Reviewer #1:

The authors present a study on the impact of different climate forcers on the regional climate of the Tibetan plateau. They used the CESM1 global model coupled to an aerosol scheme representing different types of aerosols like BC, dust, sea salt, sulfate and coupled to the CLM4 land surface scheme for the representation of snow and snow processes. The authors looked at the specific roles of CO2, BC in the atmosphere and in the snow, and sulfate in the atmosphere on the warming over the Tibetan plateau in the recent decades. They concluded that the simulations represent well the observed decrease in snow cover. They also state that BC plays a more important role in the observed warming over the Tibetan plateau compared to the global mean of the warming effect of BC. This stronger impact of BC, thus, contributes to the stronger-than-average increase of temperatures in the studied region.

While the presented results and conclusions are possibly justified based on the results of the simulations, I find that a major and indispensable step in the model validation is missing: It remains unclear how well the snow cover itself concerning parameters like extent, duration, or melting date is represented in the global model. It further remains unclear how well the BC in snow concentrations are simulated. For validation purposes the authors only show observed and simulated trends of the snow cover. Since these patterns are similar the authors assume that the model well represents the impact of a changing snow cover. In my opinion this conclusion is not justified based only on the presented data and further validation of the model is needed. Therefore, I recommend major revisions before a potential publication of the manuscript in ACP.

Response: Thanks for reviewing our paper. We agree that more model validation regard to snow cover climatology should be provided. We now added an additional supplement figure and many discussions in the text for this purpose. Please see the detailed responses below.

Comments: The authors need to define the meaning of the parameter “snow fraction”. For a full description of the snow cover multiple parameters are needed like snow height, SWE, snow cover extent, snow cover duration, snow melt-out dates, and so on. It remains unclear what parameter is used. Since the observations are based on remote sensing data, I assume that the snow fraction is related to snow cover extent? But what trend is shown in Figure 1? The trend in the maximum snow covered area or the period with snow cover? This needs to be specified.

Response: The data shown in Fig 1 is "Snow Cover Extent". We now clarify in the method section that we used "NOAA Climate Data Record of snow cover extent (Robinson et al., 2012)." The figure legend and caption of Fig 1 is modified to be "snow cover extent"). We also clarify in the Fig 1 caption that " The trend is calculated based on snow cover extent data in the entire period. " The model output of "snow fraction" refers to the same variable and that naming convention is retained in the manuscript.

It is well known that global models tend to overestimate the snow cover of the Tibetan plateau. One potential reason is that the blocking effect for the moisture transport crossing the Himalayas is too small due to the coarse resolution of the global models. As a result the precipitation over the Tibetan Plateau is overestimated. This limitation can partly be overcome with models using higher spatial resolutions (e.g. Ménégoz et al., 2013). By the way, how well are the high altitude regions represented in the used global model? The authors explicitly state that the observed warming has been important in high altitude regions. A spatial and temporal overestimation of the snow cover over the Tibetan Plateau in general will certainly lead to an overestimation of the snow-related effects. Therefore, it is crucial to validate the simulated snow cover using observations. Validating the model with simulated and observed trends can only be a second step.

Response: We now compare the simulated and observed present-day temperature, precipitation, and snow cover fraction in revised Figure S2.

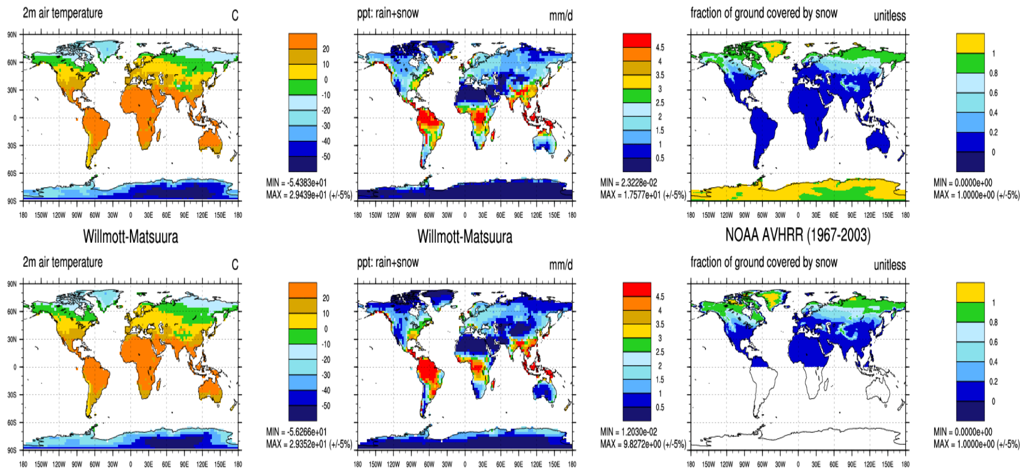


Fig S2. (Left) climatological surface air temperature (ºC) in the model simulation in the top panel, and observed surface air temperature in the bottom panel. (Middle) total precipitation (rain and snow fall) (mm/day) (Right) snow cover fraction. The model results in the top row are the 1981-2005 averages of the transient simulations under all radiative forcing. The temperature and precipitation observations are from updated dataset of Willmott and Matsuura (2001). The snow cover observations are from NOAA AVHRR as compiled by Robinson et al., (2012). In terrain-complex regions (such as North American Rockies, South American Andes and Tibet Plateau), the model tends to overestimate the precipitation and consequently snow cover, a bias commonly found in global climate model with coarse resolutions (Ménégoz et al., 2013). More detailed land model evaluations can be found in Lawrence et al., (2011).

We find that at global scale the agreement between model simulation and observation are reasonably good. However, as correctly pointed out by the reviewer, the precipitation over the Tibet Plateau tends to be overestimated by the model (Fig S2b) and therefore the snow cover is biased high in the model (Fig S2c), especially for winter season by 30-40%. We now note this model caveat in Section 3 when discussing the snow retreat trend, and we comment that future models using higher spatial resolutions (e.g. Ménégoz et al., 2013) will potentially improve the model fidelity. However, we still claim that as a global climate model used for climate attribution purpose, our current model outperforms several previous coarse-resolution models. For example, contrasting our Fig S2(c) to Fig 2 of Qian et al. (2011) which used a previous generation CAM3 with 2.8 degree, one can easily see

that the major biases in the interior of Tibet Plateau are significantly improved and the maximum snow cover along the mountain ranges are now better represented in our model.

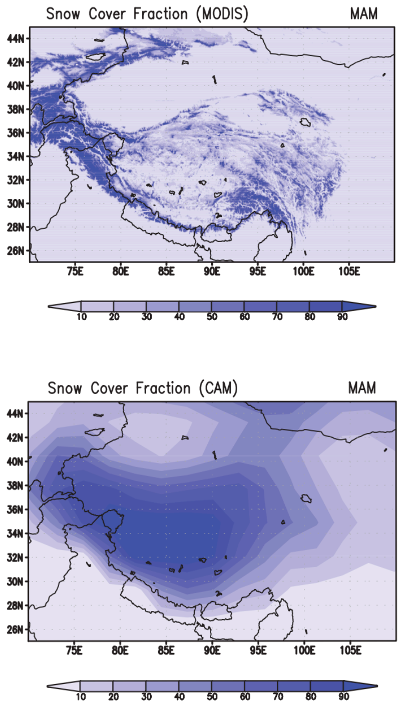


Fig. 2. Snow Cover Fraction (SCF) averaged for March-April-May (MAM) over Tibetan Plateau, (top) MODIS and (bottom) Model.

Fig 2 of Qian et al. (2011)

Qian, Y., Flanner, M. G., Leung, L. R. and Wang, W.: Sensitivity studies on the impacts of Tibetan Plateau snowpack pollution on the Asian hydrological cycle and monsoon climate, Atmos. Chem. Phys., 11, 1929–1948, doi:10.5194/acp-11-1929-2011, 2011.

We copied the discussion in Section 3 here for reviewer's reference. "Menon et al. (2010) attempted to simulate the snow reduction trends during 1990s but the spatial distribution of the observed trend was not well captured mainly due to the coarse resolution of the model. Qian et al. (2011) also acknowledged their model's limitation in representing the snow cover climatology and therefore may have biases in estimating BC impact on snow. It is well known that global models tend to overestimate the snow cover of the Tibetan Plateau, and one potential reason is that the blocking

effect for the moisture transport crossing the Himalayas is too small due to the coarse resolution of the global models and too much snowfall is simulated (Ménégoz et al., 2013). This limitation can partly be overcome with models using higher spatial resolutions. The modelling work presented here is a major step forward in terms of spatial resolution (about 1º by 1º), as opposed to earlier studies [2.8° by 2.8° in Flanner et al. (2009) and Qian et al. (2011); and 4° by 5° in Menon et al. (2010)], which helps better resolving the complex topography in this region. As a result of increased spatial resolution and also the improved land scheme, the biases in snow cover simulation is significantly reduced from its earlier model versions [Lawrence et al., (2011), also contrast Fig. S2c with Fig. 2 of Qian et al., (2011)]. However, we note that the precipitation over the Tibet Plateau is still overestimated (Fig S2b), and future studies, especially using regional climate models with even higher resolutions, are needed to improve the fidelity of model simulations of snow pack and glaciers over this topography-complicated region."

The impact of a changing snow cover and the involved feedback mechanisms are very complex and depend on many parameters: timing of the melt-out dates, incoming solar radiation, latitude, altitude, and possibly others. These parameters all influence the derived radiative forcing. For example, Jacobi et al. (2015) showed monthly averages of the radiative forcing related to the presence of BC in snow in the Himalayas. It can be assumed that if the melt-out dates are wrongly simulated the same shift in the melting of the snowpack can lead to an incorrect radiative forcing because it will not be similar for different months. Again, a correct model response regarding the impact of a changing snow cover can only be expected if the snow cover is correctly represented.

Response: We agree with these comments that radiative forcing due to BC in snow is also sensitive to the simulated snow cover in the background. We incorporated some more discussions on this in Section 4 as follows: "In addition to the uncertainty in BC loading, the forcing magnitude is also sensitive to model parameterization (Yasunari et al., 2013), and also the simulated background snow cover because the wrongly simulated melting dates of the snowpack can lead to an incorrect radiative forcing (Jacobi et al., 2015). Therefore, both in-situ (Wang et al., 2013; Zhao et al., 2014) and

laboratory measurements (Hadley and Kirchstetter, 2012) are needed to constrain model parameterizations of BC in snow."

I am also surprised to note that the simulated radiative forcing is larger for the reduction of the snow albedo due to the presence of BC compared to the radiative forcing caused by the earlier melting of the snowpack. This is opposite to results of many previous studies concerning light-absorbing impurities in snow (e.g. Flanner et al., 2007; Painter et al., 2007; Jacobi et al., 2015). Is this difference related to an overall limited representation of the snow cover in the model?

Response: Thanks for pointing out this discrepancy. We have checked the numbers in the model output again, and indeed the positive surface forcing initially due to BC deposition (Fig S4(b), calculated from the fixed SST simulation) is somewhat larger than the consequent snow-melting induced surface forcing (calculated from the fully coupled simulation). One caveat for our calculation of surface forcing due to BC deposition that it may be partially contaminated by the snow loss as the snow is already melting in the first five year of the simulation. We further clarify this issue in the caption of Fig S4 as follows "The change of surface albedo in (a) is calculated using the five years of atmosphere-only simulation in which BC emission is increased. Therefore, the albedo change largely represents the surface darkening due to BC deposition, although we cannot completely rule out the associated melting during this period. As a result, the actual radiative forcing at the surface due to BC in snow should be smaller than that in (b)."

We are currently incorporating a more proper radiation diagnostic procedure as in SNICAR without causing any fast feedback such as snow melting, which is similar to the one we used for calculating atmospheric forcing of various species. This will help better quantifying the BC surface darkening forcing from BC atmospheric heating. The relative contribution of the two is a future research topic of ours.

What are the simulated BC in snow concentrations? Do they correspond to observations? I admit that the available data are scarce, but still the few observations give an order of magnitude for the BC in snow in the Himalaya/TP region. If in the simulations the BC in snow concentrations are incorrect, but the simulated trends in the snow cover as well as in the albedo are correct, this would in my opinion suggest that the model sensitivity is incorrect.

Response: Due to accessibility to the original model output and the in-situ observation data, we were unable to perform this model validation step directly. A recent study (Zhang et al., 2015) used the same atmospheric and land snow model (but driven by realistic meteorological field in the year of 2000). They showed that simulated BC concentration is significantly larger than that from in situ sampling (Table S3 in Zhang et al., 2015), but suggested that the positive bias is smaller than what's previously reported in Ménégoz et al., (2014). However, as discussed in Ménégoz et al., (2014), they argue that "the spatial variations in BC deposition, can strongly affect the accuracy and representativeness of BC-in-snow measurements for the purpose of evaluating global models" and that "global models with coarse grid resolution cannot accurately represent elevation of sampling sites".

