

# Snow cover sensitivity to black carbon deposition in the Himalayas: from atmospheric and ice core measurements to regional climate simulations

Ménégoz, M. <sup>1,2</sup>, Krinner, G. <sup>1,2</sup>, Balkanski, Y. <sup>3</sup>, Boucher, O. <sup>4</sup>, Cozic, A. <sup>3</sup>, Lim, S. <sup>1,2</sup>, Ginot, P. <sup>1,2,5</sup>, Laj, P. <sup>1,2</sup>, Gallée, H. <sup>1,2</sup>, Wagon, P. <sup>8,9</sup>, Marinoni, A. <sup>6,7</sup> and Jacobi, H.W. <sup>1,2</sup>.

[1] {CNRS Grenoble 1, Laboratoire de Glaciologie et Géophysique de l'Environnement (LGGE, UMR5183), 38041 Grenoble, France.}

[2] {UJF Grenoble 1, Laboratoire de Glaciologie et Géophysique de l'Environnement (LGGE, UMR5183), 38041 Grenoble, France.}

[3] {Laboratoire des Sciences du Climat et de l'Environnement, IPSL, CEA-CNRS-UVSQ, Gif-sur-Yvette, France.}

[4] {Laboratoire de Météorologie Dynamique, IPSL, CNRS, Paris, France.}

[5] {IRD / UJF - Grenoble 1 / CNRS / U. Savoie / INPG / IFSTTAR / CNRM, Observatoire des Sciences de l'Univers de Grenoble (OSUG) UMS222, Grenoble, F-38041, France.}

[6] {CNR-ISAC-Institute of Atmospheric Sciences and Climate, Bologna, Italy.}

[7] {EV-K2-CNR Committee, Bergamo, Italy.}

[8] {IRD / UJF - Grenoble 1 / CNRS / G-INP, LTHE UMR 5564, LGGE UMR 5183, Grenoble, F-38402, France}

[9] {ICIMOD, GPO Box 3226, Kathmandu, Nepal}

Correspondence to Martin Ménégoz ([martin.menegoz@lgge.obs.ujf-grenoble.fr](mailto:martin.menegoz@lgge.obs.ujf-grenoble.fr))

## Abstract

We applied a climate-chemistry global model to evaluate the impact of black carbon (BC) deposition on the Himalayan snow cover from 1998 to 2008. Using a stretched grid with a resolution of 50 km over this complex topography, the model reproduces reasonably well the remotely sensed observations of the snow cover duration. Similar to observations, modelled atmospheric BC concentrations in Central Himalayas reach a minimum during the monsoon and a maximum during the post- and pre-monsoon periods. Comparing the simulated BC concentrations

29 in the snow with observations is more challenging because of their high spatial variability and  
30 complex vertical distribution. We simulated spring BC concentrations in surface snow varying from  
31 tens to hundreds of  $\mu\text{g kg}^{-1}$ , higher by one to two orders of magnitude than those observed in ice  
32 cores extracted from Central Himalayan glaciers at high elevations ( $> 6000$  m a.s.l.), but typical for  
33 seasonal snow cover sampled in middle elevation regions ( $< 6000$  m a.s.l.). In these areas, we  
34 estimate that both wet and dry BC depositions affect the Himalayan snow cover reducing its annual  
35 duration between one and eight days. In our simulations, the effect of anthropogenic BC deposition  
36 on snow is quite low over the Tibetan Plateau because this area is only sparsely snow covered.  
37 However, the impact becomes larger along the entire Hindu-Kush, Karakorum and Himalayan  
38 mountain ranges. In these regions, BC in snow induces an increase of the net shortwave radiation at  
39 the surface with an annual mean of 1 to 3  $\text{W m}^{-2}$  leading to a localised warming between 0.05 and  
40 0.3  $^{\circ}\text{C}$ .

## 41 1 Introduction

42 Black carbon (BC) is one of the major anthropogenic pollutants affecting the climate system. Bond  
43 et al. (2013) estimated the global climate forcing of BC through all forcing mechanisms to be about  
44  $+1.1 \text{ W m}^{-2}$  with a 90 % probability to be included in a range of  $+0.17$  to  $+2.1 \text{ W m}^{-2}$ . This value  
45 includes the net effect of BC on radiation and clouds, but also that on snow albedo, which has been  
46 found to strongly impact the climate system (Hansen and Nazarenko, 2004). Anthropogenic BC  
47 deposited on snow was found to shorten the current duration of the snow cover season in the  
48 Northern Hemisphere by several days (Ménégoz et al., 2013a) and to contribute to the significant  
49 decrease of the snow cover extent observed during the last decades (Dery et al., 2007). Brutel-  
50 Vuilmet et al. (2013) found the new generation of global climate model to correctly simulate the  
51 present-day snow cover extent, but to underestimate its decrease over the last decades. They also  
52 noted that coarse-gridded models simulate the snow cover particularly badly over mountainous  
53 areas.

54 The Hindu-Kush-Karakoram-Himalayas region, denominated in the following as the HKKH hosts  
55 extended glaciers (Kääb et al., 2012). In addition, wide areas of North Western and Eastern  
56 Himalayas are seasonally snow covered during long periods, whereas in central Himalaya the snow  
57 cover extent is rather limited (Ménégoz et al., 2013b). The HKKH atmosphere is strongly affected  
58 by anthropogenic emissions of BC originating from the Indian plain and highly populated  
59 mountainous areas (Ohara et al., 2007). Consequently, atmospheric BC can reach very high  
60 concentrations (Ramanathan et al., 2007) even at high altitudes (Bonasoni et al., 2010, Kopacz et  
61 al., 2011, Marinoni et al., 2013). Ice cores drilled in this region have shown that aerosol pollution is

62 incorporated into the snowpack even at very high altitudes (e.g., Ming et al., 2008, 2009, Xu et al.,  
63 2009, Ginot et al., 2013, Kaspari et al., 2013). Yasunari et al. (2010, 2013) estimated that BC in  
64 snow reduces the snow albedo on average by 5 % on the southern slopes of the Nepalese Himalaya.  
65 Kopacz et al. (2011) found the radiative forcing due to BC on snow to vary from 5 to 15 W m<sup>-2</sup>  
66 within the snow-covered areas of this region, while Flanner et al. (2007) and Qian et al. (2011)  
67 estimated peak values exceeding 20 W m<sup>-2</sup> for some parts of the Tibetan plateau. Menon et al.  
68 (2010) proposed that during the last decade BC in snow caused a significant part of the decrease of  
69 the snow cover extent observed in this region.

