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Impact of light-absorbing particles on snow albedo darkening and associated radiative forcing over High Mountain Asia: High resolution WRF-Chem modeling and new satellite observations

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Abstract

Light-absorbing particles (LAPs), mainly dust and black carbon, can significantly impact snowmelt and regional water availability over High Mountain Asia (HMA). In this study, for the first time, online aerosol-snow interactions enabled and a fully coupled chemistry Weather Research and Forecasting (WRF-Chem) regional model is used to simulate LAP-induced radiative forcing on snow surfaces in HMA at relatively high spatial resolution (12 km, WRF-HR) than previous studies. Simulated macro- and micro-physical properties of the snowpack and LAP-induced snow darkening are evaluated against new spatially and temporally complete datasets of snow covered area, grain size, and impurities-induced albedo reduction over HMA. A WRF-Chem quasi-global simulation with the same configuration as WRF-HR but a coarser spatial resolution (1 degree, WRF-CR) is also used to illustrate the impact of spatial resolution on simulations of snow properties and aerosol distribution over HMA. Due to a more realistic representation of terrain slopes over HMA, the higher resolution model (WRF-HR) shows significantly better performance in simulating snow area cover, duration of snow cover, snow albedo and snow grain size over HMA, as well as an evidently better atmospheric aerosol loading and mean LAPs concentration in snow. However, the differences in albedo reduction from model and satellite retrievals is large during winter due to associated overestimation in simulated snow fraction. It is noteworthy that Himalayan snow cover have high magnitudes of LAP-induced snow albedo reduction (4-8 %) in summer (both from WRF-HR and satellite estimates), which, induces a snow-mediated radiative forcing of ~ 30-50 W/m^2. As a result, Himalayas (specifically western Himalayas) hold the most vulnerable glaciers and mountain snowpack to the LAP-induced snow darkening effect within HMA. In summary, coarse spatial resolution and absence of snow-aerosol interactions over Himalaya cryosphere will result in significant underestimation of aerosol effect on snow melting and regional hydroclimate.

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1. Introduction

Light-absorbing aerosol particles (LAPs; airborne dust and black carbon (BC) specks), can impact on regional water availability over Asia in three ways. Firstly, LAPs can directly interact with incoming solar radiation and induce thermo-dynamical modifications to synoptic scale circulations (Hansen et al., 1997 ; Ramanathan et al., 2001; Bond et al., 2013; Lau et al., 2006; Bollasina et al., 2011; Li et al., 2016). Secondly, acting as cloud condensation nuclei, changes in concentrations of these particles can lead to microphysical modification of cloud systems and precipitations (Fan et al., 2016; Li et al., 2016 ; Qian et al., 2009; Sarangi et al., 2017). Finally, deposition of LAPs in the snowpack can also darken the snow, reduce its surface albedo and accelerate snow warming and melting (Warren and Wiscombe, 1980; Qian et al., 2015; Qian et al., 2011; Qian et al., 2009a; Lau et al., 2010; Xu et al., 2009; Hadley and Kirchstetter, 2012; Dang et al., 2017). Modeling studies have suggested that the LAP-induced snow darkening mechanism has warming and snow-melting efficacy even greater than that of greenhouse gases (GHGs) (Hansen and Nazarenko, 2004; Flanner et al., 2007; Qian et al., 2011; Skiles et al., 2012). To give a perspective, the concentration of just 100 ng of BC in 1 g of snowpack will reduce the visible-wavelength albedo of grain radius 1000 μm by 10% (Fig. 1b of Warren, 2013). A chain of positive feedback mechanisms results in such large impact of LAPs (Qian et al., 2015). Initially, as snow starts to melt, the concentration of LAPs in snowpack increases because a portion of LAPs accumulate at the surface of the snowpack instead of getting washed away with meltwater (Conway et al., 1996; Flanner et al., 2007; Doherty et al., 2010). This increase in LAP concentration leads to enhanced warming of the snowpack and thereby increases the effective snow grain size, which further lowers snow albedo (Warren and Wiscombe, 1980; Hadley and Kirchstetter, 2012). Nonetheless, at higher concentrations, grain sizes can again be reduced due to the loss of mass from surface layers with the intense melting (Painter et al., 2013). As this process continues, sufficient snow melt occurs to expose the darker underlying surface, leading to enhanced warming and snow ablation commonly-known as "snow albedo feedback" (Warren and Wiscombe, 1980; Hansen and Nazarenko, 2004; Flanner et al., 2007; Qian et al., 2015). In turn, this earlier loss of snow cover induces surface warming and perturbing regional circulations (Hansen and Nazarenko, 2004; Lau et al., 2010; Qian et al., 2011). This LAP-induced modification of snow albedo feedback is identified as one of major forcing agents affecting climate change with a high level of uncertainty (IPCC, 2013).

High Mountain Asia (HMA) includes the Tibetan plateau, central Asian mountains and the Himalaya cryosphere. It holds the largest glacial cover (~9500 glaciers) outside the polar region (Dyurgerov, 2001). Observations revealed that a historical decadal increase in the surface air temperature over HMA in a range of 0.6-1.8 ^{o}C (Shrestha et al., 1999;Wang et al., 2008), and the warming is faster over higher elevations (> 4000 m) in the last three decades (Xu et al., 2009b; Ghatak et al., 2014). The Himalaya glacier area has cumulatively decreased by ~16% during the period 1962 to 2004 (Kulkarni et al., 2010) and the spring snow cover is decreasing at a decadal rate of ~0.8 million km^2 during the last 50 years (Brown and Robinson, 2011). The average retreat rate on the north slope of Mount Everest is as high as 5.5–9.5 m y^{-1} (Ren et al., 2006). The Himalaya cryosphere contributes to the stream flow in Indus and Ganges river systems by ~ 50 % and ~10-30%, respectively (Khan et al., 2017). Warming and glacier retreat over the Himalaya cryosphere have a great potential to impact the fresh water availability for about 700 million people, modify regional hydrology, and disturb the agrarian economy of all South Asian countries (Bolch et al., 2012;Immerzeel et al., 2010;Kaser et al., 2010;Singh and

Bengtsson, 2004;Barnett et al., 2005;Yao et al., 2007). Therefore, it is critical to disentangle the factors contributing to glacier retreat and snow melt over HMA.

Regional warming due to increasing greenhouse gases (Ren and Karoly, 2006) has been reported as the primary cause of the high rate of warming and glacier retreat over HMA. However, in the last decade, advancement in remote sensing and availability of measurements from several field campaigns suggest that the contribution of LAP loading (in the atmosphere) to the warming and glacier melting over HMA is probably greater than previously believed (Ramanathan et al., 2007; Prasad et al., 2009; Menon et al., 2010). Continuous observations over the Nepal Climate Observatory Pyramid (NCO-P) facility located at 5079 m a.s.l. in the southern foothills of Mt. Everest revealed very high concentrations of black carbon (Marcq et al., 2010) and desert dust (Bonasoni et al., 2010) especially in spring from Indo-Gangetic plains. Atmospheric LAPs are scavenged to the snow/ice surface by dry and wet deposition and cause measurable snow darkening and melting (Gautam et al., 2013; Yasunari et al., 2010b;Yasunari et al., 2013; Nair et al., 2013; Ménégoz et al., 2014; Ming et al., 2008;Flanner and Zender, 2005). Thus, LAP deposited in snow and associated snow darkening has been suggested as a key factor to the early snowmelt and rapid glacier retreat over HMA (Yasunari et al., 2010; Ming et al., 2008; Xu et al., 2009a; Flanner et al., 2007; Qian et al., 2011).

While previous studies have underlined the significance of LAP-deposition in snow over HMA, the estimation of LAP-induced snow darkening and associated radiative forcing is still highly uncertain (Qian et al., 2015). Many of these studies used online global model simulations at coarse spatial resolutions of ~50-150 km (Flanner and Zender, 2005; Ming et al., 2008; Qian et al., 2011). Others used offline simulation of the snow albedo effect using measured concentrations of deposited LAP in surface snow or estimated from atmospheric loading and ice

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

cores (Yasunari et al., 2013; Nair et al., 2013; Wang et al., 2015). The complex terrain of HMA, seasonal snowfall and near surface air circulation are not well resolved by coarse global climate models (Kopacz et al., 2011; Ménégoz et al., 2014). Similarly, offline estimations are limited in scope because they are site specific and are based on simplified assumptions about deposition rates. Ideally, online high resolution simulations allowing for LAP-snow interactions should facilitate a more realistic understanding of LAP deposition to snow and LAP-induced snow darkening effect in terms of both magnitude and spatial variability over HMA.

In this study, a modified version of the online chemistry coupled with Weather Research and Forecasting regional model (WRF-Chem v3.5.1), which, is then fully coupled with SNICAR (SNow, ICe, and Aerosol Radiative) model, is used to perform first-ever high resolution (12 km) simulation over the HMA region for the water year 2013-14 (October 1, 2013 to September 30, 2014). Satellite observations of snow properties like snow albedo, grain size, and LAP-induced snow darkening from MODSCAG and MODDRFS retrievals are used for evaluation (Painter et al, 2009; 2012). The main objective of this study is to evaluate the skill of high resolution WRF-Chem model in simulating properties of snowpack, aerosol distribution, LAP in snow and LAP-induced snow darkening over HMA using spatially and temporally complete (STC) remotely sensed snow surface properties (SSP) from MODIS (Dozier et al, 2008; Rittger et al, 2016). Our second objective is to demonstrate the benefit for aerosol and snow distributions in high resolution runs by comparing to a coarse gridded quasi-global model simulation over HMA. This quasi-global simulation is run with the same WRF-Chem configuration but at 1 degree spatial resolution. Finally, the spatiotemporal variation of simulated LAP deposition, snow albedo darkening and snow mediated LAP radiative forcing (LAPRF) over HMA are discussed. The

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Discussion started: 13 November 2018

model details and datasets used are described in Section 2. Results and discussions are presented in Section 3 followed by conclusions in Section 4.

