Black carbon in snow in the upper Himalayan Khumbu Valley, Nepal: Observations and modeling of the impact on snow albedo, melting, and radiative forcing

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Abstract

Black carbon (BC) in the snow in the Himalayas has recently attracted considerable interest due to its impact on snow albedo, snow and glacier melting, regional climate and water resources. A single particle soot photometer (SP2) instrument was used to measure refractory BC (rBC) in a series of surface snow samples collected in the upper Khumbu Valley in Nepal between November 2009 and February 2012. The obtained time series indicates annual cycles with maximum concentration before the onset of the monsoon season and fast decreases in

rBC during the monsoon period. Measured concentrations ranged from a few ppb up to 70 ppb rBC. However, due to the handling of the samples the measured concentrations possess rather large uncertainties. Detailed modeling of the snowpack including the measured range and an estimated upper limit of rBC concentrations was performed to study the role of BC in the seasonal snowpack. Simulations were performed for three winter seasons with the snowpack model Crocus including a detailed description of the radiative transfer inside the snowpack. While the standard Crocus model strongly overestimates the height and the duration of the seasonal snowpack, a better calculation of the snow albedo with the new radiative transfer scheme enhanced the representation of the snow. However, the period with snow on the ground neglecting BC in the snow was still over-estimated between 37 and 66 days, which was further diminished by 8 to 15 % and more than 40 % in the presence of 100 or 300 ppb of BC. Compared to snow without BC the albedo is on average reduced by 0.027 and 0.060 in the presence of 100 and 300 ppb BC. While the impact of increasing BC in the snow on the albedo was largest for clean snow, the impact on the local radiative forcing is the opposite. Here, increasing BC caused an even larger impact at higher BC concentrations. This effect is related to an accelerated melting of the snowpack caused by a more efficient metamorphism of the snow due to an increasing size of the snow grains with increasing BC concentrations. The melting of the winter snowpack was shifted by 3 to 10 days and 17 to 27 days during the three winter seasons in the presence of 100 and 300 ppb BC compared to clean snow, while the simulated annual local radiative forcing corresponds to 3 to 4.5 and 10.5 to 13.0 W m^{-2}. An increased sublimation or evaporation of the snow reduces the simulated radiative forcing leading to a net forcing that is lower by 0.5 to 1.5 W m^{-2}, while the addition of 10 ppm dust causes an increase of the radiative forcing between 2.5 and 3 W m^{-2}. According to the simulations 7.5 ppm of dust has an effect equivalent to 100 ppb of BC concerning the impact on the melting of the snowpack and the local radiative forcing.

1   Introduction

Black carbon (BC) constitutes the most important light-absorbing aerosol in the atmosphere, where it contributes to the warming of the atmosphere (Bond et al., 2013). It further affects cloud formation either acting as cloud nuclei or increasing the evaporation rates in cloudy layers. It is formed during incomplete combustion processes and mainly emitted due to natural and anthropogenic sources like biomass burning or fossil fuel and biofuel combustion

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(Bond et al., 2013). If BC is incorporated in snow, it can lead to further warming due to its impact on the albedo of snow and ice causing an accelerated melting (e.g. Hansen and Nazarenko, 2004; Flanner et al., 2007; Ménégoz et al., 2014). BC is also a strong pollutant, dangerous for human health and the environment, and is considered as an important short-lived climate forcer. Therefore, reductions in BC emissions can potentially lead to a fast climate response, in particular for regions where elevated BC concentrations are observed. The global annual climate forcing of BC in the atmosphere and in the snow remains uncertain with the most recent estimates ranging from +0.64 ($\pm$ 0.4) to +1.1 W m^{-2} (with a 90% uncertainty range from +0.17 to 2.1 W m^{-2}) (Stocker et al., 2013; Bond et al., 2013).

Since regional warming due to BC can be much stronger than the global average, the Himalayas have become of great interest. The extended cryosphere in the high altitude regions of the Himalayas including numerous glaciers (Kääb et al., 2012) and extended snow-covered regions (Ménégoz et al., 2013a) is expected to be especially vulnerable because of the vicinity of large anthropogenic BC sources on the Indian sub-continent or in South-East Asia and high radiation intensities. Xu et al. (2009) proposed that BC in snow contributes to the retreat of glaciers observed in parts of the Himalayas, while Ramanathan et al. (2007) suggested that BC transported to the Himalayas contributed to the melting of the snow. If BC has an impact on the glacier mass balance as well as on the timing of the run-off formation due to the snow melt in springtime, this would have implications on the hydrological cycle, water resources, hydropower generation, and agriculture in the downstream regions possibly affecting the living conditions of a population of more than a billion people (e.g. Immerzeel et al., 2010). Changes in the cryosphere may further modify sensible and latent heat fluxes affecting also the Asian monsoon (Lau and Kim, 2006; Qian et al., 2011).

Measurements at the Nepal Climate Observatory at Pyramid (NCO-P) carried out since 2006 at 5079 m a.s.l. have confirmed that aerosols including BC can effectively be transported from the source regions to the high altitude regions of the Himalayas (Bonasoni et al., 2010; Marinoni et al., 2010). Ice cores and surface snow samples from different locations in the Himalayas (Ming et al., 2008; Kaspari et al., 2011; Ginot et al., 2014) and on the Tibetan Plateau (Ming et al., 2009; Xu et al., 2009) demonstrated that BC and other absorbing impurities like dust are efficiently incorporated into the snow. Based on ice core data, reductions in the snow albedo were estimated (Ginot et al., 2014) assuming that the profile of the ice core concentration directly correspond to the evolution of the BC concentration in the

surface snow layer. Furthermore, the transport of BC from the source regions and its deposition to the snow was calculated using different global transport and chemistry models (Flanner et al., 2007; Kopacz et al., 2011; Ménégoz et al., 2014). These studies resulted in an estimated annual radiative forcing due to BC in snow between 7 and 12 W m^{-2} close to the Mt. Everest (Kopacz et al., 2011) and between 1 and 4 W m^{-2} in the snow-covered areas of the Himalayas (Ménégoz et al, 2014), while peak values in the monthly radiative forcing exceeded 15 to 25 W m^{-2} for some parts of the Tibetan plateau (Flanner et al., 2007; Kopacz et al., 2011; Qian et al., 2011). However, large uncertainties in the calculated radiative forcing remain because of the low spatial resolution of the used global models preventing a precise representation of the high altitude regions and the pronounced topography of the mountain range. These model limitations constrain a multitude of simulated processes including local-scale flow, transport of water vapor and aerosols, precipitation, and snow cover formation and melting (Ménégoz et al., 2013a), which are crucial in obtaining a correct radiative forcing for BC in snow. It has been demonstrated that these deficits can cause an overestimation of the snow cover on the Tibetan Plateau producing also a likely positive bias in the estimated radiative forcing for BC in snow in this region (Ménégoz et al., 2013a, 2014). Moreover, some models generate significantly higher BC in snow concentrations compared to the few available observations in the Himalayas and on the Tibetan Plateau (Flanner et al., 2007; Ménégoz et al., 2014). However, the limited BC in snow measurements make it difficult to get a reliable idea of the spatial distribution, seasonal cycle, and inter-annual variability of BC in snow in this vast and complex region greatly diminishing our capability to validate model results (Qian et al., 2011).

It is well known that BC is not the only absorbing impurity in the snow in the high altitude region of Tibet and the Himalayas. High concentrations of dust have been observed in the atmosphere (e.g. Carrico et al., 2003; Duchi et al., in press) and in ice cores (Thompson et al., 2000; Kaspari et al., 2011; Ginot et al., 2014). In the atmosphere, Duchi et al (in press) reported the frequency of dust transport events at NCO-P with a maximum during the pre-monsoon period causing on average a 10-fold increase of PM_{10} compared to days without identified dust events. While the absorption of solar radiation due to dust is much less efficient compared to BC, this is at least partly compensated by much higher concentrations. Ginot et al. (2014) found dust concentrations up to almost 70 ppm and an average concentration throughout all seasons around 10 ppm in an ice core from the Mera glacier, which is significantly higher than any observed concentration of BC or elemental carbon (EC)

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in snow in this region (Ming et al., 2008, 2009; Xu et al., 2006; Kaspari et al., 2011; Ginot et al., 2014).

