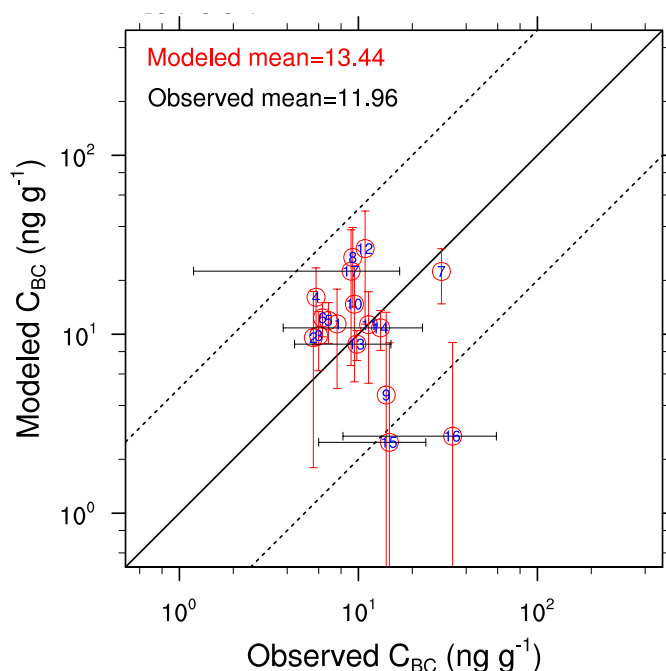
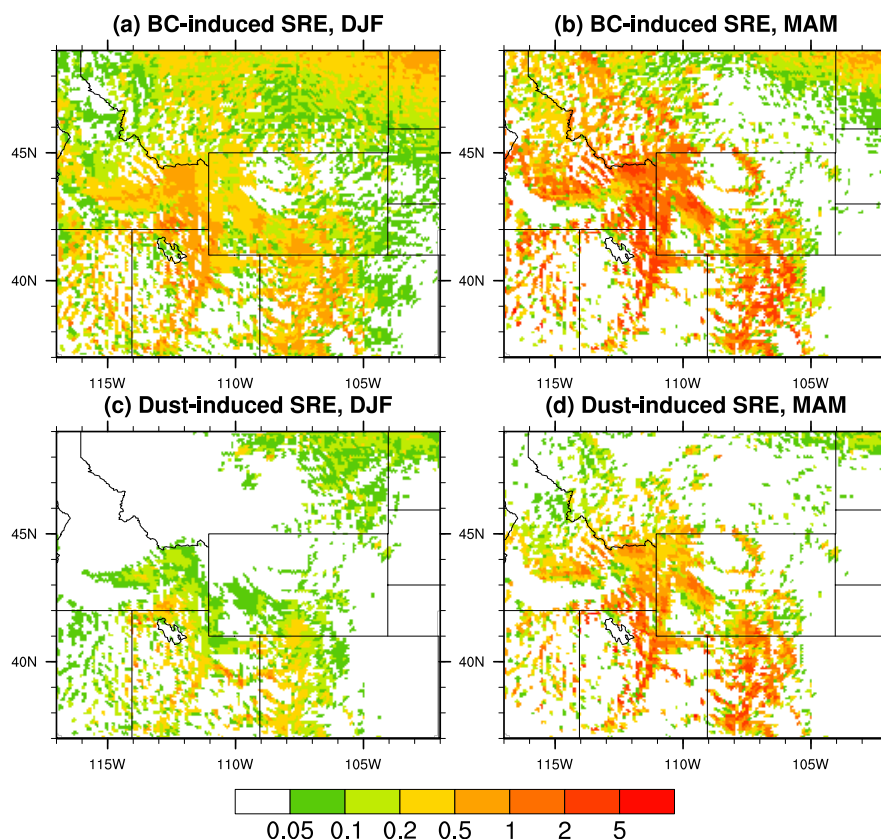