70 Since these last modelling studies have been based on relatively coarse-gridded models, their ability  
71 to simulate the snow cover over mountainous areas remains relatively limited. Here, we use a global  
72 climate model with a stretched grid to reach a fine resolution over the HKKH in order to quantify  
73 the effect of BC deposition on the snow cover duration. The temporal variations and magnitude of  
74 the BC concentrations both in the atmosphere and snow are compared to observations. Finally, we  
75 estimate the snow cover duration, surface radiation, and temperature changes induced by the BC  
76 deposition on the Himalayan snow.

## 77 **2 Experimental setup**

### 78 **2.1 The LMDZ-ORCHIDEE-INCA climate model**

79 We used the LMDZ-ORCHIDEE-INCA atmospheric general circulation model to study the  
80 interactions between the atmosphere, aerosols, and snow-covered areas in the HKKH. This model  
81 consists of three coupled modules. The LMDZ general circulation model represents the atmospheric  
82 component (Hourdin et al., 2006). The ORCHIDEE land surface model describes exchanges of  
83 energy and water between the atmosphere, the soil and the biosphere (Krinner et al., 2005)  
84 including a dynamic snow module. The coupling between LMDZ and ORCHIDEE is described in  
85 Hourdin et al. (2006). INCA (interactions between chemistry and aerosols) describes gas- and  
86 aqueous-phase chemistry (Hauglustaine et al., 2004) as well as aerosol physical properties such as  
87 size and hygroscopicity (Balkanski et al., 2010), which control the amount of wet and dry  
88 deposition. The coupling of LMDZ and INCA described by Szopa et al. (2012) allows an  
89 interactive simulation of five aerosol chemical species: sulphate, BC, organic carbon (OC), sea salt,  
90 and dust. We consider aerosol-radiation interactions for BC, OC, sea salt, and dust and aerosol-  
91 cloud interactions for sulphate, BC, and OC as described in Déandreis et al. (2012). All the  
92 experiments were conducted with the present-day global aerosol emission inventory described in  
93 Lamarque et al. (2010), a decadal resolved inventory made for the Coupled Model Inter-comparison  
94 Project Phase 5 (CMIP5, CLIVAR special issue, 2011). We further used the detailed representation

95 of snow cover implemented in ORCHIDEE by Krinner et al. (2006) and used in Ménégoz et al.  
96 (2013a). This includes a two-layer scheme describing the snow albedo as a function of snow grain  
97 size and aerosol content in the snow based on Wiscombe and Warren (1980). The representation of  
98 snow grain size and BC in the snow and the snow albedo scheme implemented in our model are  
99 detailed in Ménégoz et al. (2013a). Since we want our simulation to be in phase with the  
100 atmospheric observations, in particular with the East Asian and the Indian monsoons that bring  
101 large amounts of moisture into the Himalayas, all simulations were performed with winds nudged  
102 towards the ECMWF ERA-interim re-analysis: Each 150 seconds (i.e. with a time step 5 times  
103 longer than those used to compute wind velocities), horizontal wind velocities are nudged at all  
104 altitudes with a relaxation time of 1 hour over the HKKH region and 30 min elsewhere. Hence, the  
105 model is very constrained by the reanalysis outside the HKKH, whereas it evolves more  
106 independently inside (see details in Coindreau et al., 2007).

## 107 **2.2 Resolution of the simulation**

108 We performed simulations for the 1998-2008 period with two different grids: A first one with a  
109 regular coarse horizontal resolution (96x95 grid points corresponding to a ~350 km resolution), and  
110 a second one with a 143x144 stretched grid with a zoom on the Himalaya reaching a 50 km  
111 resolution over this region. The Himalayan Mountains located over 3000 m are poorly described  
112 with the coarse grid, whereas the stretched grid allows a more realistic representation of the  
113 topography both along the Himalayan arc and over the Tibetan Plateau (not shown). Comparisons  
114 with satellite observations (Fig. 1) show that a fine resolution is essential to simulate correctly the  
115 observed snow cover duration over the HKKH and the Tibetan Plateau: The coarse-gridded  
116 simulation (Fig. 1a) shows a strong overestimation of the snow cover duration, in particular over the  
117 Tibetan Plateau, in comparison with satellite observations interpolated onto the same grid (Fig. 1c).  
118 Interpolating the output of the simulation based on the stretched grid onto the coarse grid (Fig. 1b),  
119 we see that this bias is strongly reduced. Such an improvement is due to a better representation of  
120 the local atmospheric circulation and surface energy balance associated with climate feedbacks  
121 using a fine resolution. The simulation based on the stretched grid is quite similar to the satellite  
122 observation (Fig. 1d and 1e) with high values of snow cover duration in the high mountains of the  
123 HKKH and large parts of the Tibetan Plateau free of snow most of the year. Nevertheless, the  
124 overestimation of the snow cover duration over the Tibetan Plateau does not fully disappear when  
125 increasing the resolution of the model. In the following, we focus only on simulations performed  
126 with the stretched grid, which appears essential to describe snow cover variations in the HKKH.

## 127 **3 Aerosol deposition on snow**

### 128 **3.1 Model versus observations of atmospheric BC**

129 As part of the international exercise AEROCOM (see <http://nansen.ipsl.jussieu.fr/AEROCOM/>),  
130 Koch et al (2009) analysed the capabilities of global aerosol models to simulate BC. Like most  
131 other models, INCA was found to generally underestimate the aerosol absorption optical depth. A  
132 different behaviour was found for the modelled BC surface concentration, which is generally  
133 overestimated in Europe and underestimated in Asia. In particular, INCA using a coarse grid (96x95  
134 grid points, i.e. ~350 km) was found to underestimate by a factor of two on average the surface  
135 concentrations over entire Asia. Wang et al. (2014) reduced strongly this bias using a new BC  
136 inventory and performing the simulations with a stretched grid centred over Asia, corresponding to  
137 a resolution of ~50 km in this region. Only sparse observational data are available in the Himalayan  
138 region (Nair et al., 2013) making it difficult to evaluate the performance of global aerosol models in  
139 this region. To our knowledge, such models have only been validated in terms of aerosol deposition  
140 on snow in these areas (e.g. Kopacz et al., 2011).