The presence of absorbers in the snow has multiple impacts on the properties of the snow, which finally contribute to the radiative forcing (Painter et al., 2007; Flanner et al., 2007). The first order impact is related to the direct reduction of the snow albedo due to the incorporation of the absorbers in the snow. A second order impact is linked to the fact that the reduction of the albedo leads to a stronger warming of the snowpack compared to the clean snow causing a faster metamorphism (or snow aging) and, thus, a more efficient growing of the snow crystals. Since larger snow crystals lead to a smaller albedo, this effect leads to a further reduction of the albedo of the snowpack. Nevertheless, the forcing related to changes in the albedo remains small compared to the positive radiative forcing induced by the earlier exposition of the underlying soil caused by an accelerated melting of the warmer snow containing BC. To study in detail these multiple impacts of the absorbing impurities on the processes and properties of the snow a detailed physical snowpack model like Crocus with sufficient complexity is needed (Brun et al., 1989, 1992; Vionnet et al., 2012). Crocus is capable to calculate the internal energy budget of the snowpack, to resolve temperature gradients inside the snowpack, and to simulate the metamorphism of the snow. However, the standard model version does not allow considering absorbing impurities like BC or dust for the calculation of the albedo.

Here, we report multi-annual measurements of BC in surface snow sampled on the southern slopes of the Himalayas close to NCO-P. We compare the snow concentrations with simultaneous atmospheric measurements to investigate the role of wet and dry deposition. Moreover, we present the Crocus snowpack model with an upgraded radiative transfer scheme to study the impact of BC and dust in snow. Forced with three years of meteorological observations from the Pyramid International Laboratory (close to NCO-P) and with observed BC and dust concentrations, the model was used to study the impact of the two absorbing impurities on snow metamorphism, melting, and local radiative forcing. The model results including sensitivities of the melting and radiative forcing due to the presence of BC and/or dust in the snow are presented and compared to previous large-scale model studies.

2 Methods

2.1 Snow sampling

56 samples of surface snow were collected in Nepal during the period from 13 November 2009 to 29 February 2012 at three different locations in the Khumbu region south of the Mount Everest: At NCO-P (27.96° N, 86.81° E; 5079 m a.s.l.) and on the glaciers Changri Nup (27.98° N, 86.76° E; 5700 m a.s.l.) and Pokalde (27.93° N; 86.83° E; 5600 m a.s.l.) (Fig. 1). In most cases snow from the top layer (≤10 cm) was collected and transported to France. Using field notes, observed precipitation and snow height at NCO-P, 51 of the available snow samples were classified into fresh snow (i.e. snowfall within 24 h before sampling) and old snow. Five samples remained unclassified.

2.2 Snow sample analysis and handling

The snow samples were analyzed using a Single Particle Soot Photometer (SP2, Droplet Measurement Technologies, US) to determine refractory BC (rBC) particles. Details of the analytical procedure are described in Lim et al. (2014). The SP2 applies a laser-induced incandescence technique to measure the mass of individual rBC particles (Schwarz et al., 2006) independent of particle morphology and light-scattering coating materials (Moteki and Kondo, 2007, 2010). Each rBC particle passes through the laser beam intra-cavity, where it absorbs light, reaches a vaporization temperature, at which it incandesces, and emits visible thermal radiation proportional to the mass of the individual particles. The SP2 is highly sensitive to rBC particles, but much less to other absorbing particles like dust. It was calibrated with fullerene soot (Alfa Aesar Inc., USA), a standard BC material of known single particle mass aggregating primary particles with graphitic structure. A new nebulizer (APEX-Q, Elemental Scientific Inc., Omaha, USA) was used to increase the efficiency, with which the rBC particles in the snow were transferred to the gas phase. The rBC losses during aerosolization were determined using eight liquid Aquadag® standards resulting in an average efficiency of 56 %, which was applied to all here reported BC concentrations.

All snow samples melted during transport from the field sites to France. They were stored at < 5 °C until analysis in April 2012. Before analysis, the samples were sonicated for 15 minutes to minimize rBC losses on the container wall. Five selected samples were re-analyzed almost two years later to evaluate the rBC particle loss during long-term storage of the samples in

liquid form. The samples showed decreases in detected rBC concentrations between 0 and 80 %, which is probably related to particles attachment on the container wall and the agglomeration of particles. The loss during storage was not straightforward and seemed to be highly variable depending on storage time and rBC concentration. Therefore, all measured concentrations are presented here without any further correction for potential rBC losses during transport or storage and should be considered as minimum values.

2.3 Meteorological data and atmospheric BC

Meteorological parameters have been recorded at Pyramid International Laboratory close to NCO-P since 1994, radiation and snow depth since 2002. Moreover, continuous measurements of atmospheric BC concentrations have been performed at NCO-P using a Multi-Angle Absorption Photometer since 2006. Further details of the instrument set-up and the calculation of equivalent BC concentrations are described by Marinoni et al. (2010).

2.4 Modeling

Simulations were performed with the 1-dimensional multi-layer physical snowpack model Crocus (Brun et al., 1989, 1992; Vionnet et al., 2012), which explicitly solves the surface mass and energy budgets taking into account heat diffusion, transfer of radiation, densification, sublimation, condensation, melting, and liquid water percolation in the snow. The model is forced using meteorological data like air temperature, wind speed, relative humidity, precipitation quantity and phase, incoming direct and diffuse solar radiation, incoming long-wave radiation, and cloud cover. The simulated snowpack consists of multiple homogeneous horizontal layers, which are established according to snowfall events undergoing transformation related to a metamorphism scheme. It calculates physical properties of each modeled snow layer including thickness, temperature, density, liquid water content, snow type, grain size, and age. The model further computes budgets of the snowpack like total height, run-off, latent and sensible heat fluxes, and fluxes of infrared and short-wave radiation.

In the Crocus standard version the albedo is not prescribed, but parameterized using the snow grain size and age of only the uppermost layer of the snowpack (Vionnet et al., 2012). The albedo is subsequently applied to calculate the absorbed amount of incoming radiation, while

the penetration of the absorbed radiation is simulated using absorption coefficients estimated from the density and grain size of each snowpack layer.

The standard albedo parameterization does not offer the possibility to account for the presence of absorbing impurities. In order to be able to study the impact of BC and dust on the snowpack with Crocus, we implemented a physically-based radiative transfer scheme without using prescribed albedo values. We employed the theory of Wiscombe and Warren (1980) and Warren and Wiscombe (1980) based on a module previously used in the land surface scheme ORCHIDEE for simulations with the global model LMDZ (Krinner et al., 2006; Ménégoz et al., 2013b, 2014). Starting with a fixed soil albedo, the albedo for diffuse radiation is calculated at the top of the bottom snow layer using snow water equivalent (SWE), grain size, and BC and dust concentrations of this layer. The same procedure is applied for the overlying snow layers until the surface layer is reached. For the surface layer, the albedo for direct radiation is calculated taking into account the solar zenith angle. The albedo for diffuse and direct radiation was separately combined with the incoming direct and diffuse radiation to calculate the overall amount of absorbed radiation. Since the albedo calculation for the diffuse radiation delivers also absorption coefficients, these were used to calculate the amount of radiation energy absorbed in each snow layer assuming that within the top snow layer all direct radiation was transformed into diffuse radiation. We used the same optical properties for ice like Krinner et al. (2006). For BC we assumed a log-normal size distribution with a median number radius of 11.8 nm, a density of 1 g cm^{-3}, and a refractive index of m=1.75–0.45 i (Ménégoz et al., 2013b); for dust a log-normal size distribution was used with a median mass diameter of 2 μm and a refractive index according to its haematite content (Krinner et al., 2006). Using these typical, but fixed properties for BC and dust may lead to an underestimation of the impact of the aerosols on the simulated albedo mainly because the model only considers externally mixed aerosols (Flanner et al., 2012). Nevertheless, the derived BC mass absorption cross section of 7.6 m^2 g^{-1} at 545 nm corresponds to previously published values (Bond and Bergstrom, 2006; Flanner et al., 2007). Since the standard version of Crocus considers three different wavelength ranges for the albedo and the absorption coefficient, the values derived from the radiative transfer module were also averaged for the same bands from 300 to 800 nm, 800 nm to 1.5 μm, and 1.5 to 2.8 μm.

1 For our simulations we used observations covering the period August 2004 to July 2007

2 obtained at an altitude of 5050 m a.s.l. at Pyramid Laboratory to construct the needed forcing

3 data. Quality controlled 1-hour averages for temperature, wind speed, humidity, and radiation

4 were used without further correction. However, it is well known that the observed

5 precipitation significantly underestimates solid precipitation (Bonasoni et al., 2010; Shrestha

6 et al., 2012). Accordingly, the observed snow height shows for several instances strong

7 increases while no simultaneous precipitation was recorded as already described by Shrestha

8 et al. (2012). As a result, preliminary simulations with the standard and upgraded Crocus

9 model with the recorded precipitation did not lead to the built-up of a significant snowpack.