We now acknowledge these issues in the Section 5 related to the surface darkening effects as follows:" However, we note that model estimates of radiative forcing due to BC deposition on snow have large uncertainty. Using the same atmospheric and land model (but driven by realistic meteorological field in the year of 2000), Zhang et al., (2015) showed that simulated BC concentration in snow is biased high with respect to in situ sampling. Although the large spatial variations in BC deposition can affect the representativeness of BC-in-snow measurements for the model evaluation purposes, this potential model bias should be kept in mind."

The reviewer's concern on the fidelity of BC concentration in snow is a valid point. Although we have made previous efforts to constrain BC atmospheric radiative forcing in the model using satellite and ground radiometer measurements (Xu et al., 2013), the improvement on the accuracy of BC concentration in snow and a proper accounting of its radiative effect is a future research direction for us.

Zhang, R., Wang, H., Qian, Y., Rasch, P. J., Easter, R. C., Ma, P.-L., Singh, B., Huang, J., and Fu, Q.: Quantifying sources, transport, deposition, and radiative forcing of black carbon over the Himalayas and Tibetan Plateau, Atmos. Chem. Phys., 15, 6205-6223, doi:10.5194/acp-15-6205-2015, 2015.

SO4 should be substituted by either “sulfate” or SO2− 4. There are no SO4 emissions. The authors probably refer to emissions of SO2?

Responses: Thanks. We now define the acronym "SO4" for "sulfates" at its first occurrence, and this is consistent in the text and all figures. You are right about the emission. We now clarify in the method section that " The forcings were imposed by instantaneously increasing the emissions of BC, or the emission of SO_4's precursor sulfur dioxide, or by increasing CO_2 concentration to present-day level (400 ppm)."

References

Flanner, M.G., C.S. Zender, J.T. Randerson, and P.J. Rasch, Present-day climate forcing and response from black carbon in snow, J.Geophys.Res. 112, D11202, doi: 10.1029/2006JD008003, 2007.

Jacobi, H.W., S. Lim, M. Ménégoz, P. Ginot, P. Laj, P. Bonasoni, P. Stocchi, A. Marinoni, and Y. Arnaud, Black carbon in snow in the upper Himalayan Khumbu Valley, Nepal: Observations and modeling of the impact on snow albedo, melting, and radiative forcing, The Cryosphere 9, 1685-1699, 2015.

Ménégoz, M., H. Gallée, and H.W. Jacobi, Precipitation and snow cover in the Himalaya: From reanalysis to regional climate simulations, Hydrol.Earth Syst.Sci. 17, 3921-3936, 2013.

Painter, T.H., A.P. Barrett, C.C. Landry, J.C. Neff, M.P. Cassidy, C.R. Lawrence, K.E. McBride, and G.L. Farmer, Impact of disturbed desert soils on duration of mountain snow cover, Geophys.Res.Lett. 34, L12502, doi: 10.1029/2007GL030284, 2007.

Response: Thanks very much for providing those helpful suggestions on references. They are now cited in the paper.

Reviewer #2:

In this work, the authors estimate the role of BC increase in the tropospheric warming of the Tibetan Plateau region using a coupled AOGCM at high horizontal resolution, forced by observationally-based BC aerosol datasets.

Novelty: Use of High resolution coupled OAGCM with coupling between snow over land and BC deposition. I think this last improvement is crucial. This analysis is also based on a previous work where the BC forcing is re-estimated by using satellite + ground based optical depth, with a new methodology to separate the BC contribution to solar absorption from other aerosols and its direct rad forcing

I would recommend to publish the paper, after considering minor comments below:

Response: Thanks for reviewing our paper.

Specific comments

Lines 21-25. What do the CMIP5 models simulate as surface warming on that region in their historical simulation? Any reference about it?

Response: According to recent model evaluation papers, the strong warming trends in this region were not well captured by the CMIP5 historical simulations. For example, You et al., (2015) showed that "CMIP5 GCMs can reproduce the recent temperature evolution in the TP, but with cold biases... most CMIP5 GCMs underestimate the observed warming rates, especially the CNRM-CM5, GISS-E2-H and MRI-CGCM3 models."

We now include this reference in the 3rd paragraph of Introduction as follows: "To date global climate models forced by historical radiative forcing scenarios (such as those in Coupled Model Intercomparison Project Phase 5, CMIP5) have difficulty in simulating the observed record surface warming (You et al., 2015) or its anomalously strong altitude dependence in Tibet/Himalaya region."

You, Q., Min, J. and Kang, S.: Rapid warming in the Tibetan Plateau from observations and CMIP5 models in recent decades, Int. J. Climatol., n/a–n/a, doi:10.1002/joc.4520, 2015.

Line 15: I would suggest to add Lau et al. [2006] and Lau and Kim [2006] in the list of references

Response: Thanks for the suggestions. Now it is changed to "Many previous studies have linked Asian aerosols (including sulfates and BC) with monsoon systems and have demonstrated the aerosol impact on the summer rainfall (Ramanathan et al., 2005; Lau et al., 2006; Lau and Kim (2006); Meehl et al., 2008). "

Lau, K. M., Kim, M. K. and Kim, K. M.: Asian summer monsoon anomalies induced by aerosol direct forcing: The role of the Tibetan Plateau, Clim. Dyn., 26(7-8), 855–864, doi:10.1007/s00382-006-0114-z, 2006.

Lau, K. M. and Kim, K. M.: Observational relationships between aerosol and Asian monsoon rainfall, and circulation, Geophys. Res. Lett., 33(21), 1–5, doi:10.1029/2006GL027546, 2006.

Methodology BC treatment in the model: If you may briefly summarise here in which consists the correction you applied by the Xu et al papers, that would be very useful for the reader

Response: We now add more details on how the corrections were done in the model: "The present-day BC emission is adjusted from the standard model emission inventory (Lamarque et al., 2010) to account for the potential model underestimation of BC forcing. Emissions over East Asia regions are increased by a factor of two and South Asia regions by four. The emissions are adjusted by the same ratio in all economic sectors (energy, industrial, etc.) and all seasons by the same ratio. "

Model experiments Could you please specify here that you increase separately BC, SO4 CO2 in the perturbed equilibrium 5 ensemble members simulations?

Response: Yes, we now clarify that we used "(b) Four sets of perturbed simulations with instantaneously imposed present-day forcing: BC, SO4, CO2 and all three forcing combined. " and

"(c) Three sets of perturbed simulations but with fixed sea surface temperature. These are also forced by the instantaneous increase of BC, SO_4 and CO_2, separately,"

Which preindustrial and present day emissions have you used?

Response: We now added that "Except for the adjusted BC present-day emission as detailed in section 2.2(b), all other emission/concentration are from the standard inventory adopted by CMIP5 models, as descripted in Lamarque et al., (2010)."

Lamarque, J.-F., Bond, T. C., Eyring, V., Granier, C., Heil, A., Klimont, Z., Lee, D., Liousse, C., Mieville, A., Owen, B., Schultz, M. G., Shindell, D., Smith, S. J., Stehfest, E., Van Aardenne, J., Cooper, O. R., Kainuma, M., Mahowald, N., McConnell, J. R., Naik, V., Riahi, K. and van Vuuren, D. P.: Historical (1850–2000) gridded anthropogenic and biomass burning emissions of reactive gases and aerosols: methodology and application, Atmos. Chem. Phys., 10(15), 7017–7039, doi:10.5194/acp-10-7017-2010, 2010.

Not clear, lines 21 on: in the perturbed simulations you impose BC, SO4 and GHG as concentrations or you apply emissions for BC and sulphur (as specified in lines 8-12) and specify CO2 concentration?

Response: We now clarify that "The forcings were imposed by instantaneously increasing the emissions of BC or SO4, or by increasing CO_2 concentration to present-day level (400 ppm)."

Section 3. It seems to me by looking at figs 1, and S1 that there is an important decadal modulation of the snowcover, more than a trend. Would it be possible to look at the area averaged time series of snow cover from 1967 from dataset NSIDC? By averaging where there is a negative (blue) linear trend. Also applying a running mean could be useful in order to help in understanding the amplitude of such low-frequency variations. How different is this variability simulated by the model (at decadal timescale) w.r.t. observations in snow cover (for the 40 years)?

Response: Thanks. This is an important point regard to the snow cover variability/trend. Our argument as laid out in the submitted manuscripts was that at the multi-decadal time scale (1967-

2012 as in Fig 1), the changes along the Himalaya mountain range are long-term trends for which we aimed to attribute to various external forcings, while the changes at the multi-year time scale (positive changes in in 2001-2012, as in Fig S1(a,b) and negative changes in 1980-1991 as in Fig S1(c)) are more subject to the natural variability.

We have only showed the regional map of the snow fraction changes during the two 10-year periods in the submitted manuscript (Fig S1). Below is the time series of the regionally averaged snow cover for the last 40 years (smoothed to remove inter-annual variability) as suggested by the reviewer, and it supports our argument that the long-term trend is much weaker than short-term variation. Similar discussions were also found in many previous literatures, especially for seemingly increasing trend after year 2000 as more data become available.

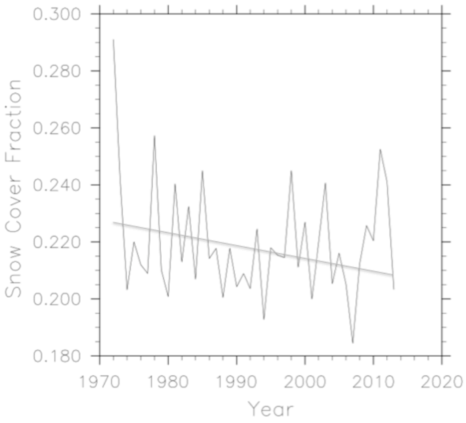


Figure. Annual mean snow cover fraction averaged over the Tibetan Plateau and its vanity region (20–50°N, 70–110°E), as the domain used in Fig 4.

However, we do acknowledge that the snow cover observational dataset over this region has issues with data availability. Therefore, we also cited a few other in-situ studies on mountain glaciers and permafrost to support the declining trend, in the beginning of section 3. " In-situ studies on regional glaciers and snow pack also reported strong declining trends. For example, Ma and Qin (2012) used 754 stations in China to document statistically significant declining trends of spring snow for the

Qinghai-Tibet Plateau for the 1951-2009 period. Consistently, permafrost degradation has been reported on the Tibet Plateau (Cheng and Wu, 2007; Li et al., 2008)."

You ascribe a better simulated snow cover to a high-resolution model (that may be ok for the better simulated orography), is this really the only factor?

Response: A number of new model features may have contributed to the improvement of snow cover simulation. We discussed in the method section that "The land model (CLM4) also includes major updates, making it more versatile in simulating snow packs (Lawrence et al., 2011). .....other parameterizations include snow compaction (Lawrence and Slater, 2010) and the albedo calculations for snow on or around vegetation (Wang and Zeng, 2009). Compared to the previous model versions, the albedo contrast between snow-covered and non-snow-covered area is more consistent with observations..." The model evaluation against observations and its improvement over older model versions were extensively documented by Lawrence et al., (2011).

However, we do think that the higher resolution and better resolved orography is a big contributing factor for this region, especially compared to a few cited previous studies on the same issue (Flanner et al., 2009, Qian et al., 2011, and Menon et al., 2010).

Is Figure 3 only over 80-100E or 0-360 longitude global average?

Response: It is the global averaged zonal mean. We now clarified in the figure caption " Globally zonal averaged radiative heating rate ....Fig. S2 shows the normalized temperature profile averaged just over the Tibetan Plateau (30 to 40ºN and 80 to 100ºE)."

Section 4 Which is the role of water vapour feedback in the T change increase versus altitude? And which is the role of changes in clouds?

Response: It is definitely an important point worth discussing. We now include the following sentences in the end of the discussion section. "Beyond the three main factors as we have discussed above, the changes of water vapour and clouds are also possible mechanisms contributing to the

elevation-dependent warming in the mountain regions. As shown in the schematic of a recent review paper (Mountain Research Initiative EDW Working Group, 2015), in a warmer and moister atmosphere, the latent heat release at the cloud condensation level may induce larger warming in high altitudes (cloud feedback) and the downward longwave radiation increase particularly fast in higher and drier atmosphere (water vapour feedback). It is difficult to identify or separate the contribution of these individual feedbacks from our current experiment setup. However, we note that these feedback mechanisms are operating regardless of forcing agents and therefore cannot explain the particularly large elevated warming in response to BC."

Mountain Research Initiative EDW Working Group.: Elevation-dependent warming in mountain regions of the world, Nat. Clim. Chang., 5(5), 424–430 [online] Available from: http://dx.doi.org/10.1038/nclimate2563, 2015.

How realistic is the simulated mean state and variability of the model temperature in this region?