141 To investigate the ability of our model to describe the atmospheric concentration of aerosols in high  
142 altitude areas of the Himalayas, we use here observations performed since 2006 at the Nepal  
143 Climate Observatory-Pyramid (NCO-P, 27.95°N, 86.82°E, 5079 m a.s.l., Bonasoni et al., 2010), in  
144 the region of Mount Everest. At NCO-P, BC is observed using a Multi-Angle Absorption  
145 Photometer (MAAP) providing measurements of the aerosol absorption coefficient that can be  
146 converted to Equivalent Black Carbon (EBC, Petzold et al., 2005, Marinoni et al., 2010). In the  
147 following, MAAP-derived EBC will be labelled BC for simplifying reading of the paper although  
148 we are well aware of denomination recommendations from Petzold et al. (2013). Table 1 shows  
149 observed and modelled atmospheric concentrations of BC at this site considering the year-  
150 dependent seasons as defined in Bonasoni et al. (2008). The order of magnitude and the annual  
151 cycle of the observed BC concentration are rather well reproduced by our model with minimum  
152 values occurring during the monsoon, higher values during the winter, and maximum values before  
153 and after the monsoon. However, the simulated BC maximum occurs in the post-monsoon period,  
154 whereas it is observed during the pre-monsoon period. The maximum daily average value observed  
155 from 2006 to 2007 reaches 2500 ng m<sup>-3</sup> whereas the simulated BC never exceeds 550 ng m<sup>-3</sup> during  
156 the same period. Such observed high values may be due to a thermally driven atmospheric  
157 circulation that cannot be represented in a model at 50 km horizontal resolution.

### 158 **3.2 Model versus observations of BC deposited on snow**

159 Simulated aerosol deposition and aerosol concentration in the snow are now compared to the  
160 information recorded in a shallow ice core extracted from the Mera Glacier (6376 m a.s.l, 27.7 °N,  
161 86.9 °E, Nepal), a high altitude site located 35 km south from the NCO-P site. This ice core was  
162 used to reconstruct both the evolution and the deposition flux of several proxies including BC and  
163 dust over the period 1998-2010 (Ginot et al., 2013). This 19.8 m core was sub-sampled with a mean  
164 resolution of 6.6 cm (the size of the samples varying between 4 and 16 cm), resulting in ~30  
165 samples per year. Note that BC is in that case determined using a Single Particle Soot Photometer  
166 (SP2) and corresponds to refractory BC (rBC) according to the recommendation from Petzold et al.  
167 (2013). Lim et al. (submitted) compared EC (from thermo-optical techniques) to rBC (SP2) in  
168 Himalayan snow. Overall, the EC/rBC is  $>1$  and often close to 3. This has to be accounted for in the  
169 comparisons, as we expect our BC simulations to be closer to EC than rBC observations, since they  
170 were performed with IPCC emission inventories mainly based on emission factors derived from  
171 thermal-optical methods (Lamarque et al., 2009, Petzold et al., 2013). In the following, we compare  
172 these ice core observations with our simulation focusing on BC and dust concentrations in the snow  
173 and their deposition fluxes. We do not expect our coarse-gridded model to reproduce the local BC  
174 observations in the snow, but we test if our model is able to reproduce the temporal variation of BC  
175 observed in snow. We use also this comparison to evaluate the order of magnitude of BC in snow.  
176 The altitude of the model grid cell containing the Mera Glacier reaches only 3000 m a.s.l., an  
177 altitude too low to simulate a continuous seasonal snow cover in winter/spring. Therefore, we used  
178 for our comparison the neighbouring grid cell located 50 km further north at an altitude high enough  
179 (5552 m a.s.l.) to conserve a continuous seasonal snow cover in the simulations. Here, the monsoon  
180 period is defined as JJAS, pre- and post-monsoon as AM and ON, and winter as DJFM. Due to  
181 strong winds eroding the snow at the surface (Wagon et al., 2013), the winter deposition fluxes  
182 could not be determined from the ice core (except for the last year of the period 1998-2010) and we  
183 grouped winter, pre- and post-monsoon into one period designed as inter-monsoon period.

184 Observed and simulated BC concentrations in the snow are reported in Table 2 showing very large  
185 discrepancies between modelled and observed BC concentrations: The modelled annual mean value  
186 ( $201 \mu\text{g kg}^{-1}$ ) is 60 times higher than the mean observed value ( $3.0 \mu\text{g kg}^{-1}$ ). Monsoon and inter-  
187 monsoon modelled values ( $285$  and  $28 \mu\text{g kg}^{-1}$ , respectively) are 30 times higher than the observed  
188 values ( $9.2$  and  $1.0 \mu\text{g kg}^{-1}$ , respectively). Note the winter difference is also very high (not shown  
189 because observed values are available only for a part of the last winter, the winter snow of the  
190 previous years was totally eroded by strong local winds), which explains the seasonal differences to  
191 be lower than the annual mean difference. Even with a difference of more than one order of

192 magnitude, the modelled and observed BC concentrations in the snow show a similar seasonal cycle  
193 with inter-monsoon values ten times larger than the monsoon values. The modelled annual BC  
194 deposition flux is 16 times higher than the flux deduced from the ice core (model:  $53 \text{ mg m}^{-2} \text{ yr}^{-1}$ ,  
195 observation:  $3.2 \text{ mg m}^{-2} \text{ yr}^{-1}$ ). Both in the model and in the ice core, the inter-monsoon period is  
196 characterised by high levels of BC deposition (model: 58 % of the total annual deposition,  
197 observation: 75 %), whereas this flux is lower during the monsoon period (model: 42 % of the total  
198 annual deposition, observation 25 %). Note that the BC deposition rates simulated in the grid cell  
199 really containing the Mera Glacier are 30% higher than those simulated in the grid cell that we used  
200 for our comparison. This difference is due to the altitude of this grid cell, lower by 2500 m than that  
201 of its neighbour, and therefore much more exposed to the transport of pollutants emitted at the  
202 foothills of the Himalayas. With a mean altitude of 3000 m, i.e.  $\sim 3400 \text{ m}$  lower than the real altitude  
203 of the drilling site, it would definitely be impossible to compare the observations with the values  
204 simulated in this grid cell. Nevertheless, our climate model has a too coarse resolution to simulate  
205 the local aerosol deposition and the final goal of such a comparison is to discuss the seasonal  
206 variations and the order of magnitude of the BC regionally deposited in snow-covered areas both in  
207 local observations and in regional simulations.