10 Therefore, a corrected precipitation data set based on the observed snow height was

11 constructed and employed for all further snowpack simulations. If the snow height showed an

12 increase while no precipitation was detected, the increase in snow height was transformed into

13 accumulation using a density of fresh snow of 0.08 g cm^{-3}. Using such a density led to a good

14 agreement of simulated and observed increases in the snowpack height during the 2004/05

15 winter season. Since the detector of the snowpack height showed regular fluctuations around

16 ±1 cm, only increases in height larger than 1 cm were considered (Shrestha et al., 2012). In

17 addition, the snow height sensor recorded several peaks with strong increases in height and

18 subsequent large decreases of several tens of cm within hours or days. These peaks were

19 removed after visual inspection of the time series. Finally, the phase of the precipitation was

20 estimated using observed air temperatures with only solid precipitation at $T_{air} < 0°C$, only

21 liquid precipitation at $T_{air} > +2°C$, and mixed phase precipitation with 50 % solid precipitation

22 in the remaining temperature range. A comparison of recorded and corrected time series of

23 precipitation is shown in the supplementary material (Figs. S1a to S1c). After applying the

24 corrections the estimated total annual precipitation corresponds to 491.7 mm (41 % solid

25 precipitation), 423.8 mm (55 % solid precipitation), and 454.8 mm (51 % solid precipitation)

26 for the years 2004/05, 2005/06, and 2006/07 compared to recorded total precipitation of only

27 360 mm (2004/05), 231 mm (2005/06), and 304 mm (2006/07). The corrected annual values

28 are in excellent agreement with an estimated multi-year average of the annual precipitation of

29 ~450 mm at NCO-P for the period 1994 to 2013 (Salerno et al., 2014).

30 In the Crocus simulations, the fraction of cloud cover is used to determine the contribution of

31 direct and diffuse radiation to the total incoming radiation. For the simulations, the cloud

32 cover fraction was set to 0 (= clear sky) if the ratio between observed and theoretical

33 incoming solar radiation was larger than 0.8, to 0.5 (= cloudy) if the ratio was between 0.2

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and 0.8, and to 1 (= overcast) if the ratio was below 0.2. Based on these derived cloud fractions, the observed incoming short-wave radiation was divided into direct and diffuse radiation using the same parameterizations as in the Crocus model. As a result the total incoming radiation for the forcing of the model corresponds exactly to the measured values. The estimated cloud cover only affects the distribution between direct and diffuse radiation, which has a slightly impact on the calculation of the albedo as described above.

3   Results and discussion

3.1   BC concentrations in surface snow

Observed rBC concentrations are highly variable and range from less than 0.1 to more than 70 ppb (Fig. 2). Calculated average and median concentrations using all samples correspond to 10 and 1.5 ppb. As described in the Methods section, the reported concentrations are potentially underestimating the real BC concentrations. Nevertheless, the increases in the reported concentrations during the dry seasons 2009/2010 and 2010/2011 are well beyond the uncertainty of the measurements, which can be as high as a factor of 5. Despite this uncertainty and the high variability, we conclude that the concentrations follow a seasonal cycle with low values in the post-monsoon and winter season and higher concentrations in the pre-monsoon culminating at maximum concentrations before the onset of the monsoon.

Regarding snow types, we obtained somewhat higher concentrations in the old snow samples (average 15 ppb, median 3 ppb) compared to the fresh snow samples with average and median concentrations of 5 and 1.3 ppb. However, these differences and also differences between the sampling sites remain questionable because they are small compared to the uncertainty of the measured rBC concentrations.

Lower rBC concentrations were measured by Kaspari et al. (2011) in an ice core from the East Rongbuk Glacier at 6500 m a.s.l. close to the Mt. Everest using the same analytical method. They found average concentrations of (0.7 ± 0.1) ppb for the period 1975 to 2000 and a maximum concentration of 32 ppb. However, the same uncertainty in the measured rBC concentrations as for our samples due to the sample handling applies to the data reported by Kaspari et al. (2011).

Finally, the concentrations reported here are lower compared to the results for EC obtained with the thermo-optical method for snow and ice core samples from the high altitude region of the Himalayas and the Tibetan Plateau. For example, Ming et al. (2008, 2009) reported EC concentrations between 2 and 981 ppb in surface snow samples from West China for the period 2004 to 2006 and average concentrations around 20 ppb in an ice core section covering 1995-2002 extracted from the East Rongbuk Glacier. Higher EC values compare well to a comparison of EC and rBC measurements using the same snow samples from Nepal leading to an average EC-to-rBC ratio of 3.4 (Lim et al., 2014). In contrast, comparable results were obtained by Xu et al. (2006), who reported a range of EC concentration between 4 and 80 ppb in surface and fresh snow samples collected between 2001 and 2004 on various glaciers on the Tibetan Plateau.

The seasonal cycle in the surface snow corresponds well to the rBC concentration profile measured in an ice core retrieved from the Mera glacier at 6376 m a.s.l. (Ginot et al., 2014). While minimum rBC concentrations are similar, maximum concentrations in the ice core remained smaller probably due to lower deposition at higher altitudes. For comparison, the results of the overlapping period in the ice core and surface snow samples are shown in Fig. 2. Low concentrations of 0.35 ppb were found in the surface snow corresponding to the November layer, which is absent in the rest of the ice core due to efficient erosion during the following winter season.

The surface snow samples as well as the Mera ice core reveal the impact of wet and dry deposition responsible for the incorporation of rBC into the snow and strong links with the seasonal cycle of precipitation and atmospheric BC concentrations as recorded at NCO-P (Figs. S1a to c, 2). It seems that wet deposition due to the accumulation of fresh snow leads to relatively small concentrations of rBC around 1 ppb. However, in the case of snowfall during the pre-monsoon season, when atmospheric BC concentrations are high, rBC concentrations in fresh snow can increase to more than 10 ppb. Additional dry deposition of rBC seems to have a relatively small impact during the winter period and old snow exposed at the surface contains relatively low rBC amounts. Maximum rBC concentrations are reached again in the pre-monsoon season potentially combining large inputs due to wet and dry deposition. Yasunari et al (2013) estimated BC concentrations in surface snow using deposition velocities calculated with meteorological measurements at NCO-P and atmospheric measurements of equivalent BC. Considering only dry deposition they obtained concentrations between 90 and

1 130 ppb in old snow for a continuous snowpack until end of May. The observed rBC maxima

2 are somewhat lower than these values, possibly because the seasonal snowpack at NCO-P

3 melts earlier and some of the BC is lost due to the handling of the samples.

3.2 Snowpack modeling: Standard vs. upgraded model

5 Although the Crocus model has so far been used in different alpine and polar regions (e.g.

6 Jacobi et al., 2010; Brun et al., 2011; Vionnet et al., 2012), it has to our knowledge never been

7 applied to simulate the seasonal snowpack in the Himalayas. Recently, a modified version of

8 the model was employed to simulate the snow on top of a debris-covered glacier in the

9 Khumbu Valley (Lejeune et al., 2013). To examine the performance of the two model

10 versions, we first compared the results of the standard Crocus model and the upgraded version

11 including the radiative transfer for the seasons 2004/05, 2005/06, and 2006/07 applying the

12 forcing data based on the observations at the Pyramid site. The simulated snowpack heights

13 for the season 2004/05 are shown in Fig. 3 (and for the seasons 2005/06 and 2006/07 in the

14 Supplementary material, Figs. S2a and b). In all three winter seasons, the standard Crocus

15 model largely overestimates the period with snow on the ground (Figs. 3, S2a and b). For

16 example, Crocus predicts the formation of a continuous snowpack starting on 14 October

17 2004 and lasting until 4 January 2005 due to several small snowfall events in October and

18 November. However, the snow height records and albedo measurements show that during this

19 period the fresh snow regularly melts within a day after precipitation. The onset of the

20 seasonal winter snowpack corresponding to the longest period with continuous snow on the

21 ground at the end of January 2005 is well represented by Crocus because the observed

22 snowpack heights are used to construct the precipitation time series. In contrast, observed

23 snowpack heights start to decrease mid-February 2005, interrupted only by additional

24 accumulation in mid-March, until the snow disappeared before the end of March 2005. In

25 contrast, the winter snowpack remains intact in the Crocus simulations until end of May 2005,

26 before it melts completely on 10 June 2005. In summary, while the observed total period with

27 snow on the ground (defined as an observed snow height > 2 cm) corresponds to 78 d, the

28 standard Crocus model predicts a period of 238 d with snow on the ground. The period with

29 snow on the ground is similarly overestimated by Crocus for the years 2005/06 and 2006/07

30 with +91 and +157 d compared to the observations (Figs. S2a and b).

31 The positive bias is strongly reduced using the upgraded Crocus model including the radiative

32 transfer even without considering any absorbing impurity. During these simulations the

snowpack shows a much stronger dynamic with faster drops in the snow height compared to the standard model. Moreover, fresh snow in the fall and early winter season is not conserved for more than 24 h (Figs. 3 and S2a) or melts in agreement with the observed snow heights (Fig. S2b). The simulated duration of the snow cover is reduced between 54 and 103 d compared to the standard Crocus model. Nevertheless, the period with snow on the ground is still overestimated by 57, 37, and 66 d for the years 2004/05, 2005/06, and 2006/07 relative to the observations.