2: Model simulations and observational datasets

Below, we provide details on the aerosol module used in WRF-Chem, interactive coupling with aerosol and SNICAR via land model, and the model setup for both 12 km and 1 degree resolution runs. The details for the remote sensing observations that are used to evaluate the models are also provided.

2.1: Coupled WRF-Chem-CLM-SNICAR Model description

The WRF-Chem simulation is performed at 12 km × 12 km horizontal resolution (hereafter refereed as WRF-HR) with 210 × 150 grid cells (64–89°E, 23–40°N) (Figure 1) and 35 vertical layers. The simulation was conducted from 20^{th} September, 2013 to 30^{th} September, 2014, to provide one year of results (following a 10-day model spin-up). ERA-interim reanalysis data at 0.7° horizontal resolution and 6 h temporal intervals are used for meteorological initial and lateral boundary conditions. The simulation is re-initialized every 4th day to prevent the drift of model meteorology. Model physics options used are the MYJ (Mellor–Yamada– Janjic) planetary boundary layer scheme, Morrison 2-moment microphysics scheme, community land model (CLM), Kain-Fritsch cumulus scheme and Rapid Radiative Transfer Model for GCMs (RRTMG) for longwave and shortwave radiation schemes.

The CBM-Z (carbon bond mechanism) photochemical mechanism (Zaveri and Peters, 1999) coupled with eight bin MOSAIC (Model for Simulating Aerosol Interactions and Chemistry) aerosol model (Zaveri et al., 2008) is used. This is the most sophisticated aerosol module available for the WRF-Chem model. The sectional approach with eight discrete size bins

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

is used to represent the size distributions of all the major aerosol components (including sulfate, nitrate, ammonium, black carbon (BC), organic carbon (OC), sea salt, and mineral dust) in the model. The processes of nucleation, condensation, coagulation, aqueous phase chemistry, and water uptake by aerosols in each bin size are included in the MOSAIC module. Dry deposition of aerosol mass and number is simulated by including both diffusion and gravitational effects as per Binkowski and Shankar (1995). Wet removal of aerosols follow Easter et al. (2004) and Chapman et al. (2009) and includes grid resolved impaction and interception processes for both in-cloud (rainout) and below-cloud (washout) aerosol removal. Processes involved in convective transport and wet removal of aerosols by cumulus clouds are described in Zhao et al. (2013).

Anthropogenic emissions used in our study is at $0.5^{O} \times 0.5^{O}$ horizontal resolution and are taken from the NASA INTEX-B mission Asian emission inventory for year 2006 (Zhang et al., 2009). Biomass burning emissions at $0.5° \times 0.5°$ horizontal resolution for the water year 2013-14 are obtained from the Global Fire Emissions Database, Version 3 (GFEDv3) (Van Der Werf et al., 2010), which are vertically distributed in our simulation using the injection heights prescribed by Dentener et al. (2006) for the Aerosol Inter Comparison project (AeroCom). Sea salt and dust emissions follow Zhao et al. (2014). Dust surface emission fluxes are calculated with the Georgia Institute of Technology-Goddard Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) dust emission scheme (Ginoux et al., 2001), and emitted into the eight MOSAIC size bins with respective mass fractions of 10^{-6} , 10^{-4} , 0.02, 0.2, 1.5, 6, 26, and 45%.

Aerosol optical properties are computed as a function of wavelength for each model grid cell. The Optical Properties of Aerosols and Clouds (OPAC) data set (Hess et al., 1998) is used for the shortwave (SW) and longwave (LW) refractive indices of aerosols and a complex refractive index of aerosols (assuming internal mixture) is calculated by volume averaging for

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

each chemical constituent of aerosols for each bin. A spectrally-invariant value of 1.53 ± 0.003i is used for the SW complex refractive index of dust. Fast et al. (2006) and Barnard et al. (2010) provide detailed descriptions of the computation of aerosol optical properties such as extinction coefficient, single scattering albedo (SSA), and asymmetry factor in WRF-Chem. Following Zhao et al. (2011) and Zhao et al. (2013a), aerosol radiative feedback is coupled with the Rapid Radiative Transfer Model (RRTMG) (Mlawer et al., 1997) and the direct radiative forcing of individual aerosol species in the atmosphere are diagnosed. Aerosol–cloud interactions are included in the model following Gustafson et al. (2007).

The increasingly used Snow, Ice, and Aerosol Radiation (SNICAR) model simulates the snow properties and associated radiative heating rates of multilayer snow packs (Flanner and Zender, 2005; Flanner et al., 2009, 2012 and 2007). Fundamentally, it employs the snow albedo theory (parameterization) based on Warren and Wiscombe (1980) and the two-stream radiative approximation for multilayers from Toon et al. (1989). SNICAR can also simulate aerosol radiative effect in snow for studying the LAP heating and snow aging (Flanner et al., 2007). Recently, laboratory and site measurements are used to validate the SNICAR simulated change of snow albedo for a given BC concentration in snow (Hadley and Kirchstetter, 2012; Brandt et al., 2011). For radiative transfer calculations, SNICAR defines layers matching with the five thermal layers in community land model (CLM) that vertically resolve the snow densification and meltwater transport (Oleson et al., 2010). In WRF-Chem-SNICAR coupled model, BC and dust deposition on snow is calculated in a prognostic approach through dry and wet deposition processes. BC in snow can be represented as externally and internally mixed with precipitation hydrometeors depending on the removal mechanism involved, but dust is considered to only mix externally with snow grains (following Flanner et al., 2012). SNICAR in WRF-Chem simulates four tracers of dust based on size (with diameters of 0.1–1, 1–2.5, 2.5–5, and 5–10 μm) and two tracers of BC (externally and internally mixed BC with 0.2 μm dry diameter) in snow. The MOSAIC aerosol model simulates dust in the atmosphere with eight size bins (0.039–0.078, 0.078–0.156, 0.156–0.312, 0.312–0.625, 0.625 - 1.25, 1.25–2.5, 2.5–5.0, and 5.0–10.0 μm in dry diameter). The first 4 bins are coupled with the smallest bin of dust particles in SNICAR. While the next two MOSAIC bins (5^{th} and 6^{th}) map into the second bin of SNICAR, the 7^{th} and 8^{th} MOSAIC dust bins correspond to the third and fourth SNICAR dust bins (Zhao et al., 2014), respectively. Deposition of LAPs to snow in SNICAR are immediately mixed in the CLM surface snow layer (< 3 cm). CLM adds excess water in the layer above to the layer beneath during melting. The scavenging of aerosols in snow by meltwater is assumed to be proportional to its mass mixing ratio of the meltwater multiplied by a scavenging factor. Scavenging factors for externally mixed BC and internally mixed BC are assumed to be 0.03 and 0.2, respectively, and 0.02, 0.02, 0.01, and 0.01 for the four dust bins (all externally mixed). Although these scavenging factors are comparable to observations (Doherty et al., 2013), the scavenging ratios can be highly heterogeneous and introduce high uncertainty into the estimation of LAP concentrations in snow (Flanner et al., 2012; Qian et al., 2014). More detailed description about the aerosol deposition and mixing processes, computation of optical properties of snow and LAPs in WRF-Chem-CLM-SNICAR coupling can be found in Zhao et al.(2014) and Flanner et al.(2012).

Configured in the way similar to the WRF-HR, a coarse (1° × 1°) gridded WRF-Chem simulation is also performed using a quasi-global model (hereafter referred as WRF-CR) with 360 × 130 grid cells (180° W–180° E, 60° S–70° N). Periodic boundary conditions are used in the zonal direction. Reanalysis of the TROpospheric (RETRO) anthropogenic emissions for the year 2010 (ftp://ftp.retro.enes.org/ pub/emissions/aggregated/anthro/0.5x0.5/) is used for anthropogenic aerosol and precursor gas emissions in the coarse gridded quasi-global WRF-Chem simulation except for Asia and the United States. INTEX-B anthropogenic emissions (Zhang et al., 2009) and US National Emission Inventory are used for Asia and the U.S., respectively. Emissions of biomass burning aerosols, sea salt, and dust are treated in the same way as described above for the WRF-HR simulation. More details about the quasi-global WRF-Chem simulation can be found in (Zhao et al., 2013b);Hu et al., 2016). Chemical initial and boundary conditions to the WRF-HR simulation are provided by this quasi-global WRF-CR runs for the same time period to include long-range transported chemical species.

2.2: Aerosol Optical Depth (AOD) dataset

The aerosol robotic network (AERONET – https://aeronet.gsfc.nasa.gov) is a global network of ground based remote sensing stations that provides quality-controlled measurements of AOD with uncertainties ~0.01 under clearsky conditions over India (Holben et al., 1998; Dubovik et al., 2000). CIMEL Sun scanning spectral radiometers are used to measure direct Sun radiance at eight spectral channels (340, 380, 440, 500, 675, 870, 940, and 1020 nm) and measure spectral columnar AOD (Holben et al., 1998). AERONET provides measurements at ~15 min temporal resolution from sunrise to sunset.