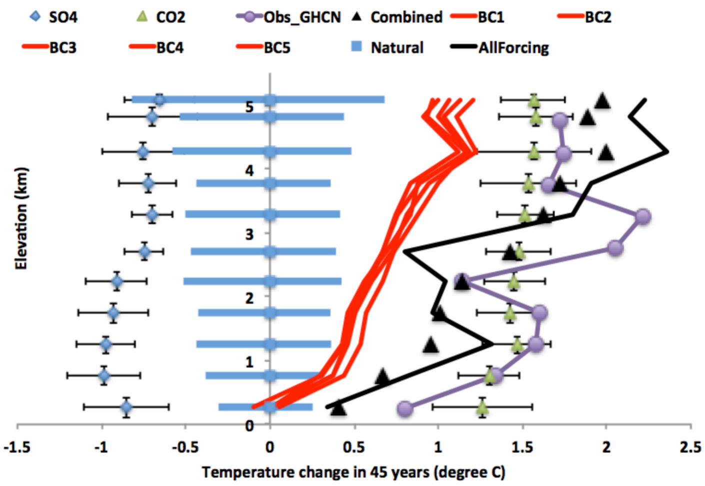
Response: The evaluation of model performance in simulating regional temperature, precipitation and snow cover are now included in Fig S2. This is also in response to comments from reviewer #1. Overall, the magnitude and spatial pattern of model simulation are in general agreement with observation. However, we now noted that the precipitation biases might lead to an overestimated snow cover, which affects the interpretation of the results. According to Reviewer #2, this is common in global climate models with coarse resolutions, and our 1-degree model has outperformed previous studies.

Related discussions are copied here for reviewer's reference. "Menon et al. (2010) attempted to simulate the snow reduction trends during 1990s but the spatial distribution of the observed trend was not well captured mainly due to the coarse resolution of the model. Qian et al. (2011) also acknowledged their model's limitation in representing the snow cover climatology and therefore may have biases in estimating BC impact on snow. It is well known that global models tend to overestimate the snow cover of the Tibetan Plateau, and one potential reason is that the blocking effect for the moisture transport crossing the Himalayas is too small due to the coarse resolution of

the global models and too much snowfall is simulated (Ménégoz et al., 2013). This limitation can partly be overcome with models using higher spatial resolutions. The modelling work presented here is a major step forward in terms of spatial resolution (about 1° by 1°), as opposed to earlier studies [2.8° by 2.8° in Flanner et al. (2009) and Qian et al. (2011); and 4° by 5° in Menon et al. (2010)], which helps better resolving the complex topography in this region. As a result of increased spatial resolution and also the improved land scheme, the biases in snow cover simulation is significantly reduced from its earlier model versions [Lawrence et al., (2011), also contrast Fig. S2c with Fig. 2 of Qian et al., (2011)]. However, we note that the precipitation over the Tibet Plateau is still overestimated (Fig. S2b), and future studies, especially using regional climate models with even higher resolutions, are needed to improve the fidelity of model simulations of snow pack and glaciers over this topography-complicated region."

Role of natural variability: why the 80%, may you also show for consistency 90 and 95% ?

Response: 80% corresponds to the 10th to 90th percentile range of the 350 member of the 45-year trend. We show below the same figure as Fig 5, but showing 90%. The observed trends and modeled trend under all forcing is still far beyond the natural variability range (blue shading, now larger than Fig 5 in submitted manuscript).



However, we still retain The 80% (10th to 90th percentile) probability range as in Fig 5. This approach is consisent with several recent papers. For example, Fig 1 of Dai et al., (2015) shows "...the blue vertical bar indicates the 10th to 90th percentile range of the internal variability of T estimated using the CESM1 30-member ensemble simulations"

Dai, A., Fyfe, J. C., Xie, S.-P. and Dai, X.: Decadal modulation of global surface temperature by internal climate variability, Nat. Clim. Chang., 5(6), 555–559 [online] Available from: http://dx.doi.org/10.1038/nclimate2605, 2015.

If we use the ctrl simulation to estimate the natural variability, how different would be the estimate with a different model, i.e. for example another ctrl simulation coming from the CMIP5 or maybe all the CMIP5 simulations. Would also in this case the observed trends be significantly “outside” the natural variability?

Response: The pre-industrial controls from other CMIP5 are generally available. Presumably, the internal variability certainly is model dependent, but there is no evidence that CESM1 is an outlier in terms of unforced natural variability. In fact, in a recent BAMS article (Kay et al., 2015), 30-member ensemble of CESM1 (1-degree and the same configuration as in our control simulations) were examined with respect to CMIP5 models. It is found that "in Fig 6, the trend spread generated by internal climate variability alone—estimated using the CESM-LE—is often statistically indistinguishable from the spread in trends within CMIP5. At least for DJF surface air temperature trends."

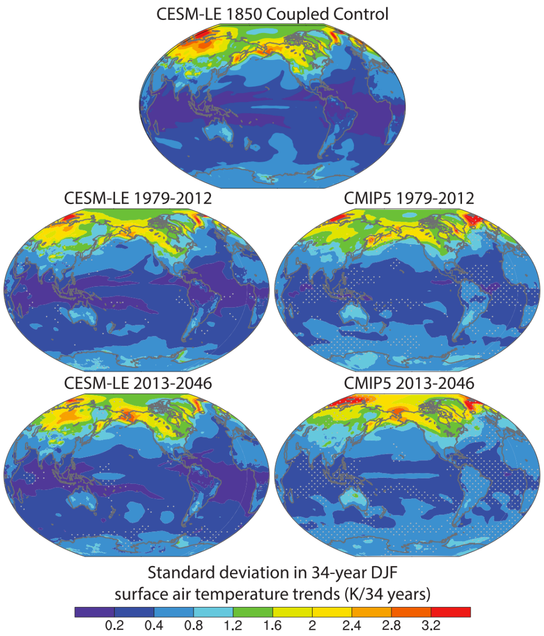


Fig. 6 of Kay et al., (2015). Global maps of standard deviation in 34-yr DJF surface air temperature trends for the (top) preindustrial (1850), (middle) historical (1979–2012), and (bottom) near-future (2013–46) periods. For the historical and near-future periods, trends are shown for both the 30-member CESM-LE ensemble and the 38-member CMIP5 ensemble (Taylor et al. 2012). Stippling on the historical and near-future CESM-LE trend maps indicates standard deviations that are statistically different than the CESM-LE preindustrial period. Stippling on the historical and near-future CMIP5 maps indicates standard deviations that are statistically different than the CESM-LE for the corresponding period. Stippling is based on an f test and a 95% confidence interval. For CMIP5, we used a single (the first) ensemble member of the following models.

Kay, J. E., Deser, C., Phillips, A., Mai, A., Hannay, C., Strand, G., Arblaster, J. M., Bates, S. C., Danabasoglu, G., Edwards, J., Holland, M., Kushner, P., Lamarque, J.-F., Lawrence, D., Lindsay, K., Middleton, A., Munoz, E., Neale, R., Oleson, K., Polvani, L. and Vertenstein, M.: The Community Earth System Model (CESM) Large Ensemble Project: A Community Resource for Studying Climate Change in the Presence of Internal Climate Variability, Bull. Am. Meteorol. Soc., 96(8), 1333–1349, doi:10.1175/BAMS-D-13-00255.1, 2015.

Therefore, we now include the following discussions in the manuscript. "One further concern is that the internal variability deduced from the long-term pre-industrial control simulation can be model dependent. However, comparing a 30-member ensemble of CESM1 simulation (1-degree resolution and the same configuration as in our control simulations) with 38-member CMIP5, Kay et al., (2015) found that the ensemble spread in the 30-year trend of surface temperature in CESM1 ensemble is statistically the same with the spread in trends within CMIP5."

Change in the UT temperature (for example as in Fig 2), do imply any significant change in convection and precipitation in the model?

Response: This is true. We now added that "The temperature response in the troposphere is associated with strong meridional circulation change. The mechanisms behind the free atmosphere circulation change, especially for the SO4 case which does not have strong atmospheric forcing, is discussed in details in Xu and Xie (2015)." In addition, we are actively studying the precipitation pattern at global scale utilizing similar model experiments.

Xu, Y., and S.-P. Xie (2015), Ocean mediation of tropospheric response to reflecting and absorbing aerosols, Atmospheric Chemistry and Physics, 15(10), 5827–5833, doi:10.5194/acp-15-5827-2015.

How important is the indirect effect in the model? Is it a minor contributor to the simulated and discussed changes?

Response: The aerosol indirect effect due to cloud changes are simulated by the model as we stated in the method section "The new cloud microphysics scheme (Morrison and Gettelman, 2008) allows the number concentration of cloud drops and ice crystals to be affected by aerosol concentrations and therefore accounts for the "indirect radiative forcing" of aerosols." The indirect forcing constitutes a large faction of SO4 forcing. A smaller fraction of BC forcing is due to indirect effect as BC is assumed less water soluble in the model (Liu et al., 2012).

Morrison, H. and Gettelman, A.: A new two-moment bulk stratiform cloud microphysics scheme in the community atmosphere model, version 3 (CAM3). Part I: Description and numerical tests, J. Clim., 21, 3642–3659, doi:10.1175/2008JCLI2105.1, 2008.

Liu, X., Easter, R. C., Ghan, S. J., Zaveri, R., Rasch, P., Shi, X., Lamarque, J. F., Gettelman, a., Morrison, H., Vitt, F., Conley, a., Park, S., Neale, R., Hannay, C., Ekman, a. M. L., Hess, P., Mahowald, N., Collins, W., Iacono, M. J., Bretherton, C. S., Flanner, M. G. and Mitchell, D.: Toward a minimal representation of aerosols in climate models: Description and evaluation in the Community Atmosphere Model CAM5, Geosci. Model Dev., 5, 709–739, doi:10.5194/gmd-5-709-2012, 2012.

You never discuss if there is any role of the dust. Is there any change in the transported dust? For example is that possible that in the simulation with increased CO2 / BC the pathways of transport of dust are changed because for example there is a different El Nino (Kim et al., Climate Dynamics 2015)?

Response: Thanks for raising this question. Firstly, we now include a paragraph in the end of section 4: "Lastly, it is also worth commenting the role of other snow impurities. In this study we used BC, a strong solar radiation absorber, to understand the climate response and the mechanisms due to absorbing aerosols that also include dust (Di Mauro et al., 2015; Gabbi et al., 2015) and organic aerosols (Qian et al., 2015). Similarly, we used SO4 to characterize all other reflecting aerosols. Any changes of dust and organics may induced changes to the snow cover, as their atmospheric heating and surface deposition are readily captured by this model, although the magnitude of response might be smaller since they are partially reflecting as well."

Gabbi, J., Huss, M., Bauder, A., Cao, F., and Schwikowski, M.: The impact of Saharan dust and black carbon on albedo and long-term mass balance of an Alpine glacier, The Cryosphere, 9, 1385-1400, doi:10.5194/tc-9-1385-2015, 2015.

Yun Qian, Teppei J. Yasunari, Sarah J. Doherty, Mark G. Flanner, William K. M. Lau, Jing Ming, Hailong Wang, Mo Wang, Stephen G. Warren, Rudong Zhang. (2015) Light-absorbing particles in

snow and ice: Measurement and modeling of climatic and hydrological impact. Advances in Atmospheric Sciences 32, 64-91.

Di Mauro, B., F. Fava, L. Ferrero, R. Garzonio, G. Baccolo, B. Delmonte, and R. Colombo (2015), Mineral dust impact on snow radiative properties in the European Alps combining ground, UAV, and satellite observations. J. Geophys. Res. Atmos., 120, 6080–6097. doi: 10.1002/2015JD023287.

Secondly, This is an interesting point that dust emission is changing as a response to climate change. We looked at the dust AOD difference between BC driven warming and pre-industrial simulation, and there is no robust changes in this region (see the figures below). The dust changes under CO2 warming are relatively larger. But the changes are still not statistically significant, and radiative forcing due to dust change is certainly smaller than CO2 radiative forcing.