The obvious reason for the different behavior of the standard and the upgraded model is related to the calculated albedo and the corresponding energy absorbed by the snowpack. Figure 3 shows a comparison of the simulated albedo together with observed albedo calculated from the ratio of the up- and down-welling shortwave radiation for 2004/05. Strongest differences between observed and simulated albedo concerning all model results are related to the overestimation of the simulated periods with snow compared to the observations without snow. Nevertheless, Fig. 3 also illustrates the differences in the simulated albedo of the two different model versions. In the standard model, the albedo rises with each precipitation event to values around 0.9 before it slowly decreases due to the albedo parameterization related to the aging of the snow. Since only the properties of the top snow layer are considered in the standard model, the simulated albedo is not affected by the thickness of the snowpack and the parameterized albedo is similar regardless of the snow height. In contrast, the effect of a thin snowpack is much better reproduced by the upgraded model including the radiative transfer inside the snowpack. Here, the SWE of each snowpack layer is an important variable and leads in the case of a thin snowpack to strongly reduced albedo values as can be seen in the cases of snowfall before December 2004 or after June 2005 (Fig. 3). During these events the simulated albedo remains between 0.2 and 0.7 causing a stronger absorption of the incoming solar radiation and, thus, a complete melting of the snow. In all model versions, the precipitation in late January leads to the formation of the seasonal winter snowpack (Fig. 2) with an albedo between 0.6 and 0.9. These albedo values of the fresh snow are relatively well reproduced by both model versions (Fig. 3). However, neither model captures the relatively strong decrease of the albedo to 0.3 until 10 March before a new snowfall event increases the observed albedo to more than 0.8. In both model versions the overestimation of the period with snow on the ground is directly linked to the positive bias in the simulated albedo. Similar results are obtained for the years 2005/06 and 2006/07 (Figs. S3a and b).

In summary, the standard Crocus model does not capture the dynamic of the snow albedo for the conditions at Pyramid especially in cases with light snowfall and quick melting of the snow early and late in the period 2004 to 2005 and also for the thin seasonal winter snowpack. Similarly, Shrestha et al. (2012) simulated a delayed melting of the snow and overestimated the springtime snow-covered area in the Dudhkoshi region with a 3-layer snow model. They also attributed a large part of the model bias to the used simplified albedo parameterization. Most of the features of the albedo are better represented with the upgraded Crocus model. As a result, the period with snow on the ground and in some cases also the maximum snow height is largely overestimated with the standard model. This positive bias is reduced if the radiative transfer inside the snowpack is considered in the model. This corresponds well to results from previous snow model comparisons indicating that the albedo parameterization is a crucial component for snow models (Etchevers et al., 2004).

Nevertheless, the overestimation of the albedo and the period with snow on the ground persists also in the model runs with the improved albedo parameterization. An important part of this bias is reduced in the presence of absorbing impurities (see below). However, the spatial variability of the meteorological as well as the snow conditions in the rugged terrain of the Himalayas cannot be captured by the point measurements used here to drive and validate the snow model. The atmospheric and snow observations at the field site may only represent localized conditions. We assume that the non-ideal conditions at the field site introduce additional variability that cannot be represented by the simulations. This variability as well as further uncertainties in the observations directly translates into errors in the snowpack simulations that can further explain the differences between simulations and observations.

3.3   Impact of BC on snow albedo

To study the impact of BC present in the snow further runs with different constant BC concentrations in the snow were performed. We selected concentrations of 100 and 300 ppb covering the range of the here reported maximum BC concentration including their uncertainty. Figure 4 shows the simulated albedo for the three different BC concentrations (0, 100, and 300 ppb) during the period 21 to 31 January 2005. During 21 to 23 January, several snowfall events led to the initial formation of the winter snowpack. As a result, the observed albedo increased from values below 0.2 on 21 January to more than 0.8 the following day and to even higher values on 23 January. It followed a period of seven days without further precipitation and slightly decreasing albedo values. Similar results are obtained in the

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simulations, during which a maximum albedo was reached early on 24 January with a subsequent decrease in the calculated albedo. Further fresh snow during the night from 30 to 31 January increased albedo values in the observations and simulations. Figure 4 shows that overall trends as well as several short-term features in the observations are well reproduced in the simulations, e.g. the diurnal cycle of the albedo with morning and evening maxima (26 and 27 January), the continuous decrease on 24 January, or the unusual behavior on 28 January. The model is also capable of simulating a positive feedback loop between albedo, snow temperature, and grain size in the presence of BC referred to as the first indirect effect (Painter et al., 2007). Initial conditions in the snowpack with different BC concentrations are very similar leading to almost indistinguishable albedo values on 22 January in all model runs. However, first small differences in the snowpack properties become apparent on 23 January (Tab. 1). While snow height, SWE, and average temperature are similar in all model runs, simulated snow temperatures in the top layer are slightly higher with BC present in the snow. For example, the presence of 100 and 300 ppb BC in the snow increase the temperature in the top 10 cm by 0.2 and 0.4 K. The simulated averaged diameter of the snow grains is still very similar in all model runs with average diameters around 300 μm and differences smaller than 5 μm. Nevertheless, they show already an increasing trend with increasing BC. On 30 January, this situation has changed with snow temperatures (average for the entire snowpack or for top 10 cm) that are at least 0.5 K higher in the presence of BC. During the same period, the simulated average grain diameter increased to 369, 386, and 400 μm, respectively, in the presence of 0, 100, and 300 ppb BC in the snow. These grain sizes still remain smaller than the average grain size of 418 to 475 μm retrieved by Negi and Kokhanovsky (2011) using satellite data for a 7-day old snowpack in the upper Himalaya, which evolved at snowpack temperatures below -20 °C.

Due to the faster growing snow grains, the simulated albedo values decrease faster in the presence of BC. The growing gap between the simulated albedo values with 100 and 300 ppb compared to snow without BC is shown in Fig. 4. On average, due to metamorphism the albedo of the pure snow decreases on average by ~0.004 per day between 24 and 31 January. In the presence of 100 and 300 ppb BC the simulated albedo decreases by additional ~0.003 and ~0.005 per day during the same period. The albedo differences in the presence of BC are partly compensated after the addition of fresh snow. The snowfall event on 30 January increases not only the absolute albedo values of the snowpack in all simulations, but the new

1   snow layer simultaneously reduces the gap in the albedo caused by the different properties of

2   the older, underlying snow.

3   On longer time scales of weeks or month, the presence of BC in the snow causes a general

4   reduction of the simulated albedo as shown in Fig. 5 for the simulated winter snowpack

5   between January and May 2005 and an earlier melting of the snowpack. In this particular

6   case, the small initial changes in the albedo shifts the melting date of the persistent snowpack

7   by 5 and 25 days in the presence of 100 or 300 ppb BC.

8   Although the differences in the simulated albedo with different BC concentrations increase

9   with the age of the snow as predicted by Warren and Wiscombe (1980), we attempt to

10   quantify the average impact of BC on the snow albedo for typical conditions at Pyramid. We

11   calculated averaged albedo values for several periods between 22 January 2005 and 30 March

12   2007 from simulations with different BC concentrations in the snow between 0 and 300 ppb.

13   Figure 6 shows the normalized albedo as differences of the averaged albedo at a certain BC

14   concentrations minus the averaged albedo of pure snow. All selected periods are characterized

15   by a continuous snowpack with a height of more than 10 cm in all simulations to exclude the

16   impact of melting snow on the albedo. We tested if the length of the selected period is

17   important and found that while the averaged albedo values are significantly higher during the

18   period 12 to 17 March 2006 compared to the period 12 March to 5 May 2006 the sensitivity

19   of the averaged albedo as a function of BC is essentially the same during both periods at least

20   for BC concentrations below 100 ppb (Fig. 6). A further comparison for the periods from 22

21   January to either 11 March 2005 or 8 April 2005 gave essentially the same values for the

22   absolute as well as normalized albedo (not shown). For all periods the relationship between

23   normalized albedo and BC is best described using quadratic polynomials with regression

24   coefficients R^2 between 0.989 and 0.998. The fit demonstrates the non-linear behavior of the

25   albedo with respect to the BC concentration in the snow, because adding BC to the snow

26   exerts a decreasing effect on the snow albedo with increasing BC in snow concentrations.

27   This behavior corresponds to the applied radiative transfer theory of Warren and Wiscombe

28   (1980) because the BC already present captures some of the solar radiation that the additional

29   BC otherwise would receive. Overall the albedo reductions remain small ranging from 0.012

30   to 0.034 (average 0.027) and 0.031 to 0.078 (average 0.060) for BC concentrations of 100 ppb

31   and 300 ppb. The changes are similar to the values of Yasunari et al. (2013) who estimated

32   reductions in snow albedo between 0.012 and 0.022 after the addition of 120 ppb BC.