Skyradiometer Network (Skynet) is another global network of ground based spectral scanning radiometer (POM-01L, Prede, Japan) stations that provides quality-controlled measurements of AOD (Nakajima et al., 1996). With an automatic sun scanner and sensor, it measures sky irradiance in five wavelengths i.e. 400, 500, 675, 870, and 1020 nm. The measured monochromatic irradiance data is processed by using Skyrad.Pack version 4.2 software. Calibration of the Sky radiometer is carried out on a monthly basis (http://atmos3.cr.chiba-u.jp/skynet/data.html). Details of the instrumentation and software protocol can be found in Campanelli et al. (2007) and Ningombam et al., (2015). In this study, we have also used AOD measurements at 500 nm over MERAK, a high altitude Skynet station in Himalaya for water year 2013-14.

The MODerate resolution Imaging SpectroRadiometer (MODIS) instrument onboard the NASA AQUA satellite provides global coverage of daily radiance observations (at 1330 LT) in 36 spectral channels. Over North India, Tripathi et al., (2005) has shown that MODIS observations correlate well with ground based measurements. For the evaluation of model simulated AOD, 1° gridded Level 3 AOD estimates (collection 6) at 0.55 µm wavelength obtained from the MODIS instrument are used during water year 2013-14. However, the MODIS land aerosol algorithm uses a dark target approach (Levy et al., 2007), which, is known to have large uncertainties over arid and mountainous surfaces (Levy et al., 2010).

2.3: Spatially and temporally complete MODSCAG and MODDRFS retrievals

Subpixel snow-covered area and snow grain size are retrieved from MODIS-observed surface reflectance data using the physically based MODIS Snow-Covered Area and Grain size (MODSCAG) (Painter et al., 2009) algorithm. In each snow covered pixel, MODSCAG attributes a fractional snow-covered area and grain size using spectral mixture analysis to determine proportion of the pixel that is snow and is not snow. MODSCAG is more accurately identifies snow cover throughout the year than the widely used MODIS snow product: MOD10A1 (Rittger et al., 2013). The MODSCAG snow-mapping algorithm for fraction of snow covered area has an uncertainty of ~ 5 % (Rittger et al., 2013). The current study incorporates pixel level snow cover area and snow grain size from MODSCAG over the HMA region to evaluate snow pack simulation and LAP-induced albedo reduction. Further, MODIS Dust

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Radiative Forcing in Snow (MODDRFS) model (Painter et al., 2012) is used to determine the LAP-induced albedo reduction over HMA. MODDRFS uses spectral reflectance differences between the measured snow spectral albedo and the modeled clean snow spectral albedo. The pixel level clean snow spectrum corresponding to MODSCAG retrieved snow grain sizes is calculated using discrete ordinate radiative transfer solutions for visible wavelengths and solar zenith angles. Coupled, these products provide the determination of snow albedo for the fractional snow cover with LAP inclusion. Reflectance inputs to MODSCAG and MODDRFS are degraded by cloud cover, off-nadir views, and data errors, but can be filtered in time and space to improve data quality and consistency. Our method for spatially cleaning and filling (Dozier et al., 2008;Rittger et al., 2016) combines noise filtering, snow/cloud interpolation and smoothing to improve the daily estimates snow surface properties (SSP). Using remotely sensed forest height maps (Simard et al., 2011) and MODSCAG vegetation fraction, we adjust the satellite viewable snow cover to account for snow under tree canopy (Rittger et al., 2016). We weight the observations based on satellite viewing angle that varies from 0 to 65 degree with larger uncertainties in off-nadir views (Dozier et al., 2008). The result is a set of spatially and temporally complete (STC) SSPs. Use of these products in an energy balance model to estimate snow water equivalent based on reconstruction produced more accurate snow cover than the Snow Data Assimilation System (SNODAS) or an interpolation of observations from snow pillows (Bair et al., 2016). In this study we use STC versions of MODSCAG and MODDRFS when comparing our WRF model output. The incomplete remotely sensed would be difficult to use given the gaps in data and uncertainties related to viewing angle (Dozier et al., 2008). Hereafter, the use of MODSCAG and MODDRFS terms will invariably refer to these STC-MODSCAG and STC-MODDRFS products.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

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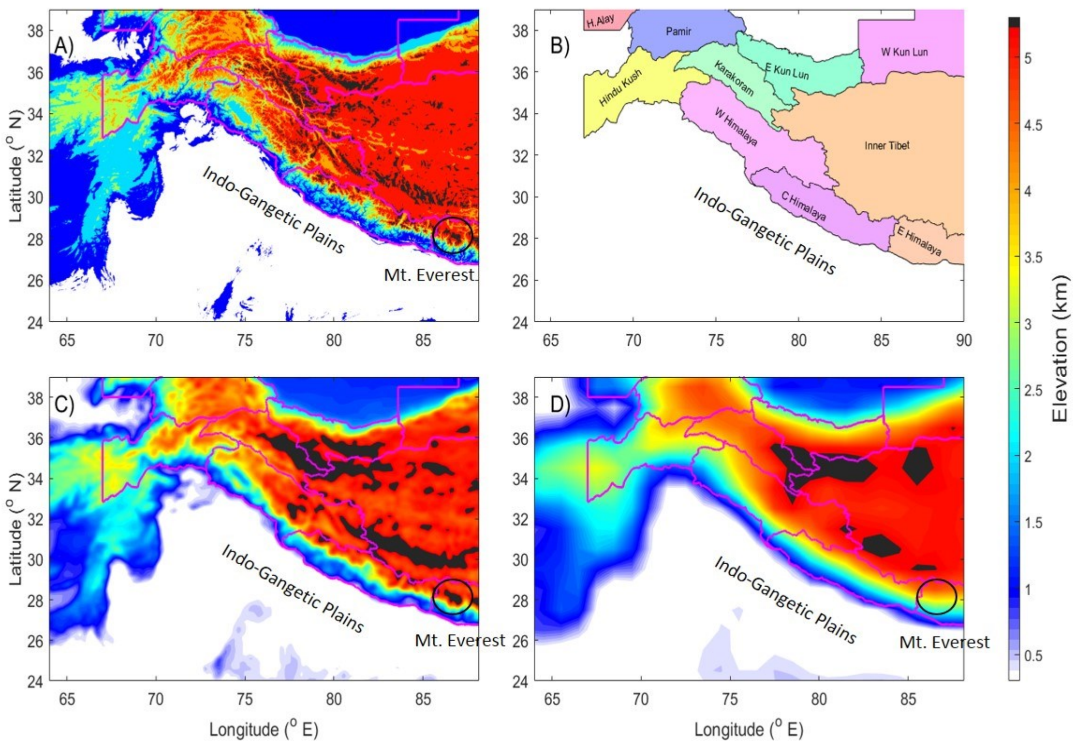

Figure 1: Panel A illustrates the terrain elevation at 1 km resolution from ETOPO1 dataset. Panel B shows glacier classification over HMA region used in this study following Randolph Glacier Inventory. Panels C and D illustrate terrain representation in WRF-HR run (12 km) and WRF-CR (1 degree) runs, respectively. For reference, Mt. Everest (shown as black circle in Figure 1C) is distinctly represented in Panel A and C, but not in Panel B.

2.4: Variation in terrain representation in WRF-HR and WRF-CR

Figure 1A illustrates the variations in terrain height over HMA at a resolution of 1 arc using ETOPO1 Global Relief Model, a publicly available global topographical dataset (Amante and Eakins, 2009). It clearly shows the enormous relief in terrain as we move from the Indo Gangetic plains (IGP) to the crest of the Himalaya and into the Tibetan Plateau (TP). The majority of HMA is above 4 km altitude with many Himalaya peaks at an altitude higher than 6 km. Figure 1B illustrates the mountain ranges and glaciers classified as per the Randolph Glacier

Inventory in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (Pfeffer et al., 2014). Specifically, Pamirs, Hindu Kush, Karakoram, Kunlun, and Himalaya hold the most number of glaciers in HMA. Figures 1C and 1D illustrate the representation of terrain elevation in WRF-HR and WRF-CR, respectively. Compared to Figure 1A, location of mountain peaks (altitude > 5.5 km) are better represented in WRF-HR compared to WRF-CR, as is particularly evident over the Karakoram, Kunlun, and Himalaya ranges. Moreover, the steep rise in elevation between IGP and TP is also well represented by WRF-HR, whereas it is more gradual in WRF-CR.

2.5: Methodology

Simulation of the snow macro- and micro-physical properties, aerosol loading and LAP in snow concentration from WRF-HR, WRF-CR and observational estimates (datasets described above) over HMA are compared in Section 3.1 and Section 3.2. In Section 3.3, the WRF-HR simulated LAP-induced snow albedo reduction values over HMA is compared with corresponding MODIS satellite based STC-MODSCAG and STC-MODDRFS. Lastly, a discussion on the high resolution model simulated LAP-induced radiative forcing estimates over HMA is also presented in context to previous studies and other atmospheric forcing.