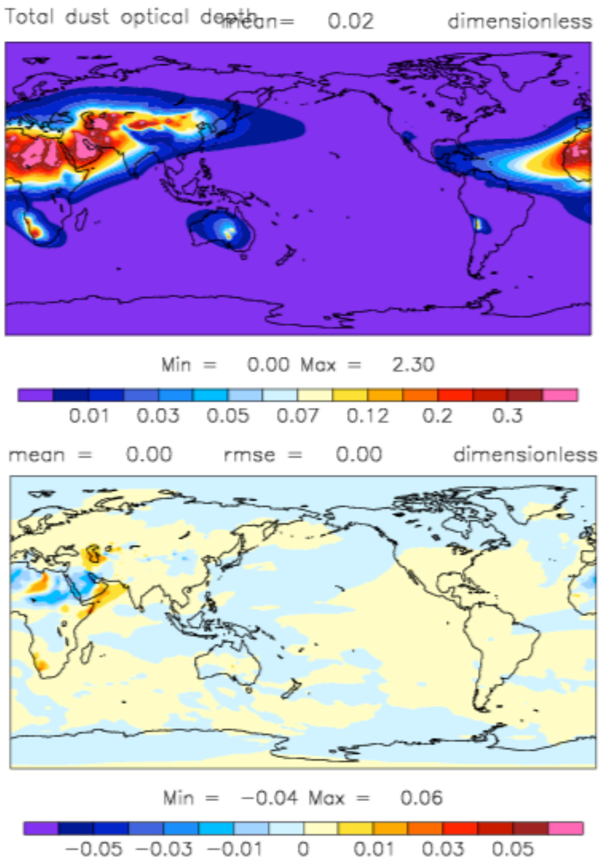


Fig. (Upper) Dust aerosol optical depth in pre-industrial run. (Lower) Change of Dust AOD in response to BC warming.

However, this "aerosol-feedback" is worth further investigation. Actually, we are studying the aerosol changes solely due to global warming. (Xu, Y., et al., Global warming impact on future PM2.5 pollutions, in preparation). The dust emission and transport under different ENSO conditions as in Kim et al., (2015) is a good reference point for our future work.

Fig S6 is missing!

Response: Sorry for the oversight. That figure is now removed from the manuscript. We now only briefly mentioned in the discussions that "A look at the seasonality of snow depth change suggests the early spring melting is important for this feedback."

Observed high-altitude warming and snow cover retreat over Tibet and the Himalayas enhanced by black carbon aerosols

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[1]{National Center for Atmospheric Research, Boulder, CO}

[2]{Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California}

Abstract

Himalayan mountain glaciers and the snowpack over the Tibetan Plateau provide the headwater of several major rivers in Asia. In-situ observations of snow cover extent since the 1960s suggest that the snow pack in the region have retreated significantly, accompanied by a surface warming of 2-2.5ºC observed over the peak altitudes (5000 m). Using a high-resolution ocean-atmosphere global climate model and an observationally constrained black carbon (BC) aerosol forcing, we attribute the observed altitude dependence of the warming trends as well as the spatial pattern of reductions in snow depths and snow cover extent to various anthropogenic factors. At the Tibetan Plateau altitudes, the increase of atmospheric CO_2 concentration exerted a warming of 1.7ºC, BC 1.3ºC where as cooling aerosols cause about 0.7ºC cooling, bringing the net simulated warming consistent with the anomalously large observed warming. We therefore conclude that BC together with CO_2 has contributed to the snow retreat trends. Especially, BC increase is the major factor in the strong elevation dependence of the observed surface warming. The atmospheric warming by BC as well as its surface darkening of snow are coupled with the positive snow albedo feedbacks to account for the disproportionately large role of BC in high-elevation regions. These findings reveal that BC impact needs to be properly accounted for in future regional climate projections, in particular on high-altitude cryosphere.

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1 Introduction

Himalayan mountain glaciers and snow packs have a major impact on the water systems of major rivers throughout Asia and the people living in the river basins. Recent observations suggested a continuing decline in Himalayan mountain glaciers and snow cover. Bajracharya et al. (2008) observed that the Himalayan glaciers are retreating at rates ranging from 10 to 60 m per year, and many small glaciers have disappeared. Gardner et al. (2013) also showed with satellite observations the steady reduction of Western China glaciers with the most rapid decline observed in the Himalayan mountain regions. Changes in the cryosphere are accompanied by documented surface warming trends over Tibet, which reveals a strong altitude dependence of surface warming with peak warming trends of 2-2.5°C at 5000 m from 1961 to 2006 (Liu et al., 2009).

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The last few decades also witnessed rapid growth in human population and economic activities, causing intense air pollution over the Asian region. Among the many air pollutants, black carbon (BC) aerosols have been shown to have a significant impact on global and regional climate change (Ramanathan and Carmichael, 2008). Many previous studies have linked Asian aerosols (including sulfates and BC) with monsoon systems and have demonstrated the aerosol impact on the summer rainfall (Ramanathan et al., 2005; Lau et al., 2006; Lau and Kim (2006); Meehl et al., 2008). The BC aerosols have also been shown to have impact on warming trends over the Himalayan/Tibetan region (Ramanathan et al., 2007), on the retreat of Himalayan glaciers (Menon et al., 2010; Qian et al., 2011), and on Eurasian snow cover (Flanner et al., 2009). Observationally, using ice-core samples to reconstruct historical BC content over Tibet, Xu et al., (2009) suggested BC is a significant contributing factor in causing the glacier change.

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To date global climate models forced by historical radiative forcing scenarios (such as those in Coupled Model Intercomparison Project Phase 5, CMIP5) have difficulty in simulating the observed record surface warming (You et al., 2015) or its anomalously strong altitude dependence in Tibet/Himalaya region. One possible explanation as we will investigate here is that few of these earlier studies of the Himalayan and Tibetan climate change have considered the combined effects of all the following factors: BC direct heating of the atmosphere, the heating of snow packs and glaciers by BC darkening the snow and ice, the greenhouse effect of CO_2 and the surface cooling effects by aerosols other than BC.

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In this study, we used a state-of-the-art global climate model to conduct a suite of model experiments to understand BC's role in the cryosphere change over the Himalaya and Tibetan region. A unique feature of the present study, compared with earlier studies, is that BC radiative forcing is constrained with multiple sources of observations (satellite observed aerosol optical depths and ground network of spectral sunphotometer measurements). We also used a newly developed method to separate the BC contribution to solar absorption from other aerosols (sulphates, organics and brown carbon) and calculate its direct radiative forcing (Bahadur et al., 2012; Xu et al., 2013). Previous studies (Ramanathan et al., 2007; Lau et al., 2010; Menon et al., 2010; Qian et al., 2011) included the effects of BC on the atmosphere and the cryosphere, but the simulated BC radiative forcing by these “standard” models used in CMIP5 is strongly biased to low values (Bond et al., 2013) due to emission inventory biases and missing physical treatments (Jacobson, 2012). As shown in Bond et al. (2013), current models are underestimating BC solar absorption over South Asia by a factor of two to five. In this study, we scaled the simulated BC forcing in the climate model by factors ranging from two to four to bring the simulated values closer to the observationally constrained values (Xu, 2014). Another improvement in this study is that the simulations were conducted using a fully coupled ocean-atmosphere-land model at a high resolution of 1° by 1°, in which a new land snow module is adopted (Lawrence et al., 2011) to account for BC deposition effect on snow and ice.

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2 Methods

2.1 The global climate model

CESM1 (Community Earth System Model 1) is a coupled ocean-atmosphere-land-sea-ice model. CESM1 climate simulations have been documented extensively (Meehl et al., 2013). The CESM1 (CAM5) used in this study is a version with a finite volume nominal 1-degree horizontal resolution (0.9º by 1.25º) and 30-level vertical resolution. The highest model level is about 36 km (4 hPa) in the stratosphere, and lower levels close to surface (boundary layers) have vertical resolutions of about 100-200 m.

CESM1 (CAM5) includes forcings from greenhouse gases (GHGs) as well as concentrations of tropospheric ozone and stratospheric ozone (Lamarque et al., 2010). The concentrations of various gases were calculated off-line and prescribed into model simulations, unlike the aerosol loading calculated online from the emissions. The three-mode modal aerosol scheme (MAM3) has been implemented (Liu et al., 2012) and provides internally mixed representations of number concentrations and mass for Aitken, accumulation, and coarse modes of various aerosol species (sulfates (SO$_4$), BC, organic carbons, dust, sea salt). The new cloud microphysics scheme (Morrison and Gettelman, 2008) allows the number concentration of cloud drops and ice crystals to be affected by aerosol concentrations and therefore accounts for the "indirect radiative forcing" of aerosols.

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The land model (Community Land Model, CLM4) also includes major updates, making it more versatile in simulating snow packs (Lawrence et al., 2011). The sub-grid processes including melting, metamorphism, deposition and redistribution are considered in a snow cover fraction parameterization (Niu and Yang, 2007). Other parameterizations include snow compaction (Lawrence and Slater, 2010) and the albedo calculations for snow on or around vegetations (Wang and Zeng, 2009). Compared to the previous model versions, the albedo contrast between snow-covered and non-snow-covered area is more consistent with observations.

2.2 BC treatment in the model

(a) BC effects on surface albedo. The deposition of BC particles, due to gravity or rainfall removal, is a mechanism to remove aerosols from the atmosphere, and therefore a sink term for the atmospheric BC mass balance. BC particles deposited onto surface of high-albedo snow or ice would reduce surface albedo. The snow model of CLM4 is significantly modified via the incorporation of SNICAR (Snow and Ice Aerosol Radiation)

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module, which represents the effect of aerosol deposition (BC, organic carbon and dust) on albedo, introduces a grain-size dependent snow-aging parameterisation, and permits vertically resolved snowpack heating (Flanner et al., 2007). This new module considers the albedo change by counting the surface concentration of BC and it calculates the surface radiative energy flux at multiple wavelengths. The surface albedo change will consequently alter the energy balance at the surface and in the atmosphere.

(b) BC atmospheric radiative forcing. The present-day BC emission is adjusted from the standard model emission inventory (Lamarque et al., 2010) to account for the potential model underestimation of BC forcing. Emissions over East Asia regions are increased by a factor of two and South Asia regions by four. The emissions are adjusted by the same ratio in all economic sectors (energy, industrial, etc.) and all seasons by the same ratio. Our previous analysis showed that such a correction would improve model-simulated radiative forcing compared with direct observations (Xu et al., 2013; Xu, 2014). Without the observationally constrained values, the modeled forcing and simulated temperature change would be lower by about a factor of two to four.

2.3 Model experiments

To isolate the climate impact of individual forcing agents, we contrasted the perturbed model simulations with present-day forcing (b) to the long-term pre-industrial control simulations (a). The approach is similar to the classical instantaneous CO_2 doubling experiment (Manabe and Wetherald, 1975). Additionally we conducted fixed-SST experiment for radiative forcing diagnostics (c) and the 20th century transient runs to better attribute the observational changes (d). The details of these simulations are given below.

(a) Control simulation for pre-industrial climate. We have a 319-year-long pre-industrial control run, and extended it with an additional 75-year run to test if there was any discernible drift in the mean climate state. The Northern Hemisphere temperature does not show any statistically significant drift. Therefore, we lay the foundation for our analysis by employing the original 319-year run and the extended 75-year run (394 years in total) as a control case. Natural variability of the climate system can be examined from the unforced 394-year pre-industrial simulations.

(b) Four sets of perturbed simulations with instantaneously imposed present-day forcing: BC, SO_4, CO_2 and all three forcing combined. The forcings were imposed by instantaneously increasing the emissions of BC, or the emission of SO_4's precursor sulfur dioxide, or by increasing CO_2 concentration to present-day level (400 ppm).

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Except for the adjusted BC present-day emission as detailed in section 2.2(b), all other emissions are from the standard inventory adopted by CMIP5 models, as described in Lamarque et al., (2010).We run the perturbed simulations in fully coupled mode for 75 years, starting from the end of the 319th year of the control simulation. The difference between the last 60 years (allowing the first 15 years for model spin-up) and the long-term control simulation provide the response signal due to the imposed forcing. With the concern that BC signal is potentially small compared with natural variability, five ensemble members of BC forced simulations are conducted to increase signal-to-noise ratio. Each model year costs about 2000 processor-hours in a high-performance computing system.

(c) Three sets of perturbed simulations but with fixed sea surface temperature. These are also forced by the instantaneous increase of BC, SO_4 and CO_2, separately, but the model runs in atmosphere and land only mode with sea surface temperature fixed at pre-industrial level. These simulations are used only for diagnosing the radiative forcing due to various species.

(d) The 20th century transient single-forcing simulations. The simulations as part of CMIP5 experiments were conducted using the same model configuration as above, except with time-evolving transient forcing of individual species (All forcing, GHGs, aerosols, and BC). Three ensemble members are available for each single forcing run. In addition to the standard BC runs, we also conducted a new BC single forcing simulation with adjusted BC emission factor as described in section 2.2(b).

2.4 Observations

The key model output in this high altitude region to be compared with observations is surface temperature and snow cover. For temperature trend, we adopted both in-situ data recorded at meteorological sites as reported in previous studies (a) and a high-resolution temperature reanalysis dataset (b). For snow cover, we adopted a long-term dataset (c) as well as the direct satellite measurement but only dated back to 2000s (d). The details of these observational dataset are below.