16

1   The sensitivity of the albedo towards BC depends further on the season with the smallest

2   impact on the snowpack in December 2006 and the strongest in the March to May 2006

3   period. The seasonal dependence of the sensitivity is linked to the radiation intensity, which is

4   lowest in December and increases until June, and the positive feedback between BC, snow

5   temperature, grain size, and albedo as described above.

6   ## 3.4   Impact of BC and dust on snow melting

7   Although the overall impact of BC in the snow on the albedo remains limited, the impact on

8   the melting of the snow can be rather large in the Himalayas as demonstrated in several model

9   studies (e.g. Flanner et al., 2007; Ménégoz et al., 2014). In the presence of BC, the melting of

10 the winter snowpack (corresponding to the longest period with a simulated continuous

11 snowpack of a height of > 2 cm) is shifted to early dates compared to the simulations without

12 BC. This shift corresponds to 3 to 10 days in the presence of 100 ppb BC and increases to 17

13 to 27 days with 300 ppb BC for the three simulated years (Fig. 7). The relationship between

14 the melting date and the BC concentrations is not always linear and depends for example on

15 the timing of the precipitation events during springtime. If the winter snowpack does not

16 persist until these events, a fast shift in the melting date is observed. One example is the shift

17 of 5 days of the melting in the season 2006/07 if the BC is increased from 80 to 100 ppb.

18 Besides the impact of the meteorological conditions, the number of days with snow on the

19 ground steadily decreases with increasing BC concentrations. While this decrease shows a

20 relatively large inter-annual variability, the overall trend is similar in all three years of

21 simulations with a stronger impact of an incremental increase of BC at higher concentrations

22 compared to lower concentrations in the snow. This behavior is, thus, in contrast to the direct

23 effect of BC on the snow albedo, which is strongest at low concentrations and becomes

24 weaker at higher concentrations (Fig. 6).

25 Although we did not measure the dust concentration in the surface snow samples, we can

26 assume that dust was also present like previously observed in ice cores from the Himalayas

27 (Thompson et al., 2000; Kaspari et al., 2011; Ginot et al., 2014). To study the impact of dust,

28 we performed calculations with a constant dust concentration of 10 ppm corresponding to the

29 average observed in the Mera ice core (Ginot et al., 2014) and BC concentrations varying

30 between 0 and 150 ppb and additional calculations without BC, but with dust concentrations

31 up to 15 ppm. In all simulations, the addition of absorbing impurities like BC and dust leads

32 to a reduction of the snow-covered periods. On average the snow-covered period is reduced

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1   by $5.6 \cdot 10^{-2}$ days (ppb BC)$^{-1}$ and $7.6 \cdot 10^{-4}$ days (ppb dust)$^{-1}$. The impact of the addition of BC

2   increases strongly in the presence of 10 ppm dust compared to pure snow because in these

3   simulations the reduction is enhanced to $8.6 \cdot 10^{-2}$ days (ppb BC)$^{-1}$. The reduction in the

4   snowpack duration is on average 50 % stronger compared to the simulations with only BC.

5   This behavior is similar to the acceleration of the melting of the snow at higher BC

6   concentrations. A linear regression using only the results for the simulations with dust = 0 and

7   80 ppb $\leq$ BC $\leq$ 250 ppb in Fig. 7 leads to a reduction of $7.9 \cdot 10^{-2}$ days (ppb BC)$^{-1}$ and is, thus,

8   similar to the impact obtained with a constant dust concentration of 10 ppm. Obviously, the

9   influence of the two different absorbing impurities in the model is comparable and exerts the

10 same processes and modifications of the snowpack. As a result, in general 100 ppb BC and

11 7.5 ppm dust can be regarded as equivalent considering the melting of the snowpack. This

12 relationship depends of course on the optical properties of the BC and dust used in the

13 simulations and can vary since the optical properties of dust are quite variable depending on

14 the chemical composition.

3.5   Radiative and net forcing

16 The reduction of the snow albedo and the earlier melting of the snowpack leads to a radiative

17 forcing since a larger proportion of the incoming radiation is absorbed at the Earth surface.

18 We calculated the radiative forcing using the observed incoming short-wave radiation and the

19 simulated albedo of the snowpack. In the absence of snow, we used a soil albedo of 0.15

20 corresponding to the observed wintertime albedo without snow on the ground (Fig. 3). All

21 values for the radiative forcing are calculated as the difference in absorbed shortwave

22 radiation with and without absorbing impurities in the snow. Since the radiative forcing can

23 partly be compensated by latent and sensible heat fluxes due to an increased sublimation or

24 evaporation of the snow, a net forcing is calculated after considering these fluxes between the

25 snow and the atmosphere. However, no further feedback mechanisms between the snow and

26 the atmosphere are included because all simulations were driven by the same meteorological

27 data sets.

28 The calculated radiative forcing (Fig. 7) as well as the net forcing (Supplement Fig. S4)

29 shows similar, but opposite trends as the reduction in snow-covered periods. A reduction of

30 the snow-covered period leads to stronger absorption due to the longer exposition of the

31 underlying soil causing an increased radiative forcing. This effect becomes obvious in the

32 seasonal cycle of the simulated forcing with 100 ppb BC in the snow shown as an example in

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Fig. 8. The maximum monthly radiative and net forcing is simulated for the end of the snow-covered period (i.e. May in 2004/05 and 2005/06 and April in 2006/07). A large inter-annual variability in the forcing is apparent for the annual mean as well as in the seasonal cycle. For example, with 100 ppb BC the net forcing varied for the three simulated years by ± 1 W m^{-2}, while the average forcing is only around 3 W m^{-2}. An even larger variability becomes apparent in the seasonal cycle (Fig. 8). In April and May, the minimum and maximum forcing can vary between less than 3 and more than 25 W m^{-2}. In contrast, at the beginning of the snow-covered period the forcing due to the presence of absorbers remains below 5 W m^{-2} and is relatively constant during the three simulated years. Several factors contribute to the inter-annual variability like differences in the incoming radiation and the length and timing of the snow-covered period (Fig. 8). However, Fig. 8 also demonstrates that the incoming short-wave radiation is not the major driver for the inter-annual variability because in April the largest forcing is observed in the year 2006/07 while the incoming short-wave radiation was smaller compared to the two preceding years. Therefore, the seasonality of the forcing is mainly driven by the timing of the snowfall and the melting of the snow. If solid precipitation occurs early in the winter season as in the simulations for 2006/07, a forcing can also occur in the period from October to December, which may be comparable or even larger than the forcing calculated for March or April.

The simulated annual mean of the forcing due to the presence of 100 ppb BC for the three years of simulation corresponds to 3 to 4.5 W m^{-2} for the radiative forcing (Fig. 7) and 2 to 4 W m^{-2} for the net forcing (Fig. S4). In the presence of 10 ppm dust the values for the forcing increase to 6 to 7 W m^{-2} and 5.5 to 7 W m^{-2} for the radiative and net forcing. This range corresponds to the detected BC in snow concentrations. Due to the uncertainties in the measurements (see Methods) the correct BC concentrations could be a factor of three higher. With 300 ppb BC the radiative and net forcing increases to 10.5 to 13 and 9 to 12.5 W m^{-2}.

Using the range of detected BC in snow concentrations ($\leq$ 100 ppb) the radiative forcing obtained with the local model is similar to results from previous global model runs. For example, Flanner et al. (2007) and Ménégoz et al. (2014) have reported annual means of the radiative forcing due to BC for the Himalayas and the Tibetan Plateau around 3.5 to 4 and 1 to 4 W m^{-2}. These calculations included either no feedbacks (Flanner et al., 2007) or only short adjustments neglecting long-term feedbacks like changes of atmospheric circulation or sea surface temperature (Ménégoz et al., 2014). Ménégoz et al. (2014) also included dust in

the snow, while Flanner et al. (2009) calculated that the addition of dust increases the total radiative forcing during springtime by 1 to 2 W m^{-2} in the considered region. However, in both studies the simulated annual averages of the BC in snow concentrations ranged from 100 to 200 ppb. Since a comparable radiative forcing was obtained with higher BC in snow concentrations it can be concluded that the sensitivities are lower in the global model compared to our local model. Since the approach to represent the radiative transfer in the snow is similar or even the same in all three models, a major reason for the lower sensitivity could be the simplified representation of the snow in the global models, which describe the entire snowpack with a limited number of layers (Flanner et al., 2007; Ménégoz et al., 2014). As a result, the energy budget and the snow temperature profiles may not be well described leading to a bias in the simulation of the snow aging in the presence of the absorbers and of the melting of the snow. Moreover, the coarse spatial resolution of the models does not allow to represent the specific local conditions at NCO-P.