The simulated fractional snow covered area (fSCA), duration of snow cover over a grid in terms of number of snow cover days (NSD), snow albedo (α) and snow grain sizes (SGS) and LAP-induced snow albedo darkening ($\Delta\alpha$) for midday (1000 -1400 LT) conditions from both the WRF models are compared with corresponding STC-MODSCAG and STC-MODDRFS retrievals over HMA. The number of snow cover days (NSD: defined as days having fSCA values ≥ 0.01) during water year 2013-14 is determined over each grid from STC-MODSCAG and both model runs. They are compared with corresponding values from STC-MODSCAG

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

retrievals, which, are observed during Terra overpasses at 10:30 LT. We have used a window from 10:00 LT to 16:00 LT for representing midday averages of modelled variables to incorporate the variability due to differences in timing (between model and real scenario) of weather conditions like precipitation and clouds. In addition, the change in snow albedo during 10:00 - 14:00 LT is < 0.01 (Bair et al., 2017), which, is low compared to other model physics- and data retrieval related uncertainties. The WRF-CR simulated variables and STC-MODSCAG and STC-MODDRFS retrievals are gridded to the resolution of WRF-HR (12 km) for ease of comparison. We have compared annual mean values as well as seasonal mean values for winter (December - February) and summer (April - June) season, separately. We have not considered the monsoon period in our analyses because the snow cover during the monsoon is negligible (except in glaciated regions at high altitudes) relative to other months (Figure S1). To evaluate spatial heterogeneity in our model, the seasonal and annual distribution of these variables are calculated separately for each sub region (shown in Figure 1) within HMA. In addition, to gain an understanding of the extent of temporal variability present in LAP-induced effects, we have also presented daily midday variation in LAP-induced snow darkening and LAP-induced radiative forcing at surface over Chotta Shingri glacier region (32.1-32.35 ^{o}N, 77.4-77.7 ^{o}E) located in the Chandra–Bhaga river basin of Lahaul valley, Pir Panjal range, in Western Himalayas. It is an accessible and representative site for glacier mass balance studies in western Himalayas. Chotta Shigri glacier has a cumulative glaciological mass loss of –6.72 m w.e. between 2002 and 2014 (Azam et al., 2016).

The simulated aerosol optical depth (AOD) is compared with available in-situ observations (described in Section 2.2). Here, quality assured (Level 2) midday (1000 to 1400 LT) averages of AOD (550 nm) at seven AERONET stations (Lahore, Jaipur, Kanpur, Gandhi

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

college, Kathmandu and CAS) and one SkyNet site within our study region are used to evaluate the simulated AOD values. Further, the simulated distribution of LAP concentration in snow at a few sites is compared with field measurements. Only a few field measurements of concentration of BC (LAP_{BC}) and dust particles (LAP_{dust}) in the snow surface or the surface layer are available over glaciated regions within our study domain. In this study, measurements of LAP_{BC} over Muztagh Ata in eastern slopes of Pamirs (Xu et al., 2006), Uttaranchal region of W. Himalayas (Svensson et al., 2018), East Rongbuk at 6.4 km altitude (Ming et al., 2012; Ming et al., 2008; Xu et al., 2009a) and composite of recent in-situ measurements from various studies near the NCO-pyramid site in Nepal at 5-6 km altitude (Kaspari et al., 2014; Yasunari et al., 2013; Jacobi et al., 2015; Ginot et al., 2014) is used. Similarly, the point measurements used for evaluating LAP_{dust} are over Abramov glacier in western slopes of Pamirs (Schmale et al., 2017), Muztagh Ata in eastern slopes of Pamirs (Wake et al., 1994) , East Rongbuk (Ming et al., 2012) and near NCO-pyramid station (Ginot et al., 2014), respectively.

3: Results and Discussions

3.1: Snow physical, microphysical and optical properties

The largest values of region-averaged annual mean fSCA within HMA are observed in both the satellite retrievals and the model runs over the Karakoram region (mean=0.45) followed by Pamirs, Himalayas and Hindu Kush in the HMA region (Figures 2A and 2B). In comparison, the fSCA over Kunlun and TP are lower (<0.3), but, pockets of very high fSCA (~0.7) are visible over the west Kunlun ranges (Figure 2A). The annual mean fSCA values and the fine spatial variability are well simulated by WRF-HR (Figure 2B) over the entire HMA region. Of exception are simulations over the Karakoram, where WRF-HR overestimates annual mean fSCA, however, the distribution of annual mean fSCA from WRF-HR and STC-MODSCAG

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Atmospheric Chemistry and Physics
Discussions

agree in all the sub regions (Figure 2D). This observation is largely valid also for summer months (Figure 2F). But, significant overestimation in distribution of fSCA (by >0.2) during winter is present over Pamirs, Karakoram, W. Himalayas, TP and Kunlun region (Figures 2E). STC-MODSCAG retrievals illustrate that the Pamirs (NSD=230 days) and Karakoram (NSD=270 days) ranges remain snow covered for 7-9 months of the year (Figure 3A). Similarly, the grids in Hindu Kush (NSD=194), W. Himalayas (NSD=189 days) and C. Himalayas (NSD=191 days) are snow covered for ~ 6-7 months. Mountains in E. Himalayas (NSD=142 days) remain snow covered for only 4-5 months of the year. The distribution of annual NSD values simulated by WRF-HR in each sub region is close to STC-MODSCAG values (Figure 3D). Also, the spatial distribution and magnitude of simulated NSD by WRF-HR is similar to that from STC-MODSCAG for different seasons, separately (Figure S2). Thus, overestimation of annual mean fSCA in WRF-HR during winter is not due to mere averaging error associated with underestimation in simulated NSD during winter (Figure S2).

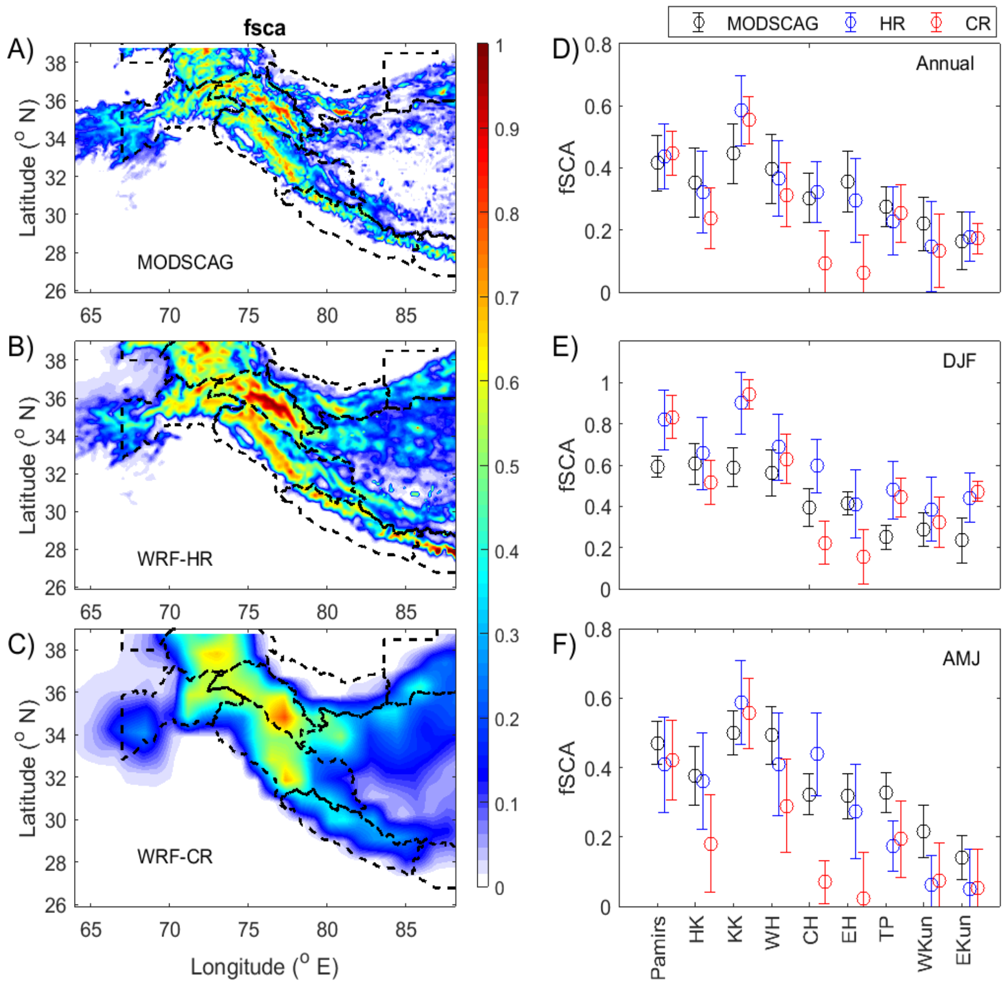

Figure 2: Spatial distribution of annual mean snow cover fraction (fSCA) during midday (1000-1400 LT) for water year 2013-14 from A) STC-MODSCAG retrievals, B) simulated values from WRF-HR and C) WRF-CR simulations. Panels D-F illustrate the distribution of midday mean fSCA over each subregion identified by glacier classification following Randolph Glacier Inventory (X-axis). The circle and vertical legs represent mean±standard deviation over each region for D) entire year, E) winter (December - February) and F) summer (April - June) season, separately. Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, WKun and EKun, respectively.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