(a) Ground-based temperature record. Monthly mean daily-minimum temperatures from 116 weather stations in the eastern Tibetan Plateau and its vicinity (with elevations ranging from 300 m to 5000 m) during 1961–2006 are reported in Liu et al. (2009). Liu et al. (2009) only analysed daily-minimum temperature because a well-recognized feature associated with climatic warming is less warming observed in maximum temperatures and

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substantially more warming in minimum temperatures (Easterling et al., 1997). Previous studies also show such asymmetric changes in maximum and minimum temperatures are particularly true for the Tibet (Liu et al., 2006) and the Alps (Weber et al., 1997).

(b) Surface temperature reanalysis dataset. Global Historical Climatology Network (GHCN) is a high-resolution (0.5° by 0.5°) analysed global land surface temperatures from 1948 to near present (Fan and van den Dool, 2008). The dataset uses a combination of two large individual data sets of station observations collected from the Global Historical Climatology Network version 2 and the Climate Anomaly Monitoring System. Data are downloaded from http://www.esrl.noaa.gov/psd/data/gridded/data.ghcncams.html.

(c) NOAA Climate Data Record of snow cover extent (Robinson et al., 2012). Prior to 1999 the NH snow cover extent is based on satellite-derived maps produced weekly by trained NOAA meteorologists. After 1999 NOAA NH snow cover extent maps were replaced by output from the Interactive Multi-sensor Snow and Ice Mapping System (IMS) processed at Rutgers University. Data are downloaded from http://climate.rutgers.edu/snowcover/docs.php?target=datareq.

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(d) Moderate Resolution Imaging Spectroradiometer (MODIS) snow cover observations. The MODIS snow products are an 8-day global-gridded product The MOD10CM product is a climate modeling grid product at a 0.05° resolution with global coverage and monthly availability. Pixel values depict the percentage of snow cover (Hall et al., 2006). For the period March 2000 to December 2006, the algorithm version 4 is used and after that version 5 of the algorithm is used. Snow cover products derived from MODIS are based on a ratioing of MODIS band 4 (green) (0.545–0.565 µm) and band 6 (near-infrared) (1.628–1.652 µm). Data are downloaded from http://nsidc.org/data/MOD10CM.

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3 Observed snow cover reduction linked with BC

The observations (Robinson et al., 2012) show that the snow cover extent over the Himalayan mountain range has declined at a rate of more than 10% per decade since the 1960s (Fig. 1). The snow cover retreat along the Himalayan mountain range is greater than in Eurasia during the same period. In-situ studies on regional glaciers and snow pack also reported strong declining trends. For example, Ma and Qin (2012) used 754 stations in China to document statistically significant declining trends of spring snow for the Qinghai-Tibet Plateau for the 1951-2009 period. Consistently, permafrost degradation has been reported on the Tibet Plateau (Cheng and Wu, 2007; Li et al., 2008).

Several satellite observations since the year of 2001 provided additional record in snow cover extent. The observed trend over this shorter period (2001-2012) is less significant (Fig. S1a) and negative trends are only found along some portion of Himalaya range. Consistently, the 5-km MODIS dataset (Fig. S1b) also shows that the snow cover extent averaged over the entire Tibet region only has a slight decrease. But as other studies have pointed out (Pu et al., 2007; Pu and Xu, 2009), the highest altitudes of 5750–6000 m exhibits larger negative trends (−6%/decade). We note that at shorter time-scale (10 years), the snow cover trend is heavily influenced by natural variability and less significant. For example, during 1980-1991 (Fig. S1c) or during 1990-2001 as shown in Fig. 5 of Menon et al. (2010), the declining trends are much larger. Nevertheless, the declining trend in the 40-50 year timescale (Fig. 1) is more robust and warrants further investigation on its causes, which is the main objective of this study.

To understand the causes of the observed trends of snow reduction over the multi-decadal timescale, we conducted global climate model simulations, in which BC emissions, CO$_2$ concentration or SO$_4$ emissions are increased instantaneously from pre-industrial to present-day levels. Fig. 2 (left column) shows the simulated change of snow fraction due to the increase of BC, CO$_2$ and SO$_4$ aerosols. The pattern of snow cover decline in the BC model simulation captures the broad features of the observed decline (Fig. 1), with the largest snow reduction along the mountain range. The Tibetan Plateau on average showed a reduction in snow fraction of 1.9% due to BC. The snow fraction shrinks by 2.9% due to present-day CO$_2$. Along the Himalayan mountain range, where the near-permanent snow cover exists, the reduction of snow fraction exceeds 10% in both BC and CO$_2$ cases.

Menon et al. (2010) attempted to simulate the snow reduction trends during 1990s but the spatial distribution of the observed trend was not well captured mainly due to the coarse resolution of the model. Qian et al. (2011)

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also acknowledged their model's limitation in representing the snow cover climatology and therefore may have biases in estimating BC impact on snow. It is well known that global models tend to overestimate the snow cover of the Tibetan Plateau, and one potential reason is that the blocking effect for the moisture transport crossing the Himalayas is too small due to the coarse resolution of the global models and too much snowfall is simulated (Ménégoz et al., 2013). This limitation can partly be overcome with models using higher spatial resolutions. The modelling work presented here is a major step forward in terms of spatial resolution (about 1° by 1°), as opposed to earlier studies [2.8° by 2.8° in Flanner et al. (2009) and Qian et al. (2011); and 4° by 5° in Menon et al. (2010)], which helps better resolving the complex topography in this region. As a result of increased spatial resolution and also the improved land scheme, the biases in snow cover simulation is significantly reduced from its earlier model versions [Lawrence et al., (2011), also contrast Fig. S2c with Fig. 2 of Qian et al., (2011)]. However, we note that the precipitation over the Tibet Plateau is still overestimated (Fig. S2b), and future studies, especially using regional climate models with even higher resolutions, are needed to improve the fidelity of model simulations of snow pack and glaciers over this topography-complicated region.

One consequence of the snow cover reduction is the decrease in surface albedo, which provides a positive feedback mechanism to localized warming. Such a surface albedo change in response to sea ice loss has been observationally detected (Kay and L'Ecuye, 2013; Pistone et al., 2014) and is important in explaining amplified Arctic warming. Flanner et al. (2011) also used observations during recent decades to calculate the surface albedo feedback in Northern Hemisphere large-scale snow-covered regions. Our simulations show that surface albedo over the Tibet region decreased by over 2% (Fig. 2, right column) in response to BC. The maximum reduction occurs right along the Himalayan mountain range and part of the Tibet-Sichuan mountain regions.

The surface albedo decrease due to CO_2 shares a similar spatial pattern with BC (Fig. 2, right column) but with a smaller magnitude (Table 2b). Moreover, the snow depth reduction in response to CO_2 is only 30% of that due to BC (Table 2c and Fig. 2), and this highlights the larger effect of BC in causing the regional cryospheric change over the Himalayas and Tibet. Not surprisingly, in the simulations the snow cover and surface albedo are increasing in response to cooling aerosols like SO_4 (Fig. 2), but in a smaller magnitude than that of the decreases due to BC and CO_2.

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4 Warming at high altitudes enhanced by BC

The Tibet region has witnessed increasing surface temperature by 0.3ºC per decade—more than twice the global average (Wang et al., 2008). One feature of the surface-warming trend over Tibet is that the warming magnitude increases significantly with altitude (Liu et al., 2009). To understand this anomalous feature, we show in Fig. 3 the tropospheric temperature responses (as a function of altitude and latitude) to BC as well as CO_2 and SO_4. BC-induced heating rate (Fig. 3a) is more concentrated over the northern hemisphere (NH) due to larger emissions there from industrial activities, consistent with radiative forcing distribution (Table 1). The notable feature of BC response is the elevated warming at altitudes of 4000 to 8000 m and 30 to 60 ºN (Fig. 3b), in particular over the Tibetan Plateau. The CO_2 warming pattern (Fig. 3b) features an amplified warming at the surface of the Arctic and in the upper tropical troposphere. CO_2 induced warming in the upper atmosphere (500 hPa) over Tibet is 1 ºC, larger than BC induced warming of 0.5ºC, but the vertical gradient is much smaller (Fig. S3). SO_4 cooling features an even stronger north-south asymmetry (north cooling; south slightly warming) but is more confined to the surface (Fig. S3). The temperature response in the troposphere is associated with strong meridional circulation change. The mechanisms behind the free atmosphere circulation change, especially for the SO_4 case that does not have strong atmospheric forcing, are discussed in details in Xu and Xie (2015).

Ground based observations have shown that the last three decades were subject to a factor of two greater warming in the high-altitude interior of the Tibetan Plateau than at the edge of the plateau and at lower altitudes. The observations in Liu et al. (2009) were made between 1965 and 2006 from ground meteorological stations on the Tibetan Plateau region, and they revealed clear altitude dependence in the daily-minimum surface temperature (purple line in Fig. 4). The vertical profile of temperature change based on daily-average measurement from another reanalysis dataset (Fig. 5) also reveals similar altitude dependence. What's driving the larger warming at high-altitude regions?

Fig. 4 shows the model-simulated change of the daily-minimum surface temperature as a function of elevation due to three different forcing agents (CO_2, SO_4 and BC). The surface temperature responses are calculated from all of the model grid cells over the Tibetan Plateau and the surrounding region (20–50ºN, 70–110°E) to capture the altitude variation in this region. As shown in Fig. 4, the altitude dependence of the surface warming is mostly determined by the response to BC forcing (red dots). At altitudes below 1000 m the warming is minimal, but with increasing altitudes the magnitude of the warming increases up to 2ºC at 5000 m. The

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dependence of the surface warming on altitude is much smaller in the CO_2 case, which only increased from 1.2ºC warming at low altitudes to 1.6ºC warming at higher altitudes (yellow dots).

The combined temperature response (black triangles in Fig. 4) by adding the individual trends due to BC, CO_2 and SO_4 is largely consistent with the observed trend. To test the additivity of the temperature response, we conducted another simulation in which all of the three forcings were imposed simultaneously. The warming profile simulated by the combined anthropogenic forcing experiment largely agree with the sum of the individual responses within 30% (Fig. 5). Some non-linearity is expected as discussed in other modelling studies (Ming and Ramaswamy, 2009). The agreement of the simulated and the observed warming profiles provides a qualitative estimate of the relative contributions of BC, CO_2 and SO_4. Over the entire Tibetan Plateau, CO_2-induced surface warming is 1.3ºC, compared to the BC-induced warming of 0.84ºC (Table 1b). Almost half of the surface warming at the highest altitudes (around 5000 m) is due to BC.

A potential complexity arises due to the internal variability of the climate systems, which has been shown to be important in determining decadal trends over individual regions (Deser et al., 2012). To examine the role of natural variability, we calculated the temperature trend from all 350 consecutive 45-year period out of 394 years simulations. The 80% (10th to 90th percentile) probability range of temperature change is shown in Fig. 5 (light blue shading). The magnitude of the warming rarely exceeds 0.5ºC in any 45-year period in the long-term pre-industrial control simulations without any external forcing. Therefore we infer that the vertical gradient of the temperature trend found in the simulations is very unlikely due to natural variability. One further concern is that the internal variability deduced from the long-term pre-industrial control simulation can be model dependent. However, comparing a 30-member ensemble of CESM1 simulation (the same spatial resolution and configuration as in our control simulation) with the 38-member CMIP5, Kay et al., (2015) found that the ensemble spread in the 30-year trend of surface temperature in CESM1 ensemble is statistically the same with the spread in trends within CMIP5.

Note that the climate simulations shown in Fig. 4 and Fig. 5 are driven by the instantaneous increase of present-day forcing. Since the real forcing trends were time dependent, we further analysed a set of 20th century transient simulation output from the same model (Fig. 6), as part of CMIP5. The relative contributions of CO_2 and SO_4 to the simulated warming are consistent between the two sets of simulations (the instantaneous forcing and transient forcing). But the trends estimated from the 20th century transient forcing simulations are smaller than the quasi-equilibrium response to instantaneous forcing (parenthesis in Table 1b). The reason is that only 70%

of SO_4 forcing and about 60% of CO_2 forcing in the transient simulation were applied after 1960. The standard BC forcing (red solid line in Fig. 6) only lead to a weak warming, not exceeding the range of natural variability. As a result, the combined all-forcing responses (black triangles) did not capture the altitude dependence in observations very well.

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Only when we adjusted historical BC forcing using the same scaling factors constrained by present-day observations, transient BC forcing induced a robust warming and amplification over high altitudes (red dots in Fig. 6), similar to what's shown in instantaneous forcing experiment. However, note that the historical time dependence of the BC forcing is more uncertain, and also we were only able to produce one ensemble of adjusted BC simulation, more subject to the influence of decadal variability. Therefore, the response to the instantaneous present-day BC forcing seems a more reliable indicator of the BC effects. While the absolute values of warming profile needs more model tests, our inference regarding the relative role of BC and CO_2 in the observed decrease of snow cover as well as the major role of BC on the altitude dependence of the warming trends is robust.