208 Modelled dust concentration in the snow reaches  $10.4 \text{ mg kg}^{-1}$ , close to the value observed in the ice  
209 core ( $10.1 \text{ mg kg}^{-1}$ ). Measurements of dust concentrations do not show any seasonal cycle whereas  
210 simulated concentrations are two times lower during the monsoon ( $5 \text{ mg kg}^{-1}$ ). Modelled and  
211 observed dust depositions are similar with an annual mean of  $10.1 \text{ g m}^{-2} \text{ yr}^{-1}$  in the ice core and  $6.4$   
212  $\text{g m}^{-2} \text{ yr}^{-1}$  in our simulation. However, observed dust deposition is three times larger during the  
213 monsoon, which is in contrast to our simulation with a flux slightly lower during the monsoon  
214 compared to the inter-monsoon flux. This may be due to a compensating effect since Ginot et al.  
215 (2013) showed that a large fraction of dust is emitted locally exhibiting a small scale process that is  
216 not represented in our coarse-gridded model.

217 We simulated an annual mean of snowfall that reaches  $83 \text{ mm}$  [water equivalent, w.eq.]  $\text{month}^{-1}$  on  
218 the model grid cell used for our comparison. At the ice core drilling site precipitation has not been  
219 measured, but Ginot et al. (2013) report an annual mean snow accumulation of  $94 \text{ mm w.eq. month}^{-1}$ ,  
220 in accordance with those measured by Wagnon et al. (2013) over the period 2007-2012 ranging  
221 between  $0.38$  and  $0.98 \text{ mm w.eq. month}^{-1}$ . The observed accumulation results from the snowfall  
222 diminished by sublimation, melting, and wind erosion (Wagnon et al., 2013), and cannot be directly  
223 compared with our modelled snowfall. Moreover, we cannot expect our model to accurately  
224 simulate the local accumulation since it remains relatively coarse-gridded even with a stretched  
225 grid. The similarity of order of magnitude of the observed snow accumulation and the modelled

226 snowfall is clearly a coincidence. Still, it indicates that the difference between modelled and  
227 observed concentration of aerosol in the snow, particularly marked for BC, cannot be explained by  
228 a difference in snow accumulation between model and observations.

### 229 3.3 How to explain the differences between simulations and observations?

230 Comparing large-scale outputs of a global climate model with local observations is challenging, in  
231 particular over a complex topography area, where models are not able to describe the high spatial  
232 variability of the atmospheric circulation and surface processes. Moreover, it is difficult to compare  
233 BC in snow modelled with a two-layer snow scheme at a resolution of 50 km with local  
234 observations. Still, such a comparison is essential to check the capabilities of the applied models  
235 and to analyse physical processes involved both at regional and local scales. In the following, we  
236 describe five points that can explain the differences between our simulations and available  
237 observational data:

238 (1) Due to its resolution, the grid cell of the model used here for comparison with the observations  
239 has an average altitude of 5552 m a.s.l., approximately 1000 m lower than the site where the ice  
240 core was drilled. Since we simulated a strong vertical gradient of the BC concentration in the  
241 atmosphere (not shown), the modelled rates of BC are more characteristic of the Himalayan valleys  
242 than high summits areas, which are less polluted since they reach the altitude of the free troposphere  
243 (Bonasoni et al., 2008). In the grid cell used for our comparison, the ratio between the BC  
244 atmospheric concentrations simulated at the surface (5552 m a.s.l.) and those modelled at 6500 m  
245 a.s.l. varies between 5 and 10 over the period 1998-2008. This assumption is consistent with the  
246 results from Kaspari et al. (2013), who reported BC concentrations in snow sampled at different  
247 altitudes between the Mera Col (6400 m) and the Mera La (5400 m). At the Mera La, located at an  
248 elevation similar to the model grid cell, they measured a mean BC concentration in snow of  $180 \mu\text{g}$   
249  $\text{m}^{-3}$  in the top three meters of snow with extreme values exceeding  $3500 \mu\text{g}\cdot\text{m}^{-3}$ . According to their  
250 study, snow is more polluted by a factor of 180 between their low altitude site (5400 m) and their  
251 high altitude site (6400 m). The high concentrations in the samples from low altitudes remain  
252 uncertain due to the sampling in exposed crevasses, which may have caused an artificial  
253 enrichment. Still, these measurements clearly indicate that snow sampled at low altitude is likely to  
254 be more polluted by one to two orders of magnitude in comparison to snow sampled at the top of  
255 the Himalayan Mountains. Our simulation provides values of BC in snow ranging from 50 to 500  
256  $\mu\text{g} \text{ m}^{-3}$  in the Nepalese Himalaya (Figure 3a). This range is therefore representative of BC  
257 concentrations in snow deposited over middle altitude areas (<6000 m), and not of those measured  
258 at high altitude sites (>6000m). Furthermore, melting and sublimation may accumulate BC in  
259 surface snow layers. These processes are more pronounced at low than at high altitudes possibly



260 contributing to a BC concentration difference between the upper and lower zone of the Mera  
261 Glacier. Our model takes into account these processes. As a result, we expect our model to  
262 reproduce the concentrations of BC in snow observed at 5500 m and not those measured at higher  
263 elevations on the Himalayan glaciers. Regarding the ice core of the Mera Glacier, the higher  
264 concentration of dust in comparison with that of BC suggests that dust is transported in the  
265 atmosphere via high-altitude pathways in contrast to BC that stays in the lower layers of the  
266 atmosphere. Similarly, Fadnavis et al. (2013) estimated from aerosol simulations and satellite  
267 observations that dust concentrations reach high and relatively homogeneous values below 6500 m  
268 (i.e. under ~450 hPa) in the Himalayan region. According to their study, the vertical gradient of  
269 aerosol in the atmosphere below 6500 m appears to be more pronounced for BC than for dust. Such  
270 a difference can explain that our model reproduces well the dust concentration in the snow observed  
271 at Mera Glacier whereas it gives higher values for BC in comparison with the observations because  
272 the altitude of the used model grid cell is 1000 m below the ice core drilling site.