We also calculated the number of days of snow cover with low fSCA (<0.5; Figure 3B) and high fSCA (>0.5; Figure 3C) values, separately. The grids in Kunlun, Northern slope of Karakoram, eastern slope of Pamirs and TP region are dominated by snow cover of relatively low fSCA for most of their snow cover duration (Figure 3B). But, grids in Hindu Kush, Himalayas and southern slopes of Karakoram are generally covered with high fSCA values throughout the year (Figure 3C). The distribution of simulated NSD values over each sub region for low and high fSCA scenario is shown in Figures 3E and 3F, respectively. WRF-HR can well simulate the NSD over grids with high fSCA (Figure 3F) but significantly underestimates NSD over grids with low snow cover (Figure 3E). Note that the regions dominated by low annual fSCA in this water year are actually the same regions where WRF-HR simulated fSCA values are being overestimated in winter (Figure 2E). Thus, simulation of fewer number of days with low fSCA (and vice versa) in WRF-HR might also be contributing partially to the overestimation of winter fSCA simulated in WRF-HR compared to STC-MODSCAG. Interestingly, WRF-CR simulated NSD values for low fSCA case is in better agreement with STC-MODSCAG values (Figure 3E). Winter mean distribution of WRF-CR simulated fSCA over Kunlun, W.Himalaya and TP region better match STC-MODSCAG values than the corresponding WRF-HR simulated fSCA values (Figure 2F). It is noteworthy that these subregions (which are dominated by low fSCA grids) receive snowfall from western disturbances during winter months. The cloud cover associated with the western disturbances over these sub regions are extensive in winter which also introduces uncertainty in MODSCAG retrievals and STC processing and contributes to the differences between WRF-HR and MODSCAG in fSCA.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Atmospheric Chemistry and Physics
Discussions

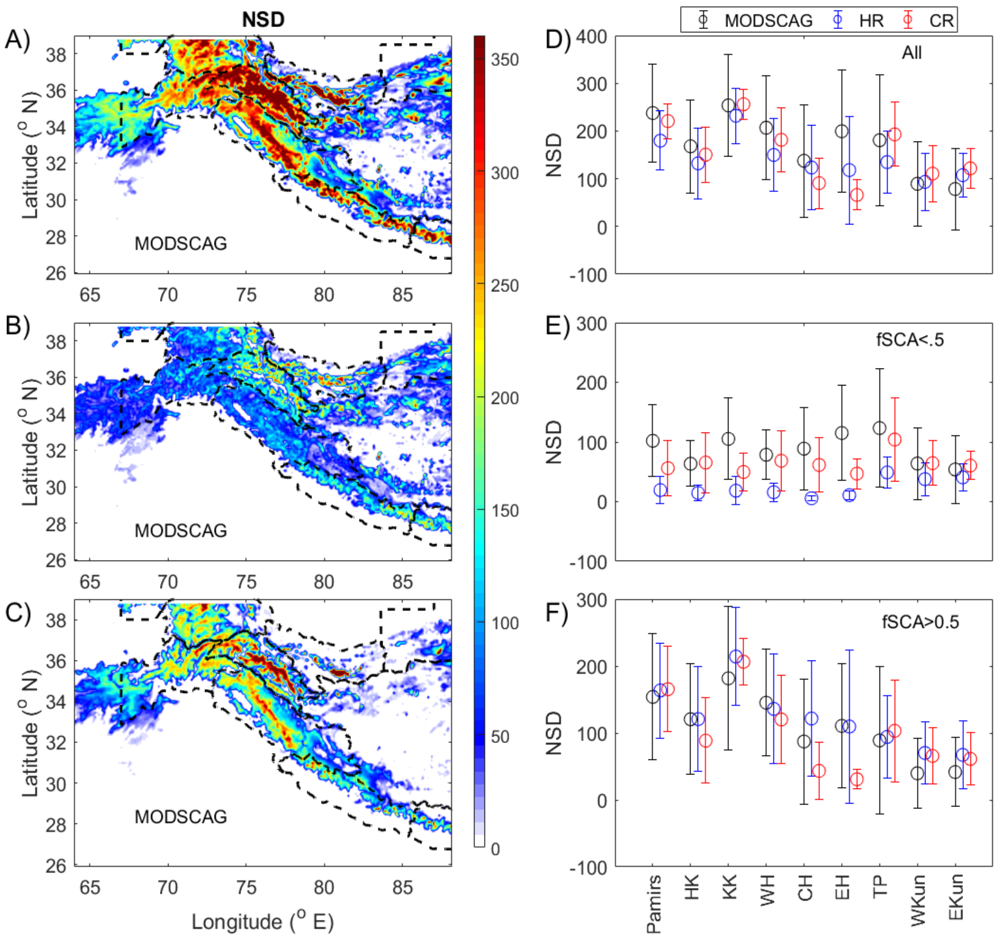

Figure 3: Spatial distribution of snow duration in terms of NSD from A) STC-MODSCAG retrievals. Panel B and C are similar to Panel A, but, shows number of days when fSCA is below and above 0.5 over each grid, respectively. Panels D illustrate the distribution of NSD over each sub region identified by glacier classification following Randolph Glacier Inventory. The circle and vertical legs represent mean ± standard deviation over each region for entire year. Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, WKun and EKun, respectively. Panel E and F are similar to Panel D, but, for NSD corresponding to fSCA values below and above 0.5, respectively.

Comparison between performance of WRF-CR and WRF-HR for fSCA clearly show significant improvements in the WRF-HR simulations over the Hindu Kush and Himalayan ranges (Figure 2D). For instance, the simulated annual mean fSCA in WRF-CR around Mt. Everest (shown as black circle in Figure 1) is less than 0.1 (Figure 2C). This is contrary to the high fSCA values observed at Mt. Everest (0.7 in Figures 2A) and simulated by WRF-HR (0.7 in Figures 2B). Moreover, the improvement is present in both winter and summer months indicating it's independence from meteorological variations (Figures 2E and 2F). Analysis of NSD values indicate that the snow cover duration in WRF-HR also improved significantly over these slopes (Figure S3) irrespective of the season. Note that WRF-CR underestimates the snow duration over Hindu Kush and Himalayas by ~2-6 months (Figure S3) and the spatial location of grids with very high annual mean fSCA values (mountain ranges) improved in WRF-HR compared to the STC-MODSCAG data (Figures 2A-C). The observed improvement in fSCA and NSD simulation over the slopes of Himalaya and Hindu Kush can be attributed to better terrain representation in WRF-HR.

Next, the simulated microphysical properties of the snow pack are evaluated against the remote sensing retrievals. Spatial patterns in annual mean SGS from STC-MODSCAG are similar to that seen in fSCA with highest values over the Karakoram and Himalayan ranges (Figures 4A) corresponding to the highest elevations and likely the coldest temperatures hindering snow grain growth. This spatial distribution of annual mean SGS values is well simulated in WRF-HR runs (Figure 4B). But, the annual mean values are largely overestimated by 30-50 micron (Figure 4D) relative to STC-MODSCAG. The seasonal distribution of region-segregated SGS values from WRF-HR also compares well with that from STC-MODSCAG retrievals (Figure 4E and 4F). Simulated annual mean SGS from WRF-CR (Figure 4C) lack the fine spatial variability seen in STC-MODSCAG and WRF-HR. Moreover, the SGS estimates are largely underestimated (by up to 100 microns) by WRF-CR specifically over grids in central and eastern Himalayas, TP and Kunlun ranges (Figure 4D). The large underestimation of SGS from WRF-CR and overestimation of SGS from WRF-HR is present for both summer and winter months (not shown). The overestimation of SGS from WRF-HR values corroborate well with the finding that the simulated fSCA distribution from WRF-HR is largely skewed towards higher values (Figure 3). Similarly, the unrealistically low mean values of SGS from WRF-CR over Himalayas, TP and Kunlun ranges are consistent with the underestimation of fSCA and NSD values over these regions (Figure 2 and 3). While, SGS retrievals from STC-MODSCAG are based on observed surface reflectance, the modeled SGS is calculated from simulated snow mass in top model layer in the grid. Hence, improvement in simulation of fSCA and NSD in high resolution WRF-HR runs also caused the SGS values from WRF-HR to be closer to STC-MODSCAG retrievals than SGS from WRF-CR runs. It is worth noting that the presence of cloud cover influences STC-MODSCAG retrievals of SGS towards smaller grain sizes if clouds are misidentified as snow. This systematic error could also contribute to the SGS differences between WRF-HR and STC-MODSCAG estimates.