5 Physical mechanisms of elevated warming due to BC

Both CO_2 and BC contribute to the elevated warming at 5 km as shown in Fig. 4. However, BC is mostly responsible for the vertical gradient of the simulated warming trend. Most of the BC aerosols in the region are emitted over India and China and subsequently transported to the Tibetan Plateau and the Himalayan mountain range. The physical mechanisms for the amplified warming at higher altitude due to BC are at least three-fold:

(1) Direct heating in the atmosphere.

BC absorbs a significant amount of solar radiation, as much as 25% in typical pollution events as directly measured by multiple unmanned aircrafts over the Northern Indian Ocean (Ramanathan et al., 2007). The BC layer placed at higher altitude is even more efficient in absorbing solar radiation than at sea level, due to stronger solar radiation and the brighter underlying cloud surface. In our model simulation, the BC atmospheric heating rate is concentrated in the Northern Hemisphere (maximum at 30ºN), coincident with the location of the maximum temperature change (Fig. 3). The elevated BC layer, due to the topography of the Tibetan Plateau and the Himalayan mountain range, contributes to the elevated heating which is more than 0.1 ºC/day and about 0.03 ºC/day at 4 km (Fig. S3a). Such anomalous heating in the atmosphere over the elevated regions will contribute to the loss of ice and snow in two ways: (a) it will increase melting of the glaciers and snowpack; and (b) more of the precipitation will fall as rain instead of snow. CO_2 increase also induces longwave heating of the atmosphere (Fig. 3c), but it is well known that warming enhancement at the upper tropical troposphere is mostly due to moist convection processes (Manabe and Wetherald, 1975) and the warming enhancement at high altitudes is not showing sharp gradient as in BC case (Fig. S3b and Fig. 4).

(2) Surface darkening by BC deposition.

Snow and ice have a high surface albedo and reflect as much as 50 to 90% of incoming solar radiation. Transported BC aerosols over the Himalayas and the Tibetan Plateau are removed from the atmosphere due to precipitation. When BC aerosols are deposited over the snow and ice, they increase the absorption of solar radiation and cause surface warming (Wiscombe and Warren, 1980; Chýlek et al., 1983). Recent studies have also suggested the influence of BC aerosols over regions like the Alps (Painter et al., 2013) and Eurasian land (Flanner et al., 2009). Menon et al. (2010) found that when the model includes snow albedo change due to BC the snow cover reduction is twice as large as the simulation with BC atmospheric heating effect only. Flanner et al. (2009) also suggested that BC surface albedo darkening effects are important in causing Eurasian springtime snow-cover decline and are comparable to that of CO_2.

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The surface radiative forcing due to BC deposition over Tibet in this model is estimated to be 4.6 W/m^2 based on the 5-year fixed SST (sea-surface temperature) simulation (Fig. S4b). Because of this strong positive surface forcing associated with surface darkening, the shortwave forcing due to BC at the surface increased from -1.5 W/m^2 (initially due to BC dimming effect) to a positive value of 3.1 W/m^2. This positive forcing imposed directly at the surface is even larger than the adjusted atmospheric heating due to BC over Tibet (1.6 W/m^2). A recent modelling study (Ménégoz et al., 2014) also examined the role of BC deposition over snow in this region (with smaller forcing estimates of 1 to 3 W/m2), but their study did not estimate the atmospheric heating effect of BC.

However, we note that model estimates of radiative forcing due to BC deposition on snow have large uncertainty. Using the same atmospheric and land model (but driven by realistic meteorological field in the year of 2000), Zhang et al., (2015) showed that simulated BC concentration in snow is biased high with respect to in situ sampling. Although the large spatial variations in BC deposition can affect the representativeness of BC-in-snow measurements for the model evaluation purposes, this potential model bias should be kept in mind. In addition to the uncertainty in BC loading, the forcing magnitude is also sensitive to model parameterization (Yasunari et al., 2013), and also the simulated background snow cover because the wrongly simulated melting dates of the snowpack can lead to an incorrect radiative forcing (Jacobi et al., 2015). Therefore, both in-situ (Wang et al., 2013; Zhao et al., 2014) and laboratory measurements (Hadley and Kirchstetter, 2012) are needed to constrain model representation of BC in snow.

(3) Snow albedo feedback. The melting snow in response to the two initial heating mechanisms discussed above will further decrease surface albedo and increase solar absorption at the surface. The results based on the 60-year coupled model simulation suggest that the surface albedo will further decrease by 1.4% and effectively impose an additional 3.2 W/m^2 shortwave forcing at the surface. In summary, the elevated heating and surface darkening due to BC are simultaneously causing local warming and snow melting. The snow cover reduction further reduces surface albedo and then provides a positive feedback. A look at the seasonality of snow depth change suggests the early spring melting is important for this feedback. The net result of such a positive loop is an amplification factor of four for BC-induced Tibet warming from the global average values and significant snow and ice retreat.

Beyond the three main factors as we have discussed above, the changes of water vapour and clouds are also possible mechanisms contributing to the elevation-dependent warming in the mountain regions. As shown in the

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schematic of a recent review paper (Mountain Research Initiative EDW Working Group, 2015), in a warmer and moister atmosphere, the latent heat release at the cloud condensation level may induce larger warming in high altitudes (cloud feedback) and the downward longwave radiation increase particularly fast in higher and drier atmosphere (water vapour feedback). It is difficult to identify or separate the contribution of these individual feedbacks from our current experiment setup. However, we note that these feedback mechanisms are operating regardless of forcing agents and therefore cannot explain the particularly large elevated warming in response to BC.

Lastly, it is also worth commenting the role of other snow impurities. In this study we used BC, a strong solar radiation absorber, to understand the climate response and the mechanisms due to absorbing aerosols that also include dust (Painter, et al., 2007; Di Mauro et al., 2015; Gabbi et al., 2015) and organic aerosols (Qian et al., 2015). Similarly, we used SO_4 to characterize all other reflecting aerosols. Any changes of dust and organics may induced changes to the snow cover, as their atmospheric heating and surface deposition are readily captured by this model, although the magnitude of response might be smaller since they are partially reflecting as well.

6 Conclusions

The observed surface warming over the Tibetan and Himalayan region of about 0.5ºC at sea level to about 2-2.5ºC at 5000 m (from 1961 to 2006) has been an outstanding feature of climate trends. The more than 2ºC warming is close to the peak warming trend observed anywhere on the planet. For comparison, the Arctic warming associated with large sea-ice retreat during this period is 1.2ºC.

The high-resolution coupled ocean-atmosphere model in this study was able to attribute the observed warming trends and their high altitude enhancement to imposed increases in CO_2, BC, and SO_4 aerosols. The simulated changes with all forcing imposed were consistent with the observations. The key to the success is that we obtained the BC forcing from the reconstruction of ground-based and satellite-based observations. The imposed BC forcing was about two to four times (depending on the regions) larger than that simulated by the models using bottom-up emission inventories. The analysis of model simulations highlights that the high-altitude warming due to BC is as large as CO_2 warming over the Tibetan Plateau and the elevated warming profile is unique in BC responses.

The observed record warming is accompanied by retreat of glaciers and snow cover as well as thinning of the snow packs. In response to the pre-industrial to the present-day increase in BC emissions, the annual averaged snow fraction over the Tibetan Plateau is reduced by more than 6% (relatively), and the snow depth by approximately 19%. The surface albedo decreases by more than 5% along the Himalayan mountain range and 1.4% over the entire Tibet, providing a positive local feedback to the enhanced local warming. In stark contrast, despite having five times larger effect in global mean temperature than BC, over Tibet CO_2 impact is only 1.5 times stronger in snow cover decrease, and only one-third in snow depth decrease. We conclude that BC is instrumental in causing snow retreat and its effects are manifested simultaneously through a three-fold process: (i) direct atmospheric heating; (ii) darkening of the snow surface and (iii) the snow albedo feedback. It is important to note that, without the scaling factor we applied to bring the model BC forcing to the observationally constrained values, the impact of BC on the observed temperature trends would have been marginal. This perhaps explains why the models used in IPCC assessments have not simulated the role of BC in the large warming trend over the Himalayas.

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Table 1. (a) TOA (top-of-atmosphere) radiative forcing (W/m^2, shortwave + longwave), due to BC (direct radiative forcing; pre-industrial to present-day; not including snow albedo effect), CO_2 (pre-industrial to 400 ppm), and SO_4 (direct and indirect effect, so-called “adjusted forcing”; pre-industrial to present-day). The radiative forcing is calculated by running the atmospheric model with fixed sea-surface temperature for 5 years. The domain of the Tibet Plateau is 30 to 40ºN and 80 to 100ºE.

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(b) Surface temperature change (ºC) in response to different forcings in (a). Surface temperature change is calculated by averaging the last 60 years of a 75-year coupled model simulation. The values in parenthesis are temperature change in the 20th century time-dependent forcing simulations (1960-2005). The linear trend (ºC/decade) is first calculated and then multiplied by 4.5 to obtain the change with 45-year time frame. BC responses include the range of using "standard" and adjusted emissions.

(a) TOA net forcing (W/m^2)	BC	CO_2	SO_4
Global	0.5	1.7	-0.9
NH	0.7	1.7	-1.5
Tibet	1.1	0.6	-0.3

(b) Surface temperature change (ºC)	BC	CO_2	SO_4
Global	0.21 (0.04-0.15)	1.2 (1.0)	-0.5 (-0.4)
NH	0.29 (0.06-0.21)	1.3 (1.2)	-0.7 (-0.5)
Tibet	0.84 (0.22-0.69)	1.5 (1.0)	-0.7 (-0.3)

Table 2. (a) Snow fraction (%), (b) surface albedo (%) and (c) snow depth over land (water equivalent, cm) change in response to different forcings. The relative change as a percentage is shown in parenthesis next to the absolute change.

(a) Snow fraction (%)	BC	CO_2	SO_4
Global	-0.13 (-2%)	-0.35 (-4%)	0.14 (2%)
NH	-0.26 (-3%)	-0.67 (-7%)	0.36 (4%)
Tibet	-1.9 (-6%)	-2.9 (-9%)	1.65 (5%)

(b) Surface albedo change (%)	BC	CO_2	SO_4
Global	-0.2 (-1%)	-0.68 (-4%)	0.28 (2%)
NH	-0.3 (-2%)	-0.79 (-5%)	0.44 (3%)
Tibet	-1.4 (-2%)	-1.1 (-2%)	1.1 (2%)

(c) Snow depth (cm)	BC	CO_2	SO_4
Global	-0.06 (-2%)	-0.15 (-4%)	0.1 (3%)
NH	-0.11 (-6%)	-0.28 (-14%)	0.2 (10%)
Tibet	-0.2 (-19%)	-0.06 (-6%)	0.16 (15%)

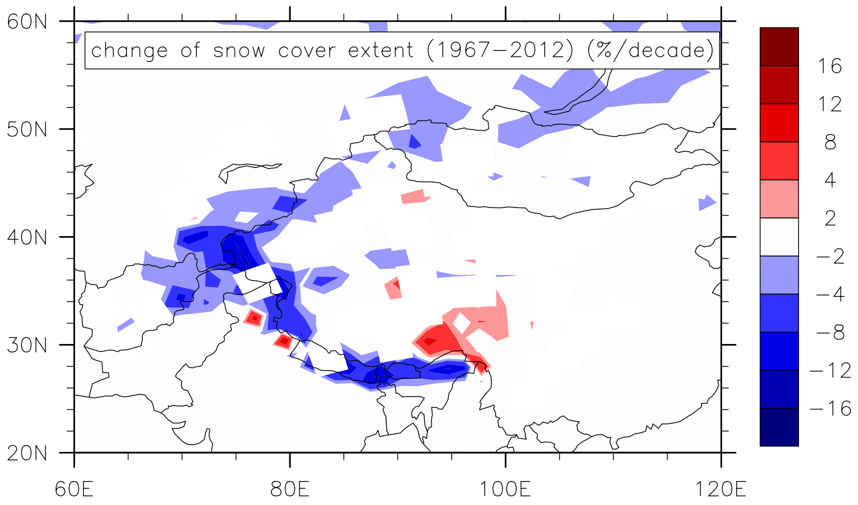

Fig. 1. Observed snow cover extent change (% per decade) from 1967 to 2012. The trend is calculated based on snow cover extent data in the entire period. Insignificant changes (confidence interval <75% calculated from student's T-test) are not shown.