Using a further global model, Kopacz et al. (2011) found BC in snow concentrations around 46 ppb in October in the Mt. Everest region. Using simple estimates of the relationship between BC in snow and albedo, they derived an average radiative forcing of 9 W m^{-2} for October with monthly means ranging from 7.5 to 12 W m^{-2}. While the simulated BC in snow concentrations are more reasonable (although still too high), the applied sensitivities are high compared to our simulations for NCO-P and do not reflect the complex processes occurring in the snowpack.

4   Conclusions

The here reported time series of rBC concentrations in the surface snow in the upper Khumbu valley indicates a seasonal cycle with maximum concentrations in the pre-monsoon period and low concentrations during the monsoon period. This cycle directly reflects the behavior of atmospheric BC in the same high altitude region (Marinoni et al., 2010) and is also conserved in ice cores from near-by glaciers (Ginot et al., 2014). Therefore, ice cores can be used to re-construct historic BC concentrations in these regions. Unfortunately, a large uncertainty remains regarding the absolute BC in snow concentrations. While the SP2 technique delivers precise concentrations of rBC, the melting of the snow samples during transport and storage potentially modified the detected concentrations. This effect probably contributed to discrepancies between the here reported snow concentrations compared to previously reported

results for EC, while the two different techniques also add to the mismatch (Lim et al., 2014). Further measurements are needed to resolve these discrepancies and to determine actual concentrations. Such accurate data are urgently needed for the validation of global and regional models used to calculate the impact of BC in snow on the regional climate (Qian et al., 2011; Ménégoz et al., 2014). While models simulating the impact of BC in snow on albedo and radiative forcing for a given snowpack exist, the quantification is currently mostly hampered by the uncertainty in BC in snow concentrations.

In its first application to the snowpack in the Himalayas, the standard Crocus model reveals significant discrepancies compared to observations with a large positive bias in the simulated albedo. While the Crocus model is well adapted to conditions in the Alps and gives satisfying results in polar regions (Jacobi et al., 2010; Brun et al., 2011; Vionnet et al., 2012), the high radiation intensities in the lower latitudes of the Himalayas make the model obviously more vulnerable due to the simplified parameterization of the albedo. While simulations were performed for typical conditions at Pyramid, it is likely that this simulated bias can be expected for wider regions of the high altitude area of the Himalayas. This concerns the simulation of the thin seasonal snowpack with a large contrast in albedo between the snow-free and –covered ground. The bias in the standard Crocus model is probably diminished in regions with the formation of a thicker snowpack like in the Western Himalayas (Ménégoz et al., 2013a) and in calculations over glaciers, where the underlying firn and ice exhibit a higher albedo than the ground. Furthermore, the parameterization of the ground heat fluxes in the Crocus model may not be well adapted to the conditions of the Himalayas possibly contributing to the over-estimation of the snowpack especially late and early in the winter season.

The bias in the albedo and the snowpack simulations are strongly reduced with the upgraded Crocus model taking into account the radiative transfer in the snow. This improvement becomes obvious even in the case when no absorbing impurities are considered in the simulations. The results are further enhanced if BC in the snow is considered and varied in a range constraint by the here presented surface snow and previous ice cores measurements (Ginot et al., 2014). Nevertheless, even in these simulations the albedo of and the length of the period with the seasonal snowpack remain overestimated. Since the simulations were all performed with constant BC concentrations, any enrichment of BC at the snow surface or in specific layers due to sublimation, dry deposition, melting, and refreezing (Doherty et al.,

2013) is not included. Nevertheless, even in the model runs with a BC concentration of 300 ppb, which is more than threefold the maximum observed BC concentration in the surface snow as well as the ice core from this region (Ginot et al., 2014), the snowpack is more persistent than observed. Thus, even if any or all of the post-depositional processes leads to an enrichment of more than a factor of three throughout the snowpack, differences in the simulations and the observations would remain. Stronger enrichments of BC in certain layers, like reported previously for the Mera glacier (Kaspari et al., 2013) or in the Arctic (Doherty et al., 2013), appear unlikely because they should have been detected in our surface snow samples or in the Mera ice core (Ginot et al., 2014). Moreover, the observed maxima of BC concentrations in the snow can easily be explained by dry deposition alone without invoking any further enrichment processes (Yasunari et al., 2013). These results do not rule out that higher BC in snow concentrations are encountered under different conditions like on glaciers or even higher altitudes, where the snowpack may not melt before April, when the air masses with the highest concentrations in BC finally arrive at the high-altitude region (Fig. 2).

A heterogeneous distribution of absorbing compounds within the snowpack may induce stronger temperature gradients inside the snowpack further accelerating the snow metamorphism compared to a snowpack with homogeneous concentrations as assumed in our simulations. Differing concentrations can be caused by processes during the melting of the snowpack as described above. Since the absorbing compounds are introduced by dry and wet deposition, concentration gradients inside the snowpack can be expected even in the not-melting snowpack. Moreover, these gradients can be different for different absorbing compounds. Vertical profiles of the absorbing compounds need to be determined to address this point. If it can also have a profound impact on the snowpack simulations remains to be seen and will be the subject of further studies.

The simulations with the Crocus model reveal further that dust and probably also other absorbers play a strong role not only for the snow albedo itself, but also for the impact of BC. While the impact of a given amount of an absorber on the snow albedo diminishes in the presence of further impurities, this is the opposite case for the increase of snow-free days and the radiative forcing. Here, the simulations demonstrated that the presence of other absorbers like dust even enhances the effect of the addition of BC. Therefore, in the future the role of dust in the snow needs to be studied together with the role BC and a correct determination of dust and its properties parallel to the determination of BC in the snow is needed. Organic

absorbers in the snow may also play a role (Wang et al., 2013) similar to dust and should also be considered in further studies. In summary, a full characterization of all absorbing compounds and their different contributions seems necessary to study the full impact of these compounds on the snow albedo and further related snow properties and processes.

Finally, this study concentrated only on the effect of the albedo for the snowpack simulations. Other processes parameterized in the snow model (e.g. turbulent fluxes, ground heat flux, snow metamorphism) and uncertainties in the forcing data may also contribute to differences between simulations and observations. Further detailed observations are needed to improve future snowpack simulations in this sensitive region. This mainly concerns measurements of the total and solid precipitation, which are crucial parameters in snowpack and hydrological modeling for this sensitive region.

Acknowledgements

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1   Table 1. Snowpack properties for 23 and 31 January 2005 simulated with the upgraded model

2   including radiative transfer and with different BC concentrations.

		BC = 0	BC = 100 ppb	BC = 300 ppb
23/01/2005	Snow height [cm]	52.8	52.6	52.4
12:00	SWE [cm]	5.66	5.64	5.64
	$T_{snowpack}$ [°C] [a]	-13.3	-13.2	-13.0
	$T_{10\,cm}$ [°C] [b]	-24.1	-23.9	-23.7
	Grain diameter [μm] [a]	301	303	305
31/01/2005	Snow height [cm]	33.6	32.4	31.1
12:00	SWE [cm]	5.51	5.45	5.41
	$T_{snowpack}$ [°C] [a]	-10.3	-9.7	-9.8
	$T_{10\,cm}$ [°C] [b]	-12.3	-11.7	-11.7
	Grain diameter [μm] [a]	369	386	400

3   [a] SWE-weighted average for the entire snowpack.

4   [b] SWE-weighted average for the top 10 cm of the snowpack.
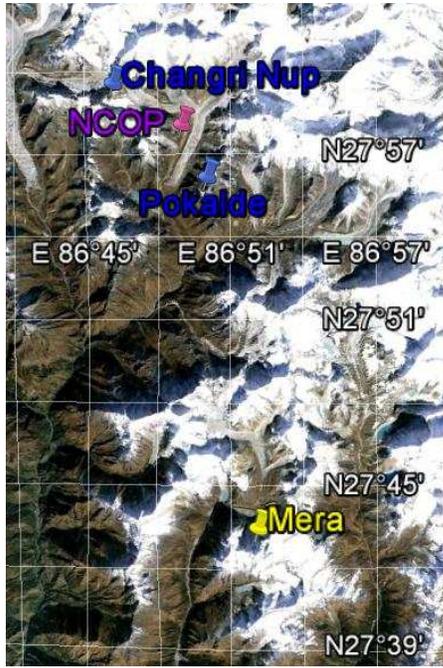
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3   Figure 1. Google earth map indicating the field sites NCO-P, Pyramid, Changri Nup, and