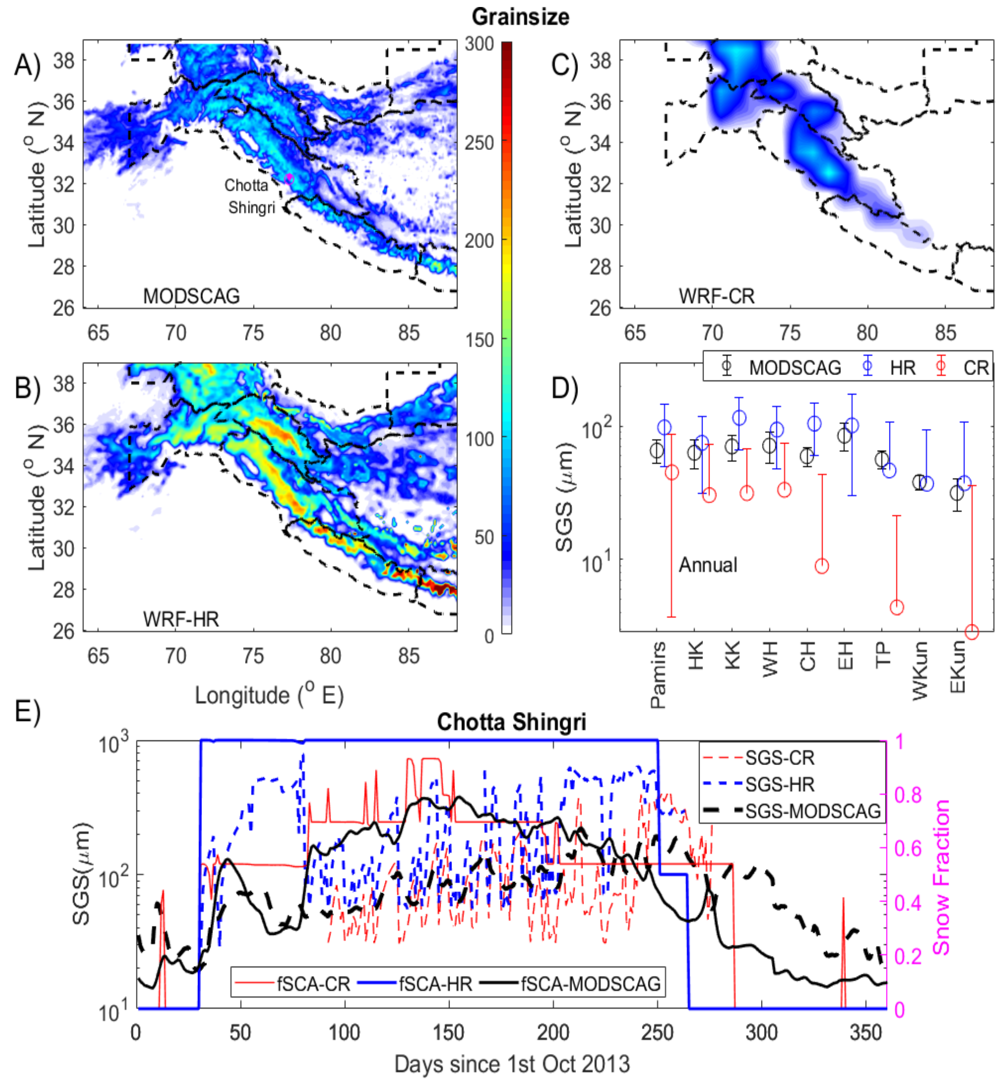

Figure 4: Spatial distribution of annual mean snow cover fraction (fSCA) during midday (1000-1400 LT) for water year 2013-14 from A) STC-MODSCAG retrievals and simulated values from B) WRF-HR and C) WRF-CR runs is shown. Panel D illustrates the distribution of midday mean fSCA over each subregion identified by glacier classification following Randolph Glacier Inventory. The circle and vertical legs represent mean±standard deviation over each region for the entire year. Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, WKun and EKun, respectively. Panel E shows time-series of daily midday SGS (hashed lines) and fSCA (solid lines) from MODSCAG (black), WRF-HR (blue) and WRF-CR (red) over a grid located near the Chotta Shingri glacier (marked by magenta diamond in Figure 3A) of western Himalaya region.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Interestingly, the SGS values from WRF-CR, over grids comprising the Chotta Shingri glacier (marked by magenta diamond in Figure 4A) of western Himalaya, are closer to STC-MODSCAG observations compared to those from the high resolution model, WRF-HR. As a sanity check of the above explanation of fSCA-SGS association, daily changes of SGS (hashed lines in Figure 4E) and fSCA (solid lines in Figure 4E) from STC-MODSCAG (black), WRF-HR (blue) and WRF-CR (red) over this glacier are compared. Fractional snow cover from STC-MODSCAG gradually increase (from below 0.2) in November, 13 (to above 0.8) in February and subsequently decrease back to 0.1 gradually by Septmeber, 14 at the glacier location. Corresponding SGS values from STC-MODSCAG closely followed the seasonal trend in fSCA varying around the values of 80-200 micron in winter. In comparison, simulated fSCA from WRF-HR values drastically increased to 1 in starting of November, 13 (from no snow cover before that), remain fully snow covered till mid-June, 14 and then steeply become snow free for rest of period after June. Compared to satellite estimates, fSCA from WRF-HR are greater in magnitude throughout the duration of snow cover indicating more snow mass simulated by WRF-HR. Associated SGS values simulated by WRF-HR values are also greater in magnitude (80-800 micron) than STC-MODSCAG estimates throughout the snow duration over the grid. In addition, the day to day variation in simulated SGS values is much larger than the STC-MODSCAG observations. However, fSCA variation from WRF-CR over this grid is very close to the variation seen by STC-MODSCAG and the associated SGS values from WRF-CR of (50-400 micron) are also closer to the estimated STC-MODSCAG values, supporting our argument of fSCA-linked bias in SGS estimates between model and satellite retrievals.

The annual mean snow albedo (α) values and the distribution over each sub region from satellite estimates (by combing grain sizes from STC-MODSCAG and decrease in albedo from

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

STC-MODDRFS (see Bair et al, 2016)) and simulated by both models are presented in Figure 4. Highest annual mean α values (~ 0.65-0.75) are observed over mountain peaks in Karakoram, Pamirs and W.Himalaya regions (Figure 5A). The location and magnitude of annual mean α over these grids are closely reproduced in WRF-HR with an underestimation of < 10% (Figure 5B). WRF-CR simulated annual mean α values over these grids have a considerably larger underestimate of ~50% (Figure 5C). Similar statistics are prevalent over all the sub regions of HMA (Figure 5D). Specifically, the distribution of α values from WRF-HR nearly matches the observed distribution, but, the distribution of albedos from the coarser model, WRF-CR, are generally 0.2-0.3 lower when compared to the observations. As above, cloud misidentified as snow could increase grain sized leading to slightly higher albedos. Also, note the opposite bias direction for albedo simulated by WRF-HR compared to simulated SGS values. This is intuitive as smaller snow particles cover greater surface area and therefore reflect more solar radiation from the surface. A similar pattern in distribution of snow albedo from WRF-HR and WRF-CR are also found over the sub regions for summer and winter months, separately (Figures 5E and 5F), indicating robust improvement in simulation of albedo values from WRF-HR throughout the year. The differences in simulated α in WRF-HR with the observations increased during summer and were lower during winter. Here, it is worth mentioning that we are comparing instantaneous estimates obtained from Terra overpass during 1000 LT with midday (1000-1400 LT) mean model values. The inherent diurnally in -α values under clear sky conditions in summer season (Bair et al., 2017) might contribute partially to the observed enhancement in differences during summer season. The improvement in α estimation from WRF-HR compared to WRF-CR can be attributed to the relatively better simulation of the overall macro- and microphysical properties of the snowpack in high resolution runs.