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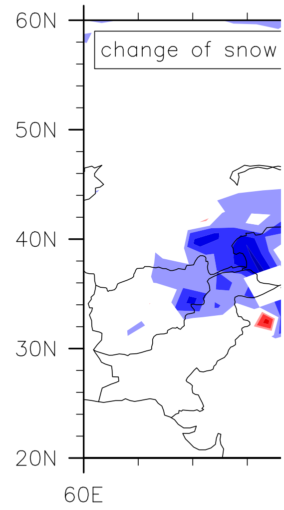

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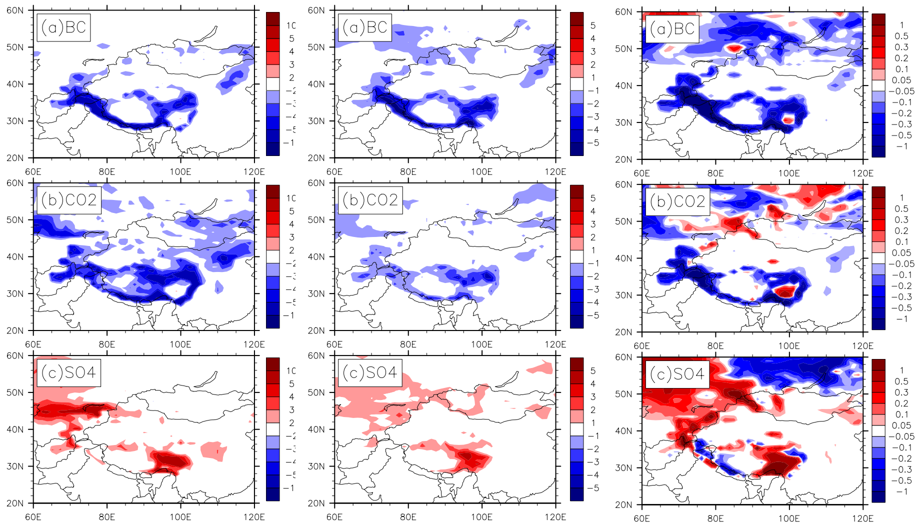

Fig. 2. Left: (a) simulated change of snow fraction (%) due to present-day BC (versus its pre-industrial level), (b) CO_2 and (c) SO_4. Middle: same as left, but for surface albedo (%). Right: same as middle, but for snow depth (water equivalent, cm). The regionally averaged statistics are shown in Table 2.

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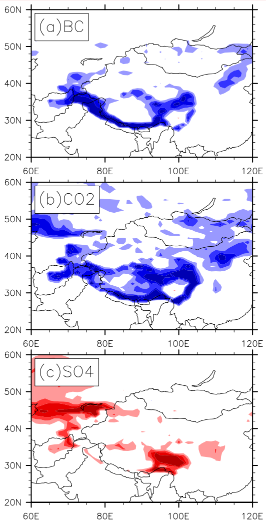

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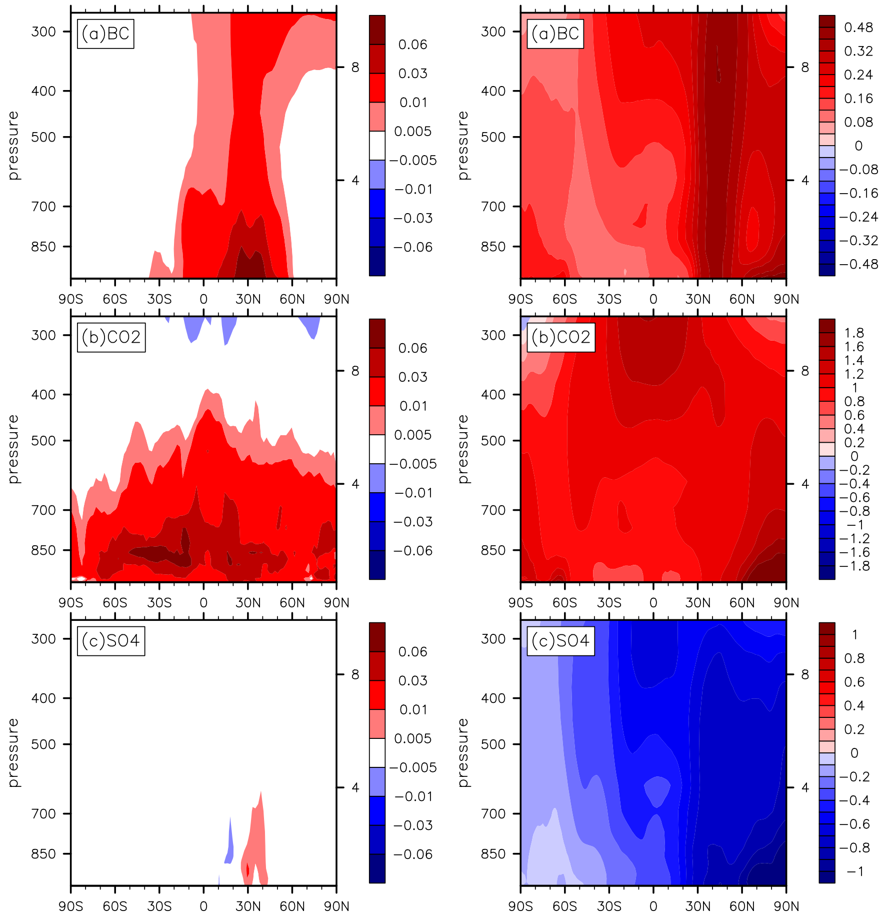

Fig. 3. Left: Globally zonal averaged radiative heating rate (°C/day) as a function of altitude and latitude due to (a) BC, (b) CO$_2$ and (c) SO$_4$, calculated from the 5-year fixed SST simulations using the instantaneous radiative diagnostic procedure. Shortwave fluxes are shown for BC and SO$_4$, and longwave flux for CO$_2$. Right: The temperature response (°C) due to (a) BC, (b) CO$_2$ and (c) SO$_4$, calculated as the difference of the last 60 years of 75-year perturbed simulation and the 319-year long-term control. Fig. S3 shows the normalized heating rate and temperature profile averaged just over the Tibetan Plateau.

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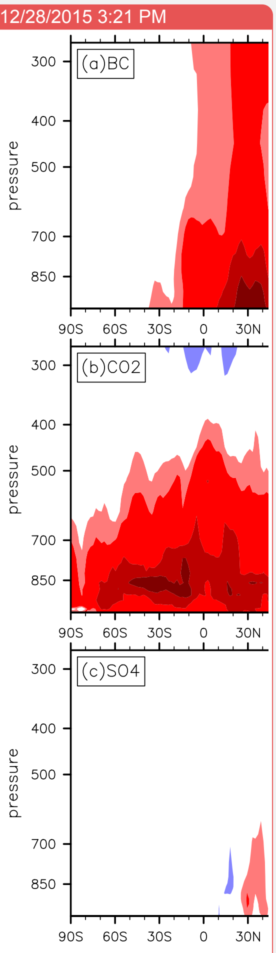

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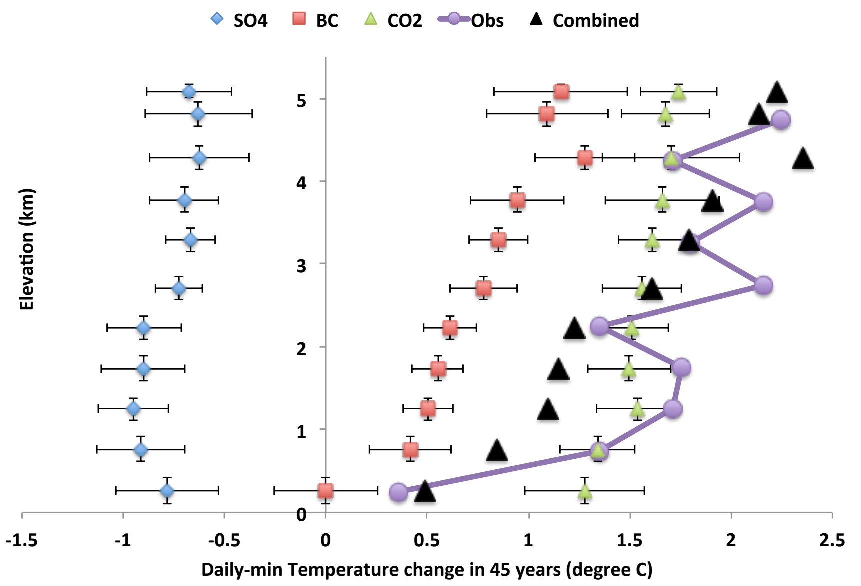

Fig. 4. The change of daily-minimum temperature (ºC) as a function of elevation (km). The observation from 1961 to 2006 are from of Liu et al. (2006). The simulated temperature responses due to instantaneous increase of forcings (CO_2, SO_4 and BC) are calculated from model grid cells over the Tibetan Plateau and its vanity region (20–50°N, 70–110°E) including low-lying regions and high-altitude regions. The standard deviation due to spatial variation of temperature response is shown as error bars. The sum of CO_2, SO_4 and BC responses are shown in black triangles.

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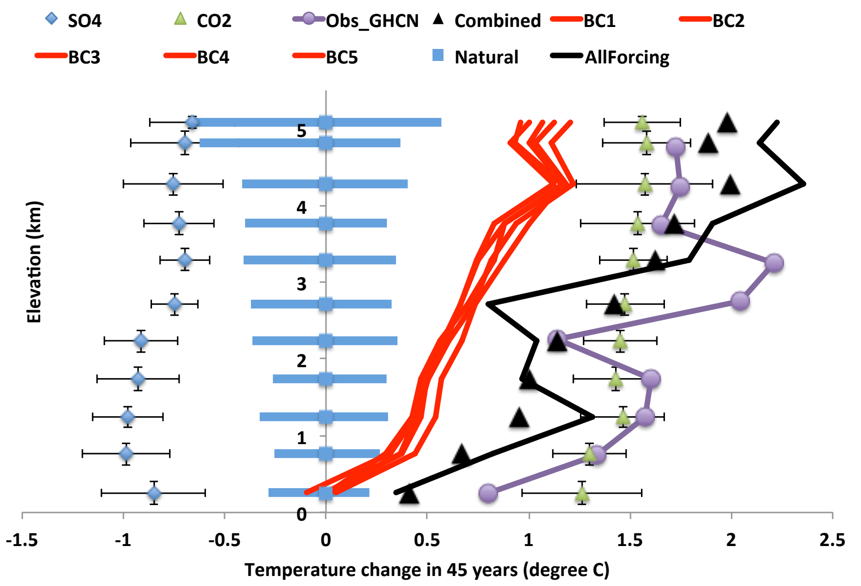

Fig. 5. Similar to Fig. 4, but with the following differences: (1) daily-mean surface temperature, not daily-minimum temperature, are shown; (2) the spread of five-ensemble member of BC simulations are shown in red lines; (3) the observations are from GHCN dataset (1961 to 2006); (4) the range of temperature change found in unforced pre-industrial control simulations is shown in blue shading and (5) the all forcing simulation (black line) is shown in comparison with the sum of individual responses (black triangles).

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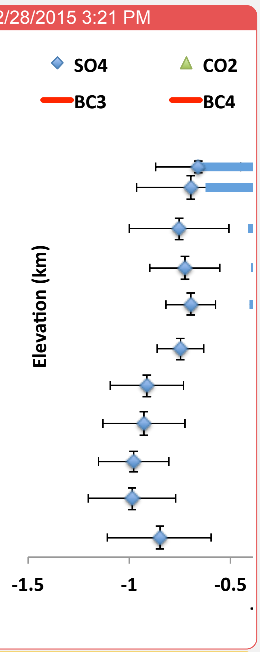

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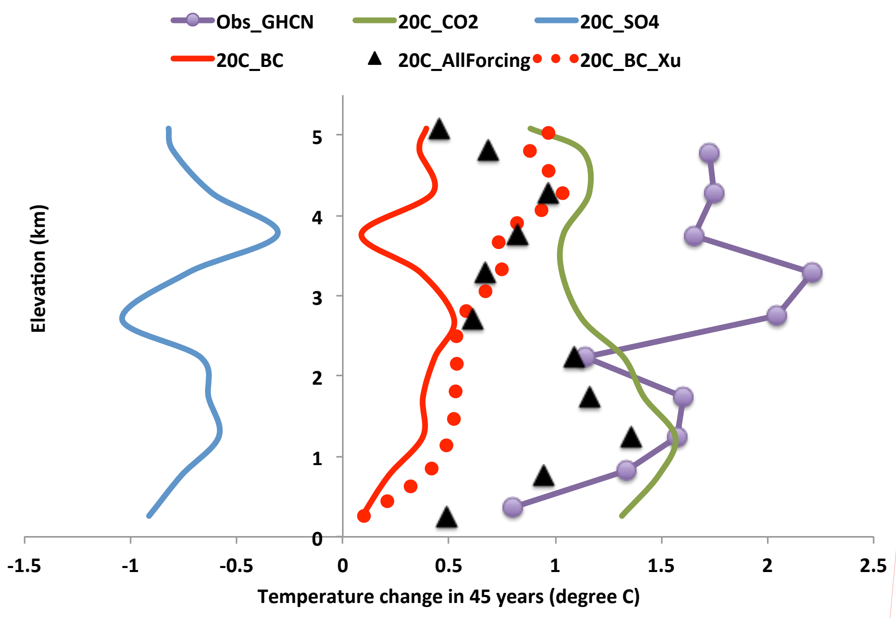

Fig. 6. Similar to Fig. 5 but showing results from the 20th century transisent simulations (3 ensemble members for each single forcing run). Note that the "starndard" BC single forcing simulation used smaller BC emissions as in other CMIP5 models (red solid line). An addtional simulation with adjusted larger BC emissions were shown (red dotted line, one ensemble member only).