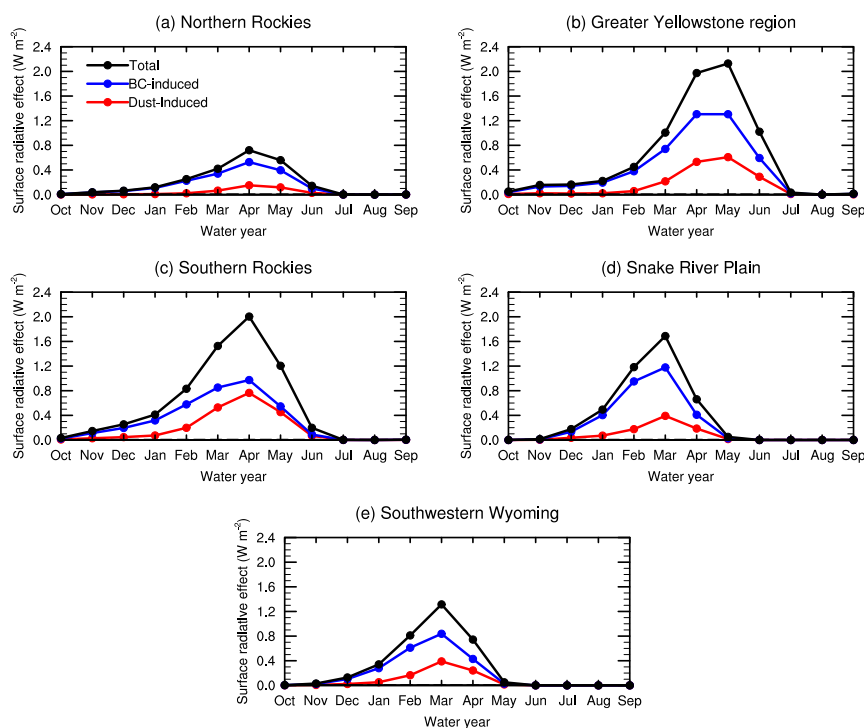
Figure 5. Comparison of BC mass concentrations in the snow column (C_{BC}) at the 17 sites (see Table 1) from VR-CESM simulations and observations with the error bars denoting the corresponding standard deviations. The observations are compiled from the previously published studies (Table 1). If multiple observations are recorded at a certain site, the observed standard deviations are calculated from these multiple observations (section 3). Simulated BC mass concentration in the snow column and its standard deviation are calculated from the 25-year mean and standard deviation of simulation in the same month as the observations (section 3).