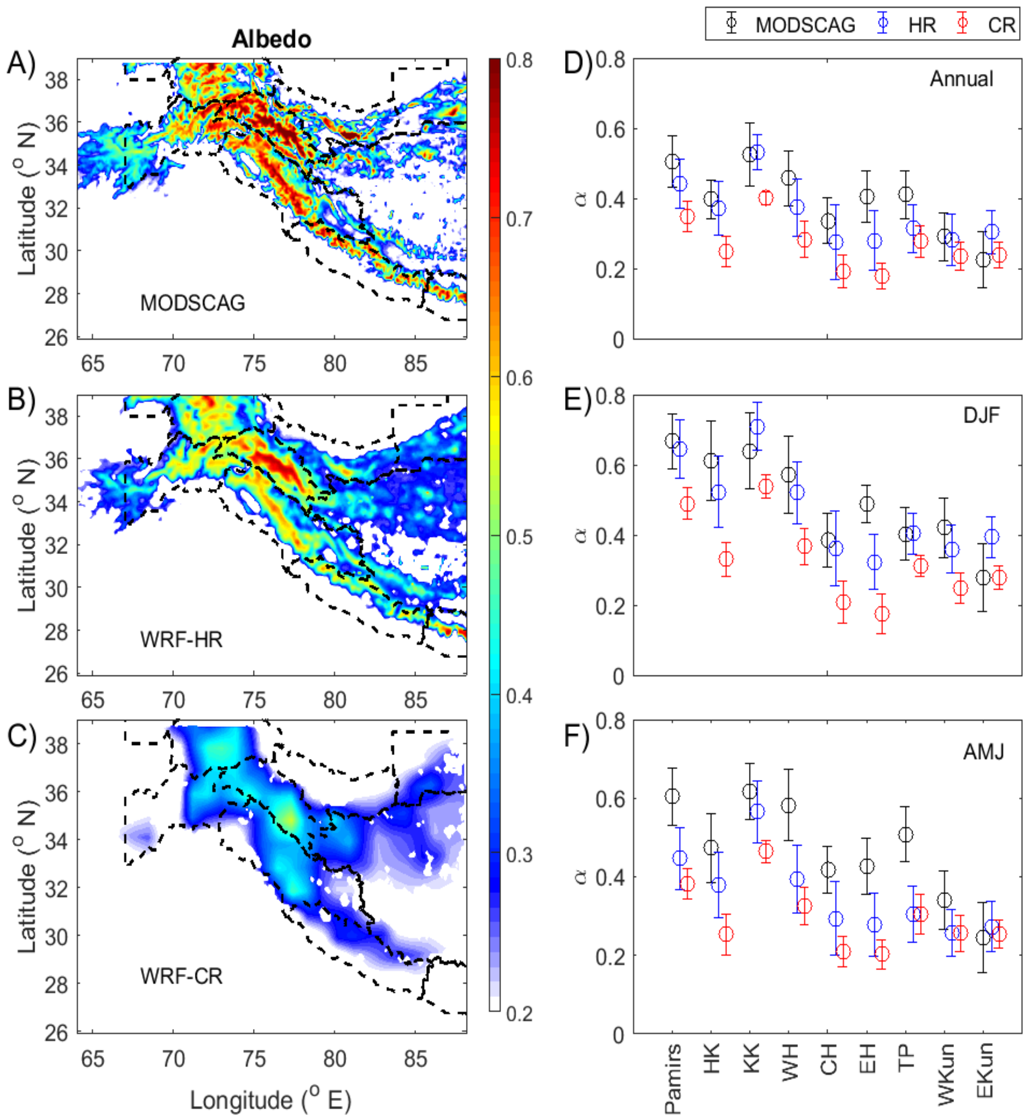

Figure 5: Spatial distribution of annual mean snow cover fraction (fSCA) during midday (1000-1400 LT) for water year 2013-14 from A) STC-MODSCAG retrievals and simulated values from B) WRF-HR and C) WRF-CR runs is shown. Panels D-F illustrate the distribution of midday mean fSCA over each subregion identified by glacier classification following Randolph Glacier Inventory. The circle and vertical legs represent mean±standard deviation over each region for D) entire year, E) winter (December - February) and F) summer (April - June) season, separately. Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, WKun and EKun, respectively.

3.2: Aerosol distribution and LAP in snow

We used available in-situ and ground sun photometer measurements from seven different sites across our study domain (location shown in Figure 6) to evaluate the simulated aerosol optical depth (AOD). The annual mean midday AOD at each site is shown in Figure 6A. Three sites (i.e. Merak, CAS and Kathmandu shown in Figures 6B-D) are located on the Himalaya slopes and the other four sites (Lahore, Jaipur, Kanpur and Gandhi College shown in Figures 6E-H) are located in the Indo-Gangetic Plains. In-situ measurements clearly illustrate a sharp decrease (4-5 fold) in mean AOD as we traverse higher up the Himalayan slope. The annual mean AOD for Lahore and Kanpur sites are 0.41 and 0.52, respectively, while the AOD over high elevated sites i.e. Merak and CAS sites are 0.07 and 0.05, respectively. Also, MODIS-observed AOD values prominently show the reduction in annual mean AOD from the Indo-Gangetic Plains (MODIS-AOD ~ 0.4-0.7) to the Tibet region (MODIS-AOD ~ 0.1-0.2) (Figure S5). Over the four sites in the Indo-Gangetic Plains, AOD simulated by both WRF-HR and WRF-CR runs are well correlated with observations (r= 0.5-0.6, Figures 6E-6H). The biases in modelled AOD are also similar (in the range of 0.2-0.4) in case of both WRF-HR and WRF-CR runs (Figures 6E-6H). Thus, no significant improvement in AOD values are achieved over the plain region with fine resolution. However, distinct and large improvement in simulated AOD is seen over the high elevation sites due to the increase in spatial resolution. Note that AOD values from WRF-CR are not strongly correlated with observations at these sites (Figures 6B-6D) and also have very high positive biases in AOD values (even higher than annual mean values at Merak and CAS stations). In contrast, the correlation between observations and WRF-HR is reasonably good (r=0.5-0.8 at these sites) using fine spatial resolution in WRF-HR. The positive biases in AOD from WRF-HR at Merak and CAS sites are lower than corresponding WRF-CR values by an order of magnitude. Presence of lower biases in AOD from WRF-HR over high elevation sites indicates that the observed sharp decrease in AOD values across the Himalayan slope are better captured by the higher resolution WRF run (WRF-HR) than in the coarser run. Greater annual mean AOD value is simulated by WRF-CR over the entire HMA region compared to WRF-HR (Figure S5) supporting an overestimation of AOD from WRF-CR at higher elevation in addition to the few sites. The presence of high biases (0.3-0.4) over Kathmandu valley even in WRF-HR runs indicate that model resolution even finer than 12 km is likely needed to better resolve the AOD distribution in complex terrain around valleys in Himalayan slope regions. Moreover, Jayarathne et al., 2018 shows that many local emissions are not accounted in global emissions which causes underestimation in simulated regional AOD values in these valleys. Temporal variability in monthly mean AOD (relatively higher AOD in summer) is simulated reasonably well by both the model versions (Figure S5).

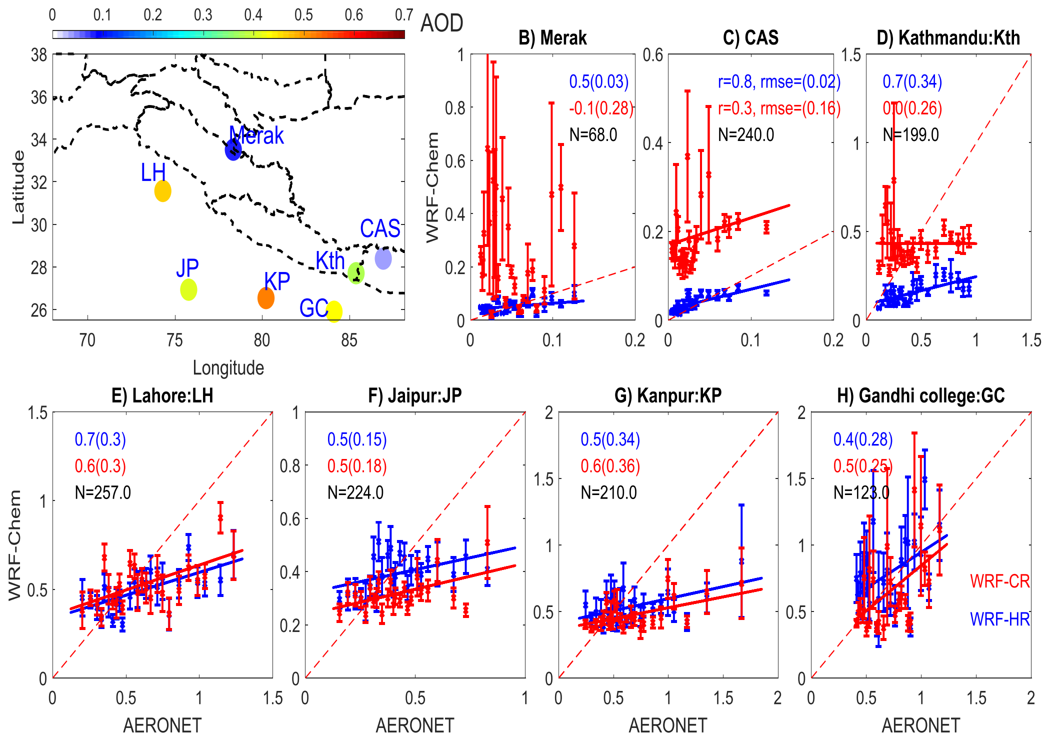

Figure 6: Comparison of midday (averaged over 1000-1400 LT) aerosol optical depth (AOD) measured by ground based sun photometer at seven sites within the study domain with corresponding simulated AOD values from, both, WRF-HR and WRF-CR. The annual mean AOD values over each site is shown by shade in topmost left Panel. The other panels illustrates the comparison over one of each stations shown by dots in topmost left Panel. The 'N' mentioned in the legend in each panel is the total number of days when collocated data between model and measurement is available over that site. These sample points are divided into 50 equal bins of ascending AERONET-AOD values (2 percentile each) and averaged. The standard deviation in each bin is shown by the vertical bars. The correlation coefficient values (r) are also mentioned in the legend followed in brackets by the relative error values (∑rmse/mean obs).