273 (2) The BC concentration shows large vertical variations throughout the snowpack (Ming et al.,  
274 2009). The estimate of the BC surface concentration in the snow strongly depends on the snow  
275 thickness considered when analysing both observations and model outputs. In the model the surface  
276 snow layer corresponds to the top 8 mm Snow Water Equivalent (SWE, see Ménégoz et al., 2013a  
277 for model details) of the snowpack, whereas the vertical resolution of the Mera Glacier ice core is  
278 6.6 cm SWE. Since most of the BC deposited is accumulated in the top layer of the model,  
279 considering a thicker top layer in our simulation would result in lower concentrations. As an  
280 example, with a concentration of BC in the bottom snow layer equal to zero and considering a depth  
281 of 8 cm instead of 8 mm for the top layer corresponds to an artificial reduction of the surface  
282 concentration by a factor of 10. Ming et al. (2009) considered a 1-meter depth for the surface snow  
283 layer, for which they found a BC concentration of  $18 \mu\text{g kg}^{-1}$  for the East Rongbuk Glacier making  
284 it difficult to compare with our modelled value. In addition, the question about a possible flushing  
285 of BC particles through snow layers is still unresolved. Performing observations at 2000 m a.s.l. in  
286 the Northern United States under particularly high rates of snow melting, Conway et al. (1996)  
287 observed a diffusion of BC particles through the snow layers. Similar experiments performed in  
288 Spitsbergen by Aamaas et al. (2011) led to a contrary conclusion based on the observations that BC  
289 tends to stay at the surface of the snowpack even during melting conditions. We concluded that we  
290 could not implement a parameterisation describing such processes in our model due to the lack of  
291 consistent observations. Still, we considered in our model that all the deposited BC stays at the  
292 surface as long as there is no snowfall without any flushing of particles through the snow layers  
293 (Ménégoz et al., 2013a). This assumption could overestimate both the BC concentration in the top

294 snow layer and the magnitude of the BC effects on the snow cover and the climate. However, we  
295 assume this hypothesis to be realistic for snow covered areas located below 6000 m, since Kaspari  
296 et al. (2013) observed at the Mera La particularly polluted snow layers, with a BC concentration  
297 exceeding  $3500 \mu\text{g}\cdot\text{m}^{-3}$ , higher than the maximum that we simulated in the whole Himalayan region  
298 (Figure 3a). Neglecting the BC flushing through snow layers in our simulations can also partially  
299 explain why the ratio between modelled and observed BC concentrations in the snow is larger than  
300 that for the deposition fluxes (60 versus 16, see Table 2). However, this assumption does not impact  
301 the modelled flux of BC deposited at the surface modifying only its vertical distribution within the  
302 snowpack.

303 (3) The strong winds observed at the drilling site erode unknown amounts of both snow and aerosol  
304 in particular during the winter (Wagon et al., 2013, Ginot et al., 2013). This process is not  
305 simulated in our model.

306 (4) Measurements of BC concentrations, both in the atmosphere and in the snow, differ widely  
307 according to the method used to perform the observations (Petzold et al., 2013). Overall, different  
308 methodologies and sample treatment can lead to differences in BC up to a factor of 5 as reported by  
309 Lim et al. (submitted) and Kaspari et al. (2013). However, the differences between model and  
310 observations are too high to be explained alone by the difference in measuring techniques.

311 (5) BC deposition strongly varies both spatially and temporally. In an ice core from the East  
312 Rongbuk Glacier ( $28.03^\circ \text{N}$ ,  $86.96^\circ \text{E}$ , 6518 m a.s.l. Everest region), Kaspari et al. (2011) measured  
313 average rBC concentration in snow of  $0.7 \mu\text{g kg}^{-1}$  for the recent period using an SP2. They  
314 estimated the seasonal mean of rBC concentration in snow to reach a maximum close to  $10 \mu\text{g kg}^{-1}$   
315 during winter or spring and a background value during the monsoon around  $0.1 \mu\text{g kg}^{-1}$ . These  
316 values are similar to those measured at the Mera Peak (Table 2). Some kilometres farther at the  
317 Repula Col ( $28.02^\circ \text{N}$ ,  $86.96^\circ \text{E}$ , 6500 m a.s.l.) Ming et al. (2008) found a mean concentration of  
318 EC in snow around  $20 \mu\text{g kg}^{-1}$  using a thermo-optical technique with an opposite seasonal cycle  
319 compared to the seasonal cycle observed by Kaspari et al. (2011). In their observations, the  
320 maximal concentration of EC in snow occurs during the monsoon with values sometimes exceeding  
321  $50 \mu\text{g kg}^{-1}$ . Kaspari et al. (2011) estimated the dry deposition to be the main sinks of atmospheric  
322 BC in the Everest region, whereas Ming et al. (2008) estimated BC to be incorporated in snow  
323 mainly by wet deposition. In a further study, Ming et al. (2009) measured the BC concentration in  
324 the snow of different mountainous areas in Western China. On the South of the Tibetan Plateau,  
325 they measured vertical profiles of BC in snow with a resolution of 5 cm finding concentrations  
326 ranging between 22 and  $600 \mu\text{g kg}^{-1}$  closer to our modelled values. Modelled results are similar to  
327 highest BC mixing ratios derived by thermo-optical methods. As explained previously, emission

328 factors used in inventories made for climate models are often derived from EC measurements.  
329 Therefore, a comparison with thermo-optical methods may be more appropriate than a comparison  
330 with SP2-derived values. Nevertheless, it is clear that individual measurement sites may not be  
331 representative of a model grid-box.