4   Pokalde. Also shown is the drilling site of the Mera ice core (Ginot et al., 2013).
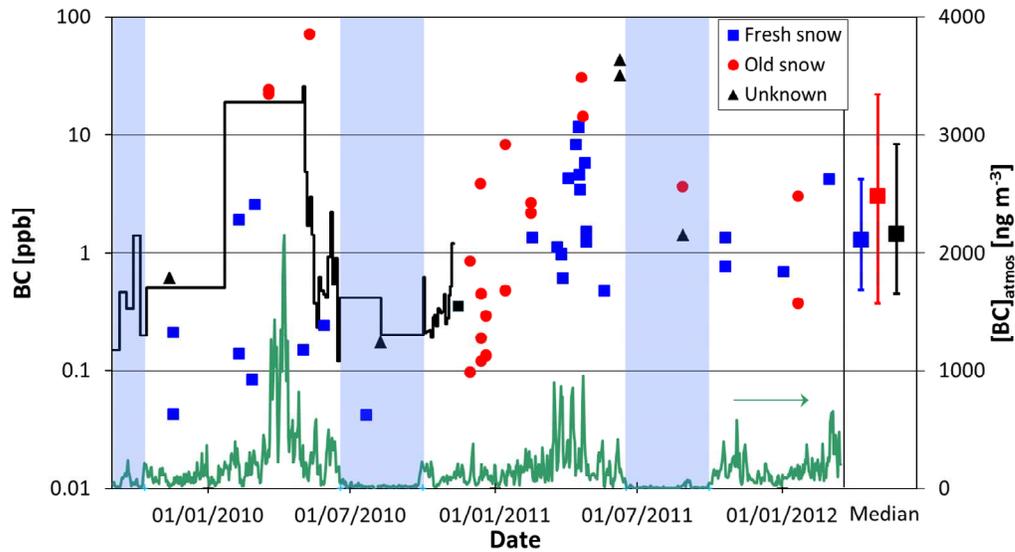
Figure 2. Time series of measured BC concentration in surface snow samples from the Khumbu Valley. The samples were classified into fresh (blue), old (red), and unknown snow (black). The symbols on the right show median concentrations for fresh (blue), old (red), and all snow samples (black). The error bars correspond to the 25th and 75th percentile. Shaded blue areas indicate the monsoon periods 2009, 2010, and 2011 over Nepal according to Tyagi et al. (2010, 2011) and Tyagi and Pai (2012). The black line corresponds to the BC concentration for the period between 1 September 2009 and 8 November 2010 season determined in the Mera ice core (Ginot et al., 2013) with the surface snow concentration from 11 November 2010 shown as black square. The green line shows the atmospheric BC concentrations measured at NCO-P.

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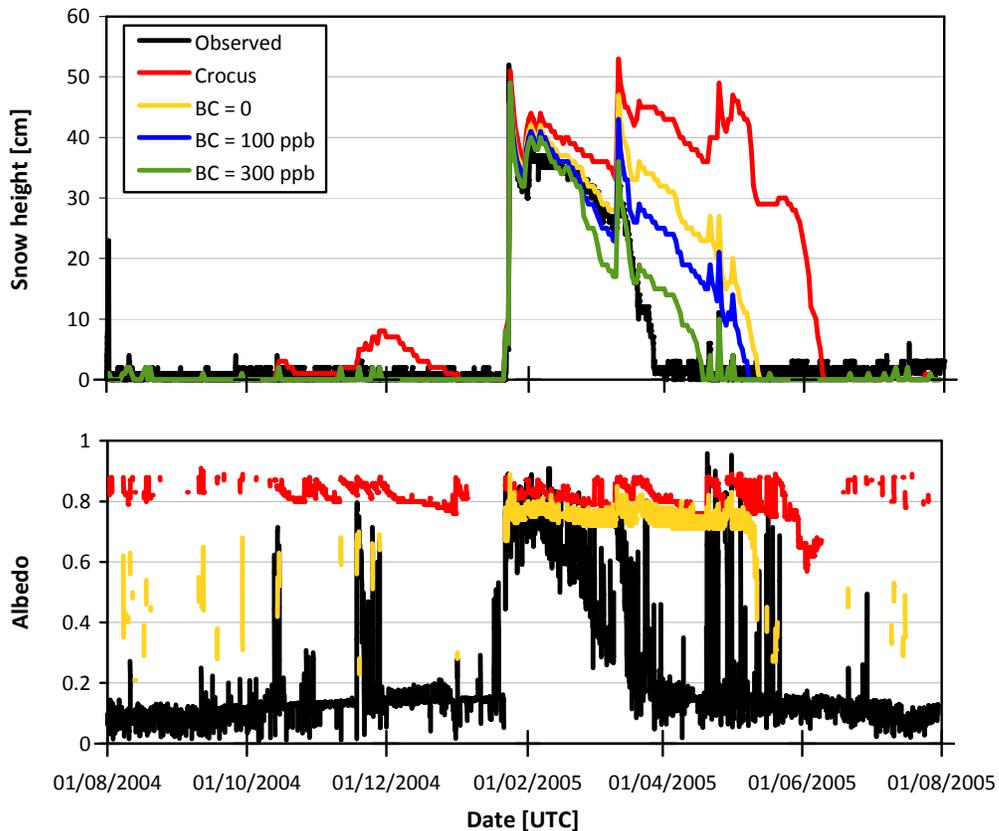
3   Figure 3. (Top) Comparison of observed (black) and simulated snowpack heights at NCO-P

4   for the winter season 2004/05. Simulations were performed with the standard crocus model

5   (red) and with the upgraded model including radiative transfer with constant BC

6   concentrations of 0 (yellow), 100 (blue), and 300 ppb (green). Snowpack heights simulated

7   with the upgraded model are indistinguishable for the different BC concentrations until late

8   January and after mid-May. (Bottom) Comparison of observed (black) and simulated albedo

9   at NCO-P. Simulations were performed with the standard crocus model (red) and with the

10   upgraded model including radiative transfer but without BC (yellow).
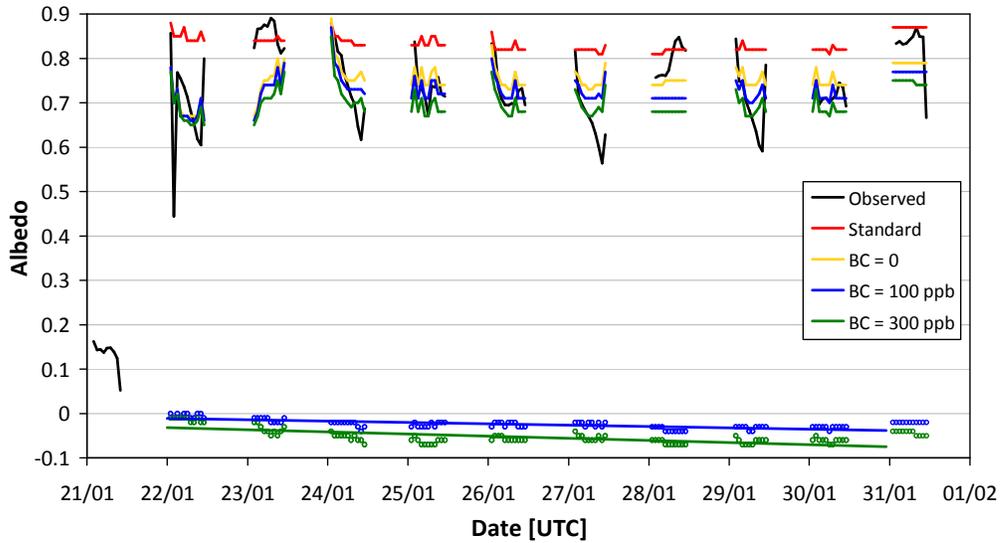
Figure 4. Comparison of observed (black) and simulated albedo at NCO-P for the period 22 to 31 January 2005. Model results are obtained with the upgraded model including radiative transfer with different BC concentrations: 0 (yellow), 100 ppb (blue), and 300 ppb (green). Circles at the bottom indicate the differences of the simulated albedo between BC = 100 ppb and 0 (blue) and BC = 300 ppb and 0 (green). Straight blue and green lines show the results obtained by linear regressions for the albedo differences during the period 22 to 30 January 2005.
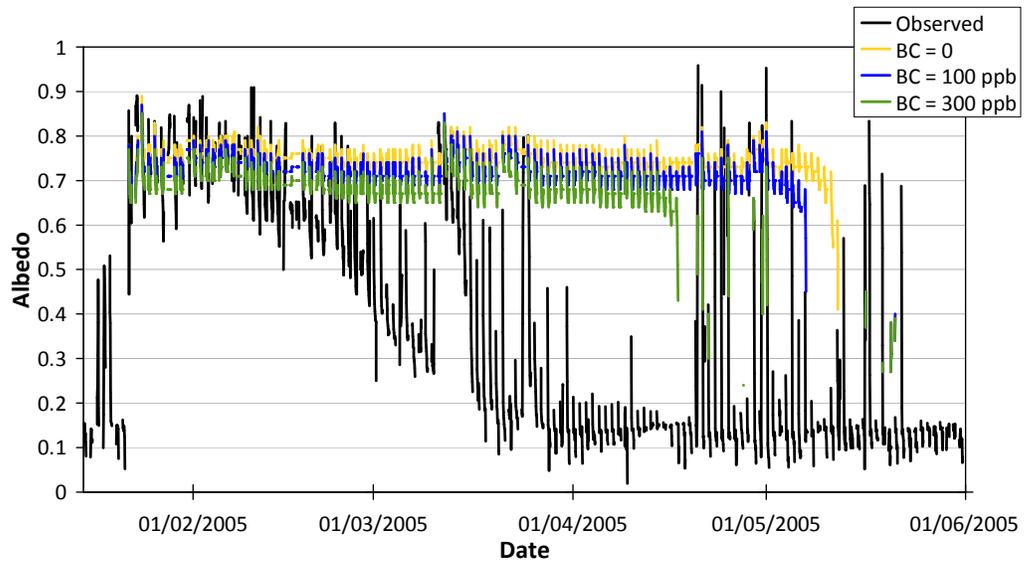
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3   Figure 5. Comparison of observed (black) and simulated albedo at NCO-P for the period 15