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 1106 **Figure 6.** Winter (December-January-February (DJF), left) and spring
 1107 (March-April-May (MAM), right) mean surface shortwave radiative effect (SRE, W
 1108 m^{-2}) induced by BC (top) and dust (bottom).
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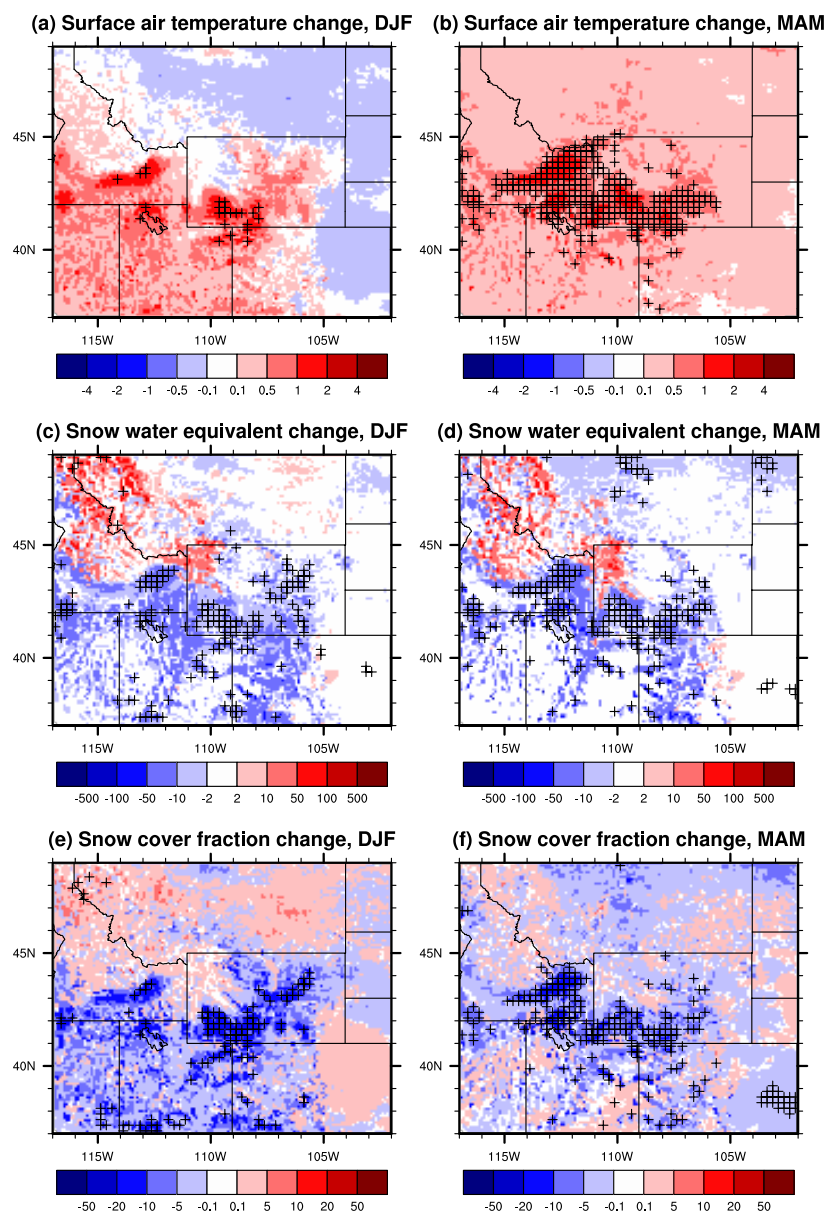
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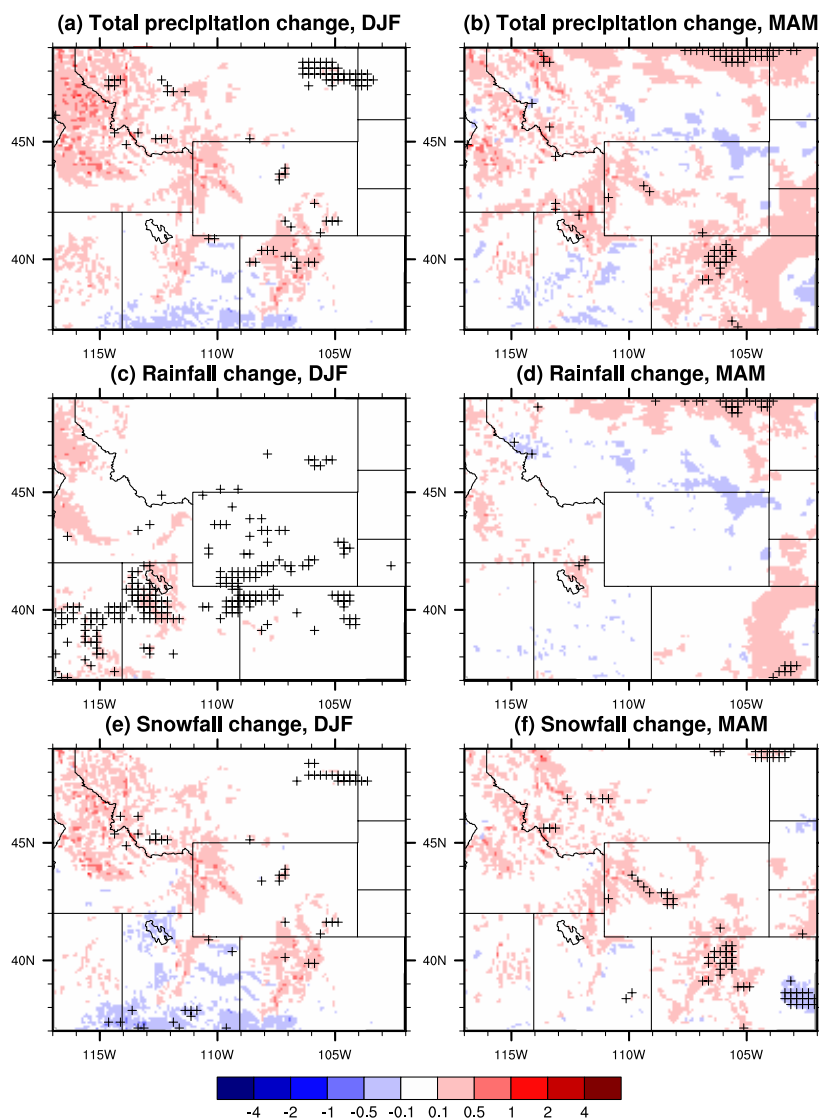
1111

1112 **Figure 7.** Monthly variations of surface radiative effect (SRE; W m^{-2}) during the
 1113 water year (October 1st to September 30th) averaged over the Northern Rockies,
 1114 Greater Yellowstone region, Southern Rockies, Eastern Snake River Basin, and
 1115 Southwestern Wyoming, respectively.