### 332 **3.4 Estimating wet and dry aerosol deposition in the HKKH**

333 We now examine the relative importance of wet versus dry deposition of BC in the HKKH. We  
334 simulate high amounts of BC and dust wet deposition in the region of the Mera Glacier, while dry  
335 deposition represents locally only 11 % of the total simulated deposition (Table 2). Many previous  
336 published studies have focused only on the role of dry deposition of BC on the snow albedo in the  
337 Himalayas. As an example, Yasunari et al. (2010, 2013) estimated dry deposition velocities and  
338 corresponding snow albedo variations at the NCO-P site. They forced their simulations with the  
339 atmospheric measurements of Bonasoni et al. (2010), who pointed out the high spring atmospheric  
340 BC concentration. Yasunari et al. (2010, 2013) estimated the BC deposition from March to May  
341 2006 to reach 900-1300  $\mu\text{g m}^{-2}$  inducing a concentration of BC in snow ranging between 26 to 68  
342  $\mu\text{g kg}^{-1}$ . Considering only dry deposition in our simulations leads to a similar deposition flux (480  
343  $\mu\text{g m}^{-2}$  for the same period) and to an equivalent BC concentration in snow of 43  $\mu\text{g kg}^{-1}$ . However,  
344 the model indicates that large amounts of BC in snow originate from wet deposition, both during  
345 monsoon and inter-monsoon periods. We suggest, therefore, that the maximum of BC observed in  
346 the Mera ice core is largely due to wet deposition and not only to dry deposition as often suggested  
347 (Yasunari et al., 2010, 2013). This assumption is reinforced by Ming et al. (2008, 2009) who found  
348 that the BC concentration in snow reaches a maximum during the monsoon period exceeding values  
349 of 50  $\mu\text{g kg}^{-1}$ . Reconstructing the atmospheric concentration of BC, they considered that most of the  
350 BC in the ice core originated from wet deposition. In our simulation, dry deposition reaches strong  
351 values during the inter-monsoon period on the southern slopes of the Himalayas (Figure 2a). During  
352 this dry period, simulated wet deposition plays a minor role over the Indian subcontinent, but there  
353 is still some precipitation in particular over the Western and the Eastern part of the Himalayas  
354 inducing a large amount of wet deposition in these regions (Figure 2b). During the monsoon  
355 simulated dry deposition is weaker than during the inter-monsoon period (Figure 2c). During the  
356 same period, due to high precipitation rates wet deposition is enhanced over the entire Indian  
357 subcontinent (Figure 2d), in particular in the Indo-Gangetic Plain, where very high amounts of BC  
358 are emitted (not shown). The atmospheric residence time of BC (i.e. the ratio between the  
359 atmospheric BC concentration and the BC deposition rate) is, therefore, limited, but still long  
360 enough to allow a significant transport of BC towards the Himalayas in our simulation leading to  
361 large wet deposition rates over the HKKH and the Tibetan Plateau. Finally, we assume that the

362 diminution of the atmospheric BC concentration observed at the NCO-P observatory is not due to a  
363 reduction of the aerosol transport from polluted areas in the south, but rather to a decrease of the  
364 atmospheric residence time of BC. Even with significant amounts of BC deposited on snow during  
365 the monsoon, the BC concentration in snow decreases as snowfall occurring at the same time  
366 strongly dilutes the aerosol concentration in the snow. The pronounced spatial heterogeneity of  
367 precipitation in the Himalayas (Ménégoz et al., 2013b) certainly induces large spatial variations of  
368 BC wet deposition, which may explain parts of the difference between the observations of BC in  
369 snow performed by Ginot et al. (2013), Ming et al. (2008), and Kaspari et al. (2011, 2013). There is  
370 a strong need for more observations to quantify accurately both the BC wet deposition and the BC  
371 concentration in the snow at different altitudes. Still, our model and most of the observations show  
372 BC in snow to reach a maximum in spring, which is characterized by an increase of both dry and  
373 wet BC deposition concomitant with low levels of snowfall and increasing levels of sublimation  
374 and melting. In the absence of flushing, all these factors enhance the BC concentration at the snow  
375 surface. In our simulations, the spring maximum concentration of BC in the surface snow layer (i.e.  
376 the upper 8 mm SWE) reaches spatially highly variable values ranging from 50 to 500  $\mu\text{g kg}^{-1}$   
377 (Figure 3a).

#### 378 **4 BC deposition impact on snow cover and surface energy balance**

379 To quantify the effect of BC deposition on Himalayan snow, also called the “snow darkening  
380 effect” (Bond et al., 2013), we performed two simulations with and without BC in snow. Note that  
381 the atmospheric effects of BC are taken into account in both simulations. **The snow albedo changes  
382 induced by dust deposition, well known to minimize the forcing of BC in snow (e.g. Ginot et al.,  
383 2013, Kaspari et al. 2013), are also taken into account in both simulations.** We analysed the  
384 modifications induced by the forcing of BC deposited on snow including the so-called “rapid  
385 adjustments” (IPCC, 2013), but neglecting large parts of longer-timescale feedbacks since the winds  
386 were nudged toward the reanalysis of the ECMWF in the two simulations (see Section 2.1). Note  
387 that snow aging processes including snow grain size growth is taken into account in our model. This  
388 rapid adjustment significantly enhances the BC forcing in snow. We computed the difference  
389 between the two simulations to estimate the snow cover duration change induced by BC in snow.  
390 The snow cover duration is defined as the number of days per year with a SWE higher than 0.01  
391 mm. In parts of Nepal, the annual mean of the snow cover duration is reduced between one and five  
392 days (Figure 3b). However, due to the relatively small surface covered by seasonal snow in Nepal  
393 this decrease remains limited to a small area. In contrast, larger areas in the Karakoram and in the  
394 Western and Eastern Himalayas undergo a decrease of one to eight days per year of the snow cover  
395 duration due to the snow darkening effect. These variations are statistically significant at a 95 %