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

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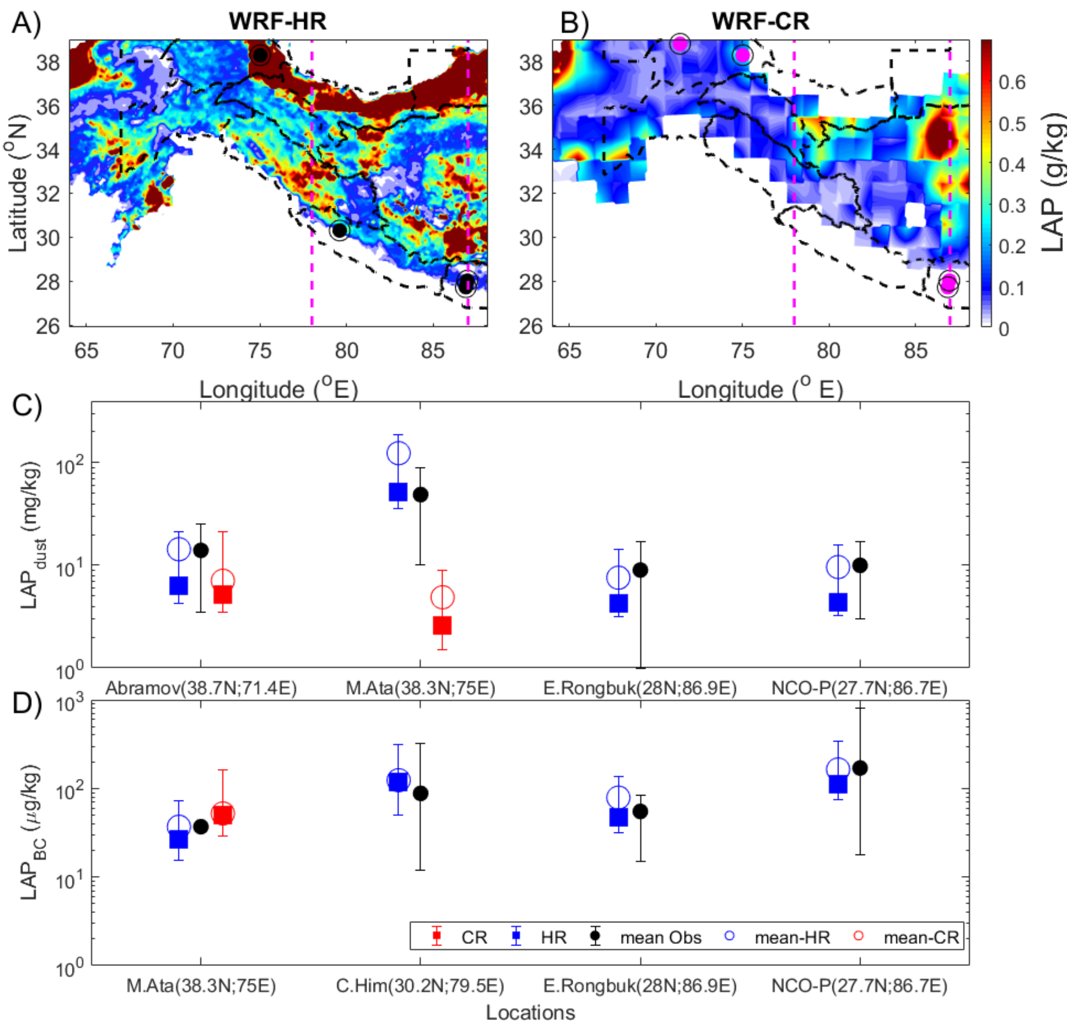

Figure 7: Spatial distribution of annual mean LAP concentration in snow top layer for water year 2013-14 simulated by A) WRF-HR and B) WRF-CR. The black dots in Panel A denote the locations where observations of BC in snow is available. Similarly, the magenta dots in Panel B denote the locations from where observation of dust in snow is available. Panels C-D illustrate comparison of simulated LAP_{BC} (top) and LAP_{dust} (bottom),respectively, in topmost snow layer with observed values over the marked locations in the Himalayan cryosphere. Annual mean (circle) and distribution (box plot) is used as metric of comparison. The WRF-HR and WRF-CR values are represented by blue and red, respectively. The pink lines in Panel A are the cross-sections shown in Figure 8.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

Significant differences in simulated AOD over high elevations of Himalaya slopes and TP indicate that considerable differences might also be present in LAP concentrations in snow between the two WRF simulations. Annual mean LAP concentrations in top snow layer from WRF-HR and WRF-CR simulations are compared in Figures 7A-7B. The comparison shows that LAP concentration in WRF-HR are significantly higher than WRF-CR simulated values. Quantitatively, the WRF-HR simulated annual mean LAP concentrations over the Pamir (0.5 g/kg), Karakoram (0.45 g/kg), Hindu Kush (0.2 g/kg), W. Himalaya (0.3 g/kg), C. Himalaya (0.2 g/kg) and E. Himalaya (0.08 g/kg) ranges is 3-5 times higher than the same from WRF-CR runs. In contrast, WRF-HR simulated LAP over TP (0.21 g/kg) and Kunlun ranges (0.8 g/kg) is similar to the mean magnitude simulated by WRF-CR runs. As a sanity check, we evaluate the simulated LAP concentrations against those previously reported in the literature. Figures 7C and 7D illustrate the evaluation of mean annual LAP concentration from WRF-HR and WRF-CR associated with BC (LAP_{BC}) and dust (LAP_{dust}), respectively, against the reported data (shown as filled black circles). The locations of reported LAP_{BC} (black filled dots) and LAP_{dust} (magenta filled dots) are shown in Figure 7C and Figure 7D, respectively. First, the difference in the magnitude of LAP_{dust} and LAP_{BC} over HMA is striking. The LAP_{dust} is more than 1000 times greater than LAP_{BC} both in observations and the models. Secondly, LAP_{BC} and LAP_{dust} values from WRF-HR are much closer to reported values compared to WRF-CR values. The differences in mean of reported LAP_{BC} and LAP_{dust} distribution to that simulated by WRF-HR at various sites are in range of 5-30 μg/kg and 5-20 mg/kg, respectively. WRF-CR well simulates the concentrations of LAP_{BC} and LAP_{dust} over Pamirs (~ 10 mg/kg), but significantly underestimates the LAP_{BC} and LAP_{dust} (by an order of magnitude) over the Himalayan ranges. Although the reported data are not specific to water year 2013-14, it can be reasonably assumed that the inter-

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

annual variations of LAP concentration in snow is of the order of magnitude as uncertainty in the observations. Thus, the WRF-HR better simulates aerosol and LAP concentration than the WRF-CR over the HMA region.

It is interesting to note that finer spatial resolution resulted in lower AOD but greater LAP values in snow over some places in HMA. For more insight, the vertical distribution of aerosol concentration in altitude-latitude space (Figure 8) across two latitudinal cross-sections (magenta colored lines in Figure 7A) is analyzed. Figure 8 illustrates the differences in simulated vertical distribution of mean aerosol number concentration along 78°E (row 1) and 87°E (row 2) for WRF-HR (left column) and WRF-CR (right column) runs, respectively. Corresponding terrain elevation (black solid line) and snow depth (magenta bars) are also overlaid in these plots. The latitude-altitude plots clearly illustrate that improved representation of the terrain in WRF-HR shows the sharp change of elevation over Himalayan foothills and causes a steeper natural barrier to the transport of aerosols uphill from IGP to HMA region than in the WFR-CR model. Also, high spatial resolution enhances snowfall in WRF-HR over the HMA region relative to the WRF-CR model. While the former change increased annual dry deposition flux, more snowfall caused greater wet deposition annually in WRF-HR compared to WRF-CR (Figure S6). The combination of these effects increases the deposition of aerosols and therefore LAP on the southern slopes of Himalaya in the WRF-HR run. This explains the coexistence of higher LAP concentration/deposition and lower AOD across HMA in WRF-HR, compared to corresponding WRF-CR results.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

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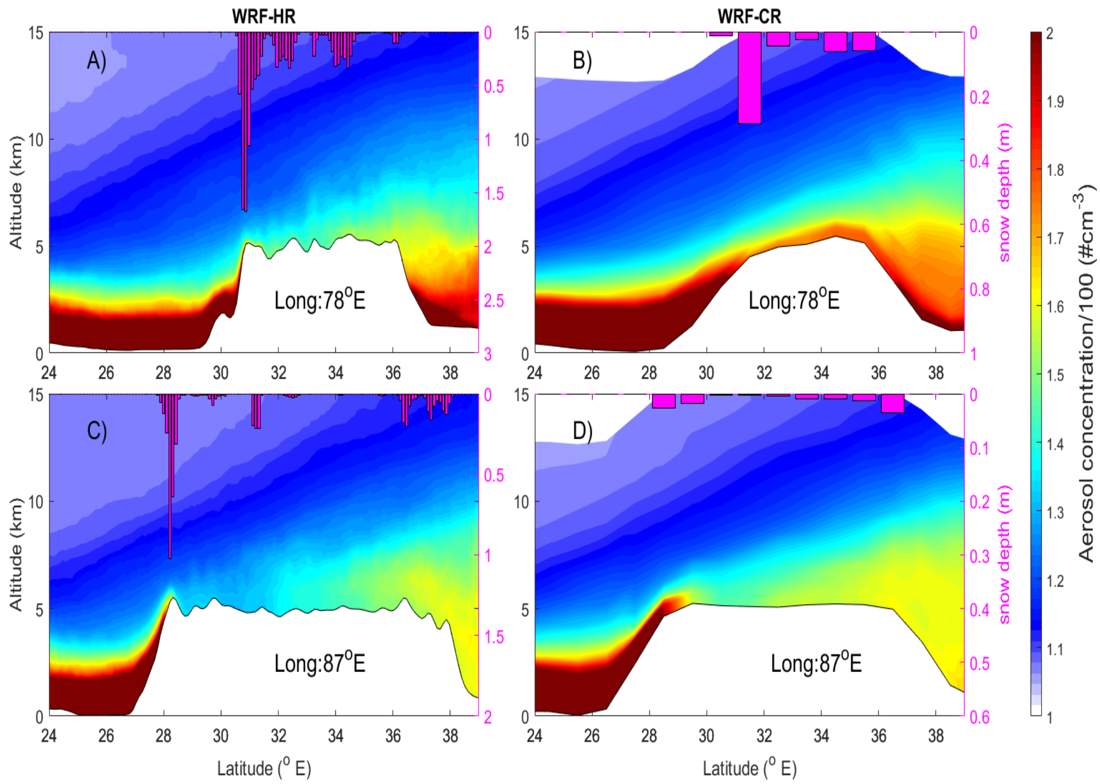