4   January to 1 June 2005. Model results are obtained with the upgraded model including

5   radiative transfer with different BC concentrations: 0 (yellow), 100 ppb (blue), and 300 ppb

6   (green).
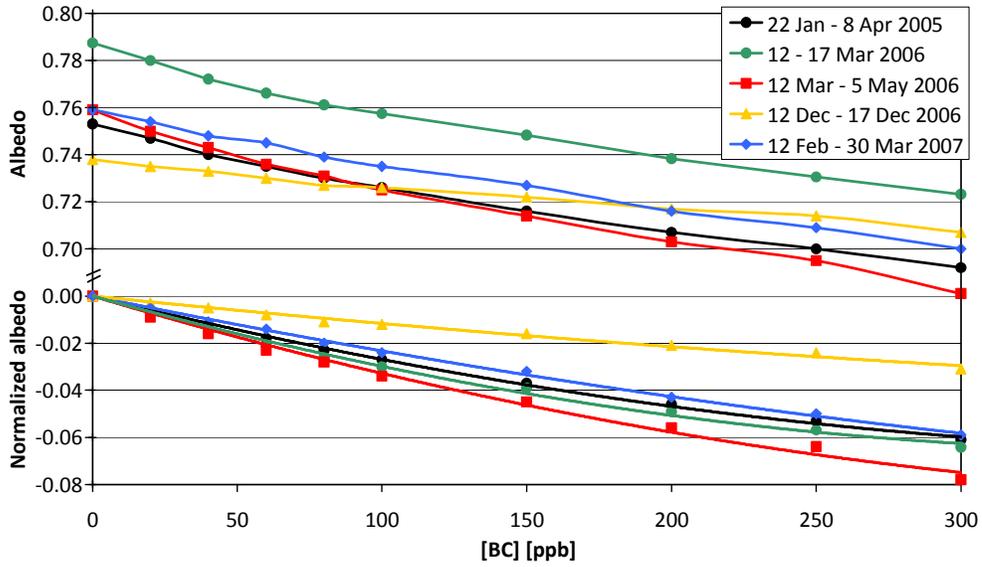
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3   Figure 6. (Top) Simulated average albedo for several periods with continuous snow higher

4   than 10 cm as a function of BC concentration. (Bottom) Normalized albedo as differences of

5   the averaged albedo minus the averaged albedo at BC = 0. The lines correspond to the best fit

6   of a quadratic polynomial forced through zero for each set of the normalized albedo.
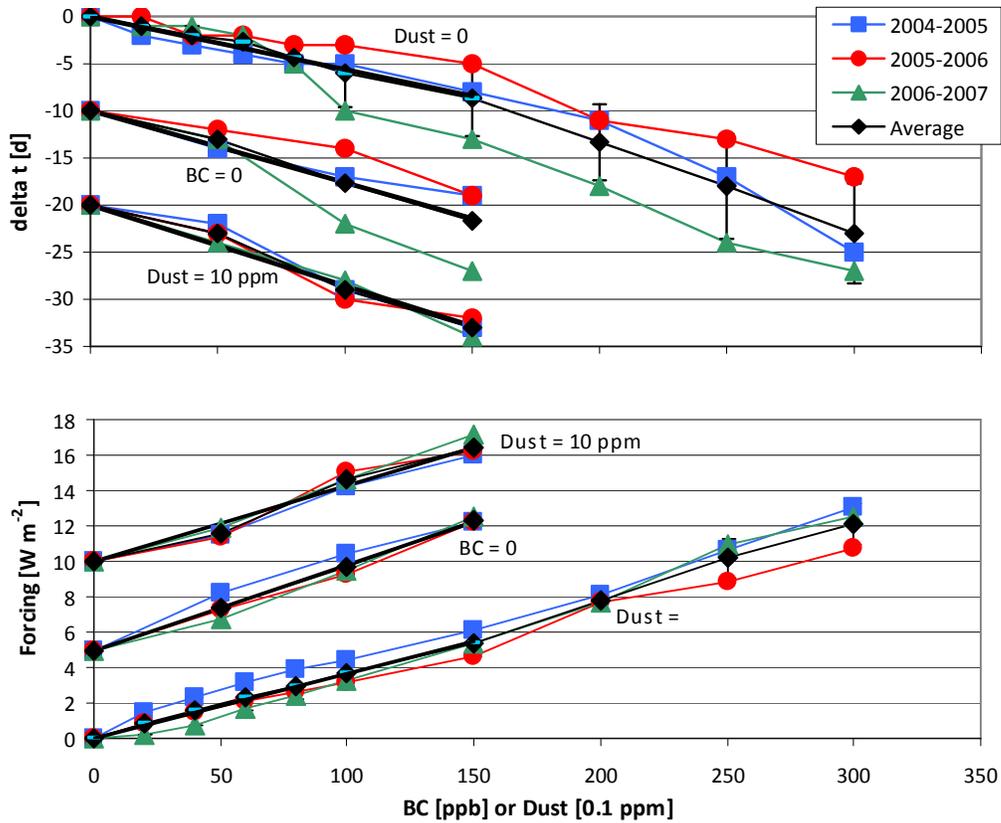
Figure 7. (Top) Reduction in the snow-covered period in days simulated for different BC and dust concentrations in the snow. (Bottom) Simulated annual radiative forcing related to shortwave radiation due to the presence of BC and dust in the snow. Simulations are performed without dust, without BC (shifted by -10 days or +5 W m^{-2}), and with dust = 10 ppm (shifted by -20 days or +10 W m^{-2}). In the last case, the reductions are calculated relative to the case with BC = 0 and dust = 10 ppm. Black symbols indicate the 3-year averages of the radiative forcing with the error bars representing the standard deviation. Black lines correspond to linear regressions forced through the origin for the average values for BC ≤ 150 ppb.
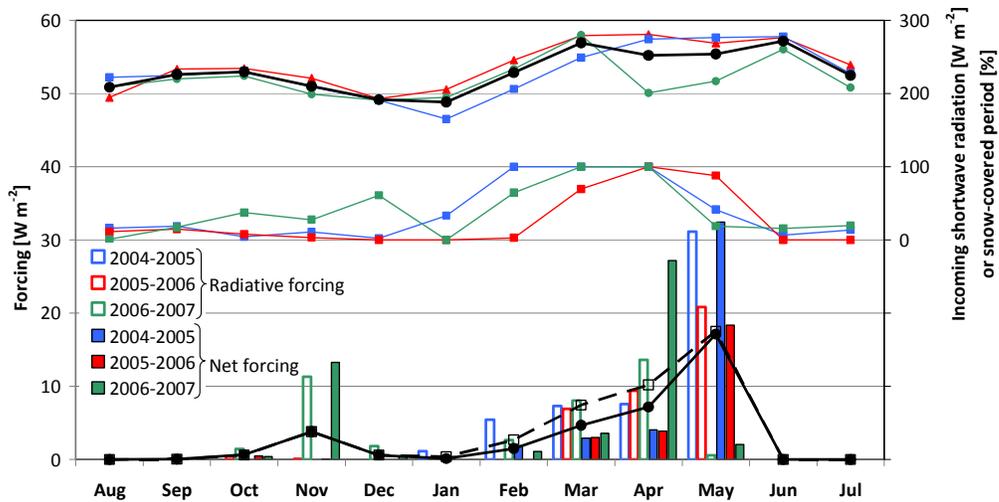
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3   Figure 8. Simulated monthly mean radiative (open bars) and net forcing (filled bars) for the
4   seasons 2004/05 (blue), 2005/06 (red), and 2006/07 (green) due to the presence of 100 ppb of
5   BC in the snow. Black open squares and filled circles indicate 3-year averages of the monthly
6   means of the radiative and net forcing. Also shown are snow-covered periods in percent based
7   on the simulations without BC in the snow (middle) and the monthly means of the observed
8   incoming short-wave radiation (top) for the three years. The black circles in the top panel
9   indicate the 3-year average of the monthly means of the incoming radiation.