396 level according to a two-sample t-test. As stated in Section 3.3, we simulated relatively high BC  
397 concentrations in snow, representative of those observed at intermediate altitudes (< 6000 m). The  
398 snow cover variations that we simulated are therefore representative of these areas whereas the  
399 much lower values obtained by Ginot et al., (2013) and Kaspari et al. (2013) suggest the BC forcing  
400 to be weaker at higher altitude (> 6000 m). Maskey et al. (2011) pointed out that the areas located  
401 higher than 6000 m concern only 1% of the mountainous regions (> 3000 m) in Nepal. Thus, the  
402 largest areal extent of snow cover area lies in the elevation zone between 3000 m and 6000 m,  
403 where snow is more likely to be polluted stronger. Our simulation is representative of intermediate  
404 altitude areas, where snow cover is not continuous from one year to another, and cannot be used to  
405 assess the “snow darkening effect” at regions higher than 6000m. The simulated annual mean of  
406 snow cover duration is quite low over the Tibetan Plateau (Figure 1d). Therefore, its sensitivity to  
407 aerosol deposition remains limited and we modelled no significant variation of the snow cover  
408 duration in this region. We assume that previous studies based on coarse-gridded models (e.g.,  
409 Flanner et al., 2007; Menon et al., 2010, Qian et al., 2011) strongly overestimate the forcing of BC  
410 in the Tibetan Plateau. Such overestimation is not due to the representation of the BC forcing itself,  
411 but rather to an overestimation of the snow cover extent in this region, which cannot be well  
412 simulated with a coarse resolution model. Two reasons explain that snow cover duration is not  
413 reduced by BC deposition over the Tibetan Plateau: (i) The Tibetan Plateau is snow-covered during  
414 longer periods only during the winter (DJFM, Ménégos et al., 2013b), when the solar radiation is  
415 low. Moreover, the aerosol transport from the Indian plains is limited during this period because of  
416 low temperatures, a highly stable atmosphere, and very strong Westerlies (Ménégos et al., 2013b).  
417 (ii) During spring, summer, and fall, the Tibetan Plateau is more affected by BC deposition, but  
418 snow covers the surface only during brief periods that are too short to allow post-depositional  
419 processes to accumulate BC at the surface of the snow cover.

420 In the regions where the BC deposition on snow induces a decrease of the snow cover duration, we  
421 found an increase in the average annual net surface solar radiation that varied between 1 and 3 W m<sup>-2</sup>  
422 for the period 1998-2008 (Figure 3c). This surface net radiative forcing is of the same order of  
423 magnitude as previous local estimates. Based on a simple radiative model transfer, Ginot et al.  
424 (2013) found the BC measured in their ice core to increase the net surface radiative balance by 2 W  
425 m<sup>-2</sup>. Ming et al. (2008) reported a mean surface forcing for the BC deposited on snow of 1 to 1.5 W  
426 m<sup>-2</sup> at the end of the 20th century and Kaspari et al. (2011) calculated a forcing of 0.5 W m<sup>-2</sup> over  
427 the last decades. The surface radiative forcing estimated in our study is necessarily higher as these  
428 last studies because it takes also into account rapid adjustments (in particular the faster growth of  
429 the snow grain size if the snow becomes warmer in the presence of BC). In addition, the number of

430 days when the surface albedo has values corresponding to snow-free surfaces instead of values of  
431 snowy surfaces is increased when BC in the snow is taken into account for the snow albedo  
432 computation. The forcing estimated in Figure 3c includes also this excess of solar energy absorbed  
433 by the surface when it is free of snow. Finally, the surface forcing of BC in snow that we estimated  
434 here includes the BC particles itself with all the rapid adjustments, but excludes parts of slow  
435 feedbacks that are associated with modifications of the atmospheric circulation, the hydrological  
436 cycle, and changes in sea surface temperature. Further sensitivity studies may be performed with  
437 our model to evaluate separately these processes as done by Menon et al. (2010) with a coarse-  
438 gridded GCM. In our simulation, the BC deposition on snow results in an increase of the mean 2m  
439 atmospheric temperature over 1998-2008 ranging from 0.05 to 0.3 °C (Figure 3d). This temperature  
440 modification is statistically significant according to a two-sample t-test (not shown). It represents an  
441 upper estimation because we modelled BC concentrations in the snow typical for especially  
442 polluted areas (Kaspari et al., 2013).

## 443 **5 Conclusion**

444 We applied a coupled climate-chemistry model to evaluate the impact of BC deposition on snow  
445 cover in the HKKH from 1998 to 2008 through snow albedo variations. When compared to satellite  
446 observations, the snow cover simulated with coarse-gridded models appears particularly biased by  
447 the absence of the representation of the complex topography. Simulating atmospheric circulation  
448 and surface energy balance with a finer resolution allows a more realistic representation of the snow  
449 cover duration. Even with some differences induced by local atmospheric processes not described  
450 by our large-scale model, this one reproduces the seasonal variations of the atmospheric BC  
451 concentrations observed in the Mount Everest region with maximum values occurring in the post  
452 and pre-monsoon period. Estimating the BC concentration in snow is more difficult due to the high  
453 spatial variability of dry and wet deposition and to the complex vertical distribution of BC in the  
454 snowpack. Our model simulates a BC dry deposition flux in accordance with previous local  
455 analysis. However, we purport that wet deposition brings also large amounts of BC to the  
456 Himalayan snow. This wet deposition does not increase directly the concentration of BC in the  
457 snow, in particular during the monsoon, because snowfall also brings fresh snow at the surface that  
458 is cleaner than the old surface layer, highly concentrated in BC. However, it plays a significant role  
459 in reducing the snow albedo: when periods of sunny days occur after snowfall events, melting and  
460 sublimation dramatically concentrate BC at the surface. Such events likely occur in spring during  
461 the progressive onset of the monsoon (Mölg et al., 2012) and occasionally in summer. Then, the  
462 atmosphere is quickly cleaned by higher rates of wet deposition causing significant levels of BC in  
463 the snow. More observations are needed to estimate the actual rates of BC wet deposition in this