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 1118 **Figure 8.** Changes in surface air temperature (upper row; °C), snow water equivalent
 1119 (bottom row; mm), and snow cover fraction (bottom row; %) in winter (left) and
 1120 spring (right) induced by BC- and dust-in-snow. The crosses denote the regions where
 1121 changes are statistically significant at 0.1 level.
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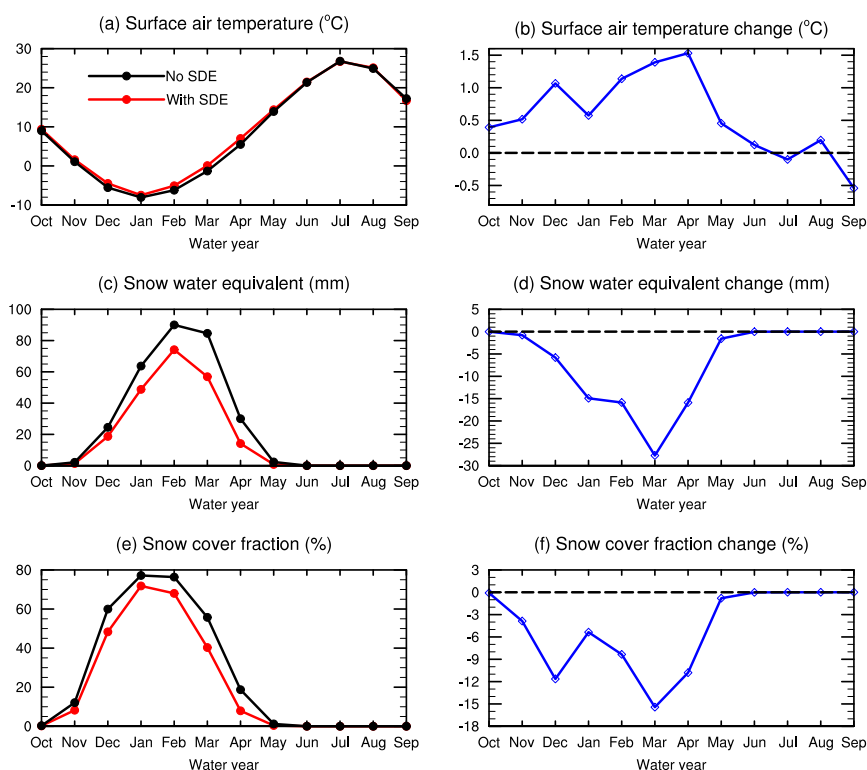
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1124 **Figure 9.** As Figure 8, but for total precipitation change (top), rainfall change (center),1125 and snowfall change (bottom). The unit is mm day^{-1} .

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 1130 **Figure 10.** Seasonal evolution of (a) surface air temperature, (c) snow water
 1131 equivalent, and (e) snow cover fraction and their changes due to SDE (b, d, and f)
 1132 averaged over the Eastern Snake River Plain.
 1133