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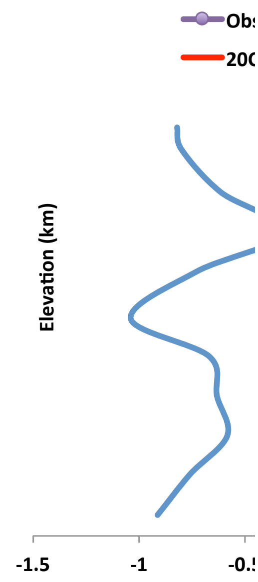

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Supplementary materials

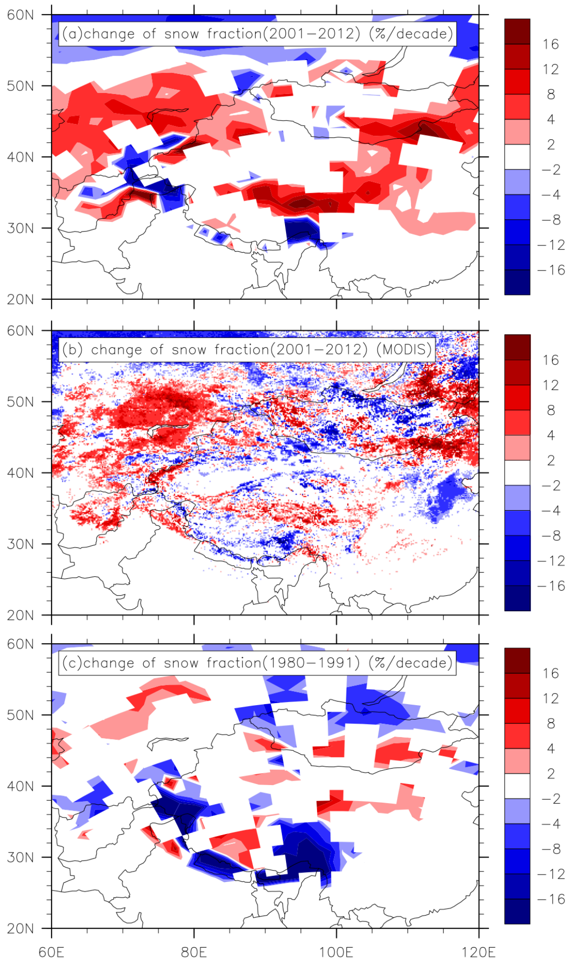

Fig. S1. (a) Snow cover extent same as Fig. 1, but for 2001-2012. (b) same as (a) but from MODIS. (c) same as (a) but for 1980-1991.

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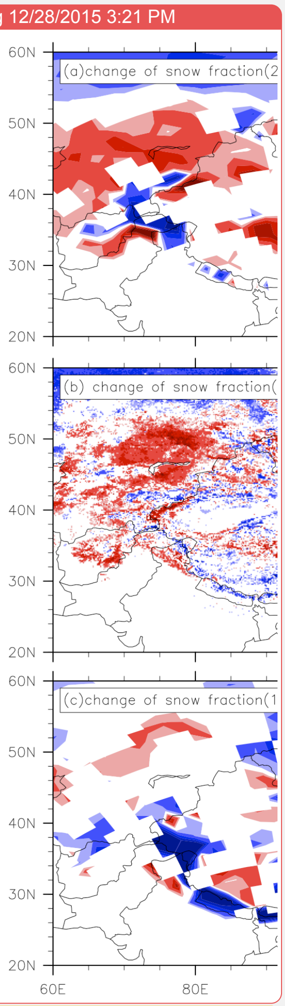

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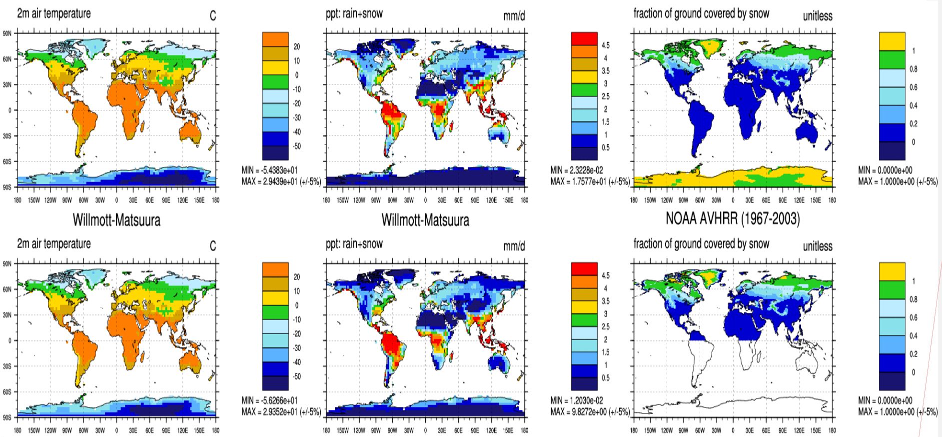

Fig S2. (Left) climatological surface air temperature (ºC) in the model simulation in the top panel, and observed surface air temperature in the bottom panel. (Middle) total precipitation (rain and snow fall) (mm/day) (Right) snow cover fraction. The model results in the top row are the 1981-2005 averages of the transient simulations under all radiative forcing. The temperature and precipitation observations are from updated dataset of Willmott and Matsuura (2001). The snow cover observations are from NOAA AVHRR as compiled by Robinson et al., (2012). In terrain-complex regions (such as North American Rockies, South American Andes and Tibet Plateau), the model tends to overestimate the precipitation and consequently snow cover, a bias commonly found in global climate model with coarse resolutions (Ménégoz et al., 2013). More detailed land model evaluations can be found in Lawrence et al., (2011).

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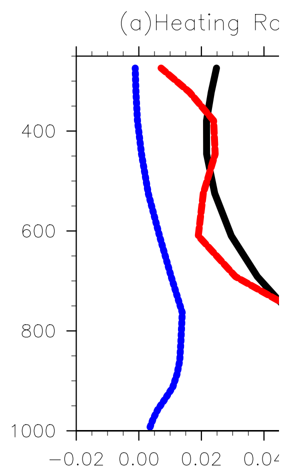

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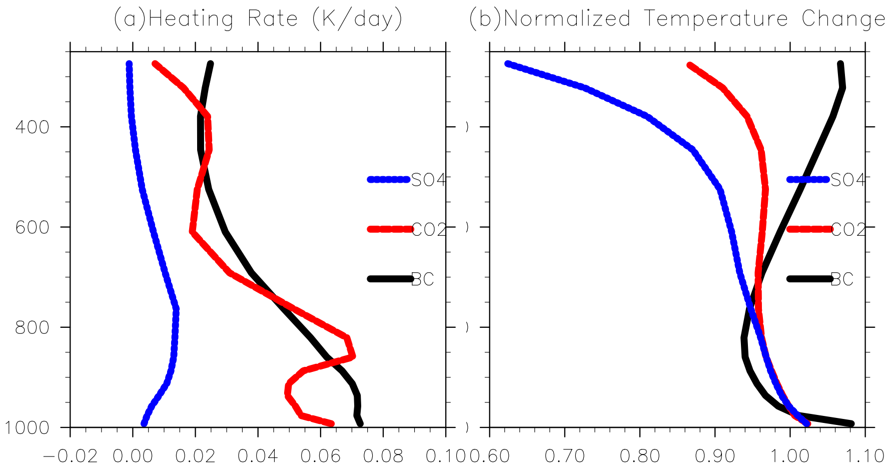

Fig. S3. Similar to Fig. 3 but showing the vertical profile averaged over the Tibet region. (a) Radiative heating rate (ºC/day). Shortwave fluxes for BC and SO_4, and longwave flux for CO_2. (b) Normalized temperature change relative to the average below 900 hPa. Note that the changes are tropospheric atmospheric temperature change, not surface temperature. The domains of the Tibet Plateau (as in Table 1) are 30 to 40ºN and 80 to 100ºE.

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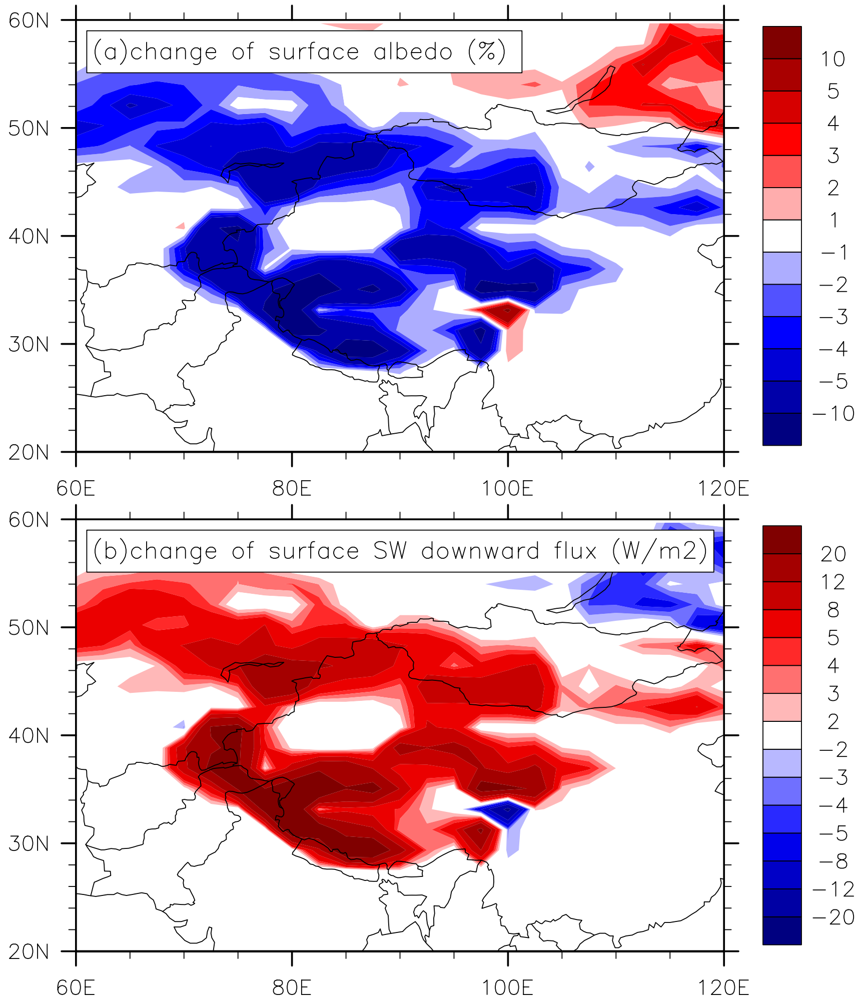

Fig. S4. (a) Change of surface albedo due to BC deposition on snow; (b) Change of net shortwave radiation (downward as positive, W/m^2). Over Tibet Plateau, the surface albedo is reduced by 2.2%, causing an increase in shortwave radiation reaching the surface by 4.1 W/m^2 (heating). Globally, the radiative forcing at the surface is about 0.1 W/m^2. The change of surface albedo in (a) is calculated with the five-year atmosphere-only simulation in which BC emission is increased. Therefore, the albedo change largely represents the surface darkening effects due to BC deposition, although we cannot completely rule out the associated melting during this period. As a result, the actual radiative forcing at the surface due to BC in snow should be smaller than that in (b).

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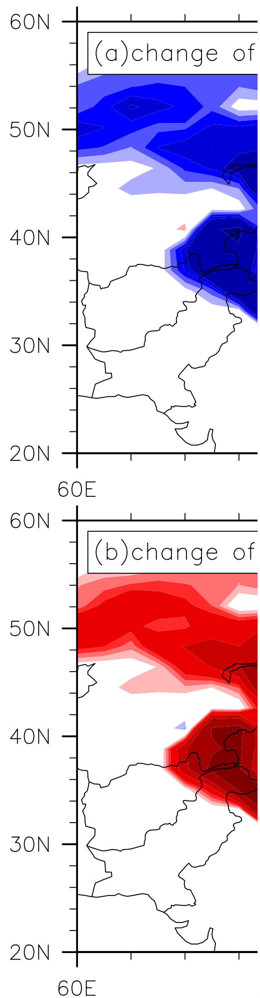

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