464 region. In addition, field campaigns dedicated to observe both the BC altitudinal gradients and the  
465 BC vertical profile throughout the snowpack are helpful to improve our understanding of the snow  
466 darkening effect: Kaspari et al. (2013) observed a highly variable BC concentration in snow  
467 sampled on the Mera Glacier, as they measured concentrations varying from two orders of  
468 magnitude between the upper and lower parts of the glacier. According to their study, the BC  
469 concentrations varied from  $\sim 10 \mu\text{g kg}^{-1}$  in a snow pit at (6400 m a.s.l.) to thousands  $\mu\text{g kg}^{-1}$  in a  
470 vertical profile sampled at 5400 m a.s.l. in a crevasse. It remains difficult to validate the ability of  
471 coarse gridded models to simulate the BC concentration in the snow as it depends on several  
472 parameters including the snow depth considered both in simulations and observations. Further  
473 observations of BC in snow could help to force models with realistic vertical profiles of BC in  
474 snow. Nevertheless, we assume that our study is based on BC concentrations typical for seasonal  
475 snow cover at middle elevation areas ( $< 6000$  m a.s.l.), which are not representative of permanently  
476 snow covered areas located at high elevations ( $> 6000$  m a.s.l.). We estimate that the BC deposited  
477 on the mountains of the HKKH decreases the snow cover duration between one and eight days per  
478 year. We found this anthropogenic forcing to have a limited impact on the snow cover of the  
479 Tibetan Plateau, a dry region only partially snow covered even during the monsoon-accumulation  
480 period. Considering the BC forcing with rapid adjustments, part of the slow feedbacks, and in  
481 particular the removal of snow during some days per year, we found the surface of mountainous  
482 region of the HKKH to absorb an excess of 1 to 3  $\text{W m}^{-2}$ . This forcing cannot be considered as  
483 representative of glaciated areas since these are never free of snow except for debris-covered  
484 glaciers. Further climate simulations based on higher spatial resolution allowing the representation  
485 of permanent snow cover on current glaciated areas could be used to simulate the forcing of BC  
486 over glaciers. Finally, we estimate that the BC deposition on the Himalayan snow increased the  
487 annual mean 2m-atmospheric temperature by 0.05 to 0.3  $^{\circ}\text{C}$  between 1998 and 2008.

488

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646

647 Table 1: Summary of atmospheric concentrations of BC observed at the Nepal Climate  
 648 Observatory-Pyramid (NCO-P, 27.95°N, 86.82°E, 5079 m a.s.l., from March 2006 to February  
 649 2008, Bonasoni et al., 2010) and modelled with LMDZ-INCA. The corresponding grid cell of the  
 650 model is located at 4480 m a.s.l. and seasons are defined as in Table 1 of Bonasoni et al. (2010).

[BC] (ng.m <sup>-3</sup> )	Pre-monsoon	Monsoon	Post- monsoon	Winter	Maximum
Observations	316	50	135	119	2500
<b>Model</b>	<b>206</b>	<b>76</b>	<b>224</b>	<b>138</b>	<b>550</b>

651

652 Table 2: BC and dust concentration in the snow, and BC and dust deposition reconstructed from an  
 653 ice core drilled at 27.7° N, 86.9° E, 6376 m a.s.l. (Ginot et al., 2013); All percentages are computed  
 654 from the annual deposition values. The model grid cell is located 50 km northward, where the  
 655 altitude of the model surface is high enough to allow a continuous seasonal snow cover (28.0° N,  
 656 86.9° E, 5552 m a.s.l). Ginot et al (2013) computed the annual mean aerosol concentration by  
 657 multiplying the seasonal snow accumulation by the seasonal aerosol concentration. For the entire  
 658 period 1998-2008, they used the winter concentration of the last year, since it was the only one  
 659 available, the others being eroded by winds. Inter-monsoon modelled concentrations include winter  
 660 values. Annual modelled values are the temporal average of the aerosol concentration in the top  
 661 snow layer of the model (i.e. a constant surface depth of 8 mm snow water).

		Annual	Inter-monsoon	Monsoon
BC concentration ( $\mu\text{g kg}^{-1}$ )	Observation	3.0	9.2	1.0
	<b>Model</b>	<b>201</b>	<b>285</b>	<b>28</b>
BC deposition ( $\text{mg m}^{-2} \text{ yr}^{-1}$ )	Observation	3.2	75 %	25 %
	<b>Model</b>	<b>53</b> <b>(11% dry)</b>	<b>58 %</b> <b>(8% dry; 50% wet)</b>	<b>42 %</b> <b>(3% dry, 39% wet)</b>
Dust concentration ( $\text{mg kg}^{-1}$ )	Observation	10.1	11.1	10.1
	<b>Model</b>	<b>10.4</b>	<b>13</b>	<b>5</b>
Dust deposition ( $\text{g m}^{-2} \text{ yr}^{-1}$ )	Observation	10.1	28 %	72 %
	<b>Model</b>	<b>6.4</b> <b>(10% dry)</b>	<b>60 %</b> <b>(6% dry, 54% wet)</b>	<b>40 %</b> <b>(4% dry, 36% wet)</b>

662 **Figure captions :**

663 Figure 1: Annual mean snow cover duration, days per year) averaged over 1998-2008: (a) LMDZ  
664 coarse-gridded simulation (~350 km). (b) LMDZ simulation based on a stretched grid (~50 km)  
665 interpolated on the coarse grid of LMDZ. (c) Satellite observation interpolated on the coarse grid of  
666 LMDZ. (d) LMDZ simulation based on a stretched grid (~50 km). (e) Satellite observation  
667 interpolated on the LMDZ stretched grid. LMDZ simulations were performed with the BC effect on  
668 snow albedo. IMS satellite observations are from NSIDC (2008).

669 Figure 2: BC deposition ( $\text{mg m}^{-2} \text{ month}^{-1}$ ) modelled over the Indian subcontinent averaged over  
670 1998-2008. (a) Inter-monsoon (ONDJFMAM) dry-deposited BC. (b) Inter-monsoon wet-deposited  
671 BC. (c) Monsoon (JJAS) dry-deposited BC. (d) Monsoon wet-deposited BC. LMDZ simulation  
672 based on a stretched grid reaching a resolution of ~50 km over the Himalayas.

673 Figure 3: (a) Simulated spring BC mixing ratio ( $\mu\text{g kg}^{-1}$ ) in the surface snow layer (8 mm SWE). (b)  
674 Difference in annual mean of snow cover duration (days per year) between two simulations  
675 performed with and without the snow albedo variations induced by BC deposition. Areas with  
676 statistically significant differences, according to a two-sample t test, are red-contoured. (c) Same  
677 difference but for annual mean net surface solar radiation ( $\text{W m}^{-2}$ ). (d) Same difference but for  
678 annual mean 2m atmospheric temperature ( $^{\circ}\text{C}$ ). LMDZ simulations are based on a stretched grid  
679 reaching a resolution of ~50 km over the Himalayas. Wind fields have been nudged toward the  
680 ERA-Interim reanalysis (see details in Section 2.1).