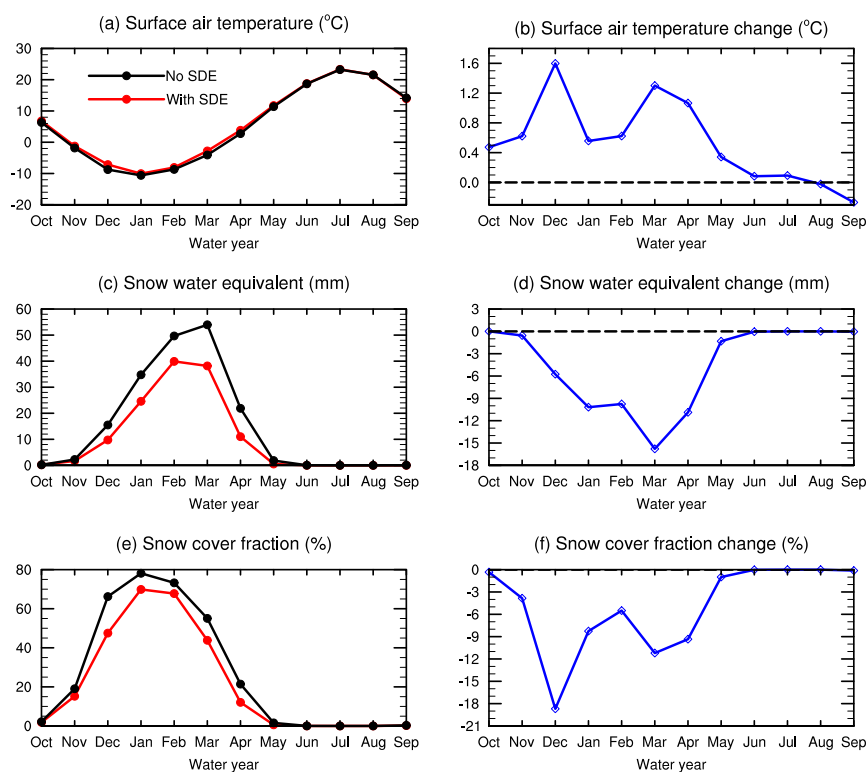
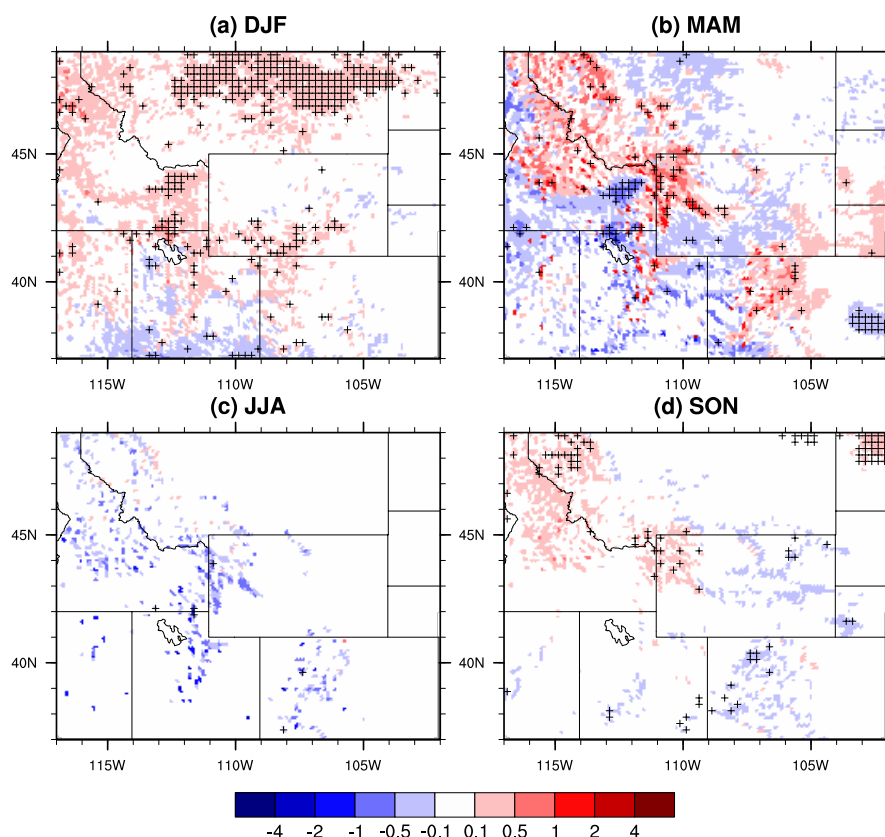


Figure 11. As Figure 10, but for Southwestern Wyoming.



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1138 **Figure 12.** Snowmelt change (mm day^{-1}) due to SDE of BC and dust in four seasons:

1139 (a) December-January-February (DJF), (b) March-April-May (MAM), (c)

1140 June-July-August (JJA), and (d) September-October-November (SON). The crosses

1141 denote the regions where changes induced by SDE are statistically significant at 0.1

1142 level.

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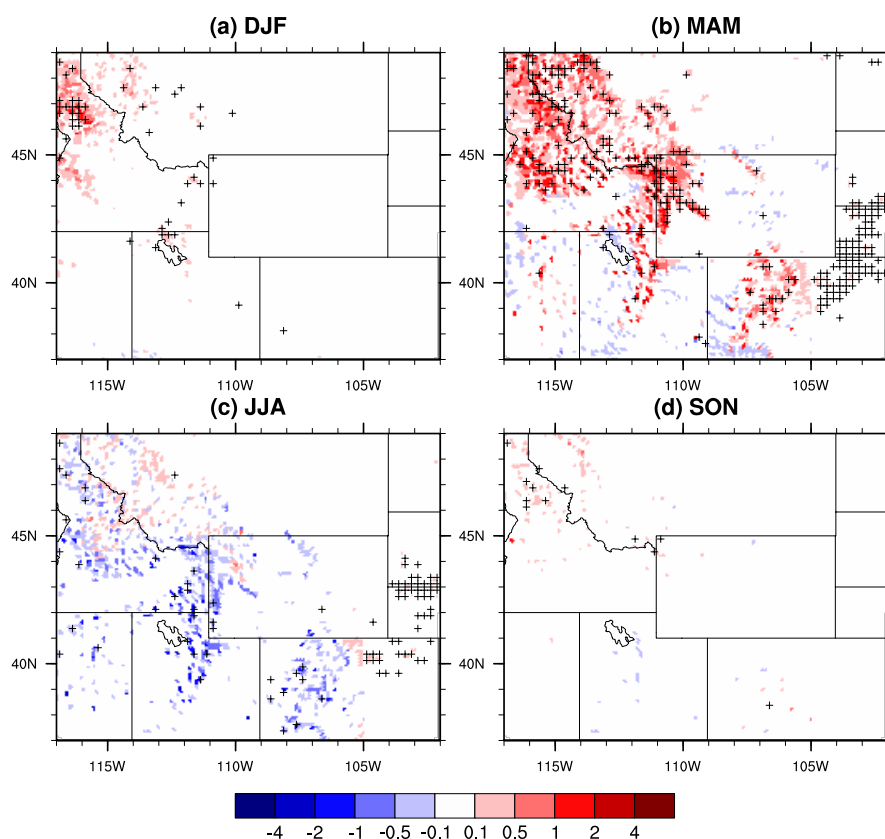
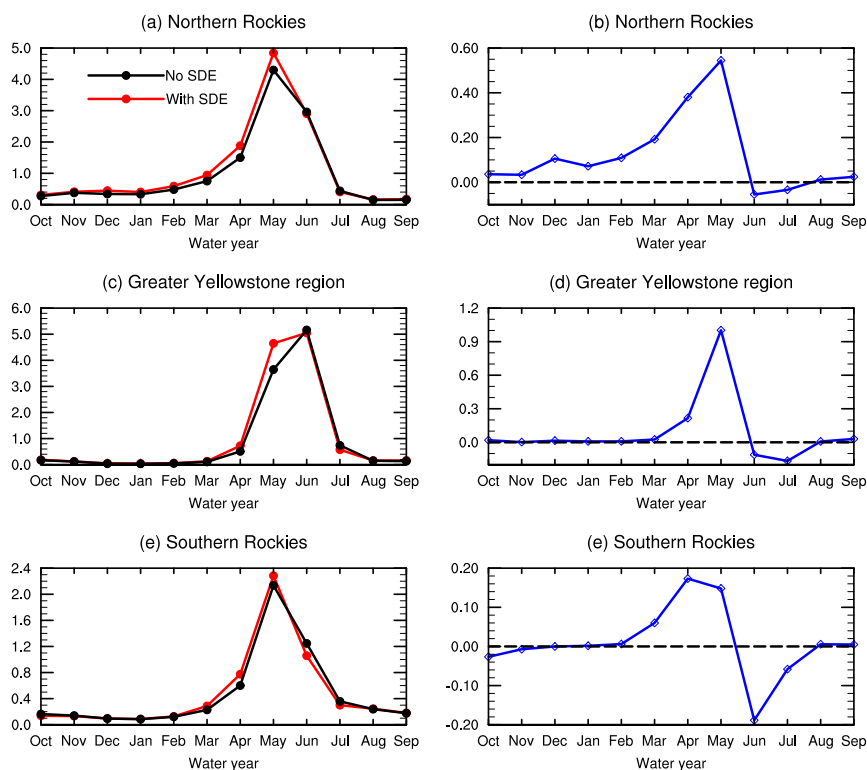


Figure 13. As Figure 12, but for runoff change (mm day^{-1}).



1149
1150 **Figure 14.** Seasonal evolution of runoff (left) and their change (right) in the Northern
1151 Rockies (top), the Greater Yellowstone region (center), and Southern Rockies
1152 (bottom). The unit is mm day⁻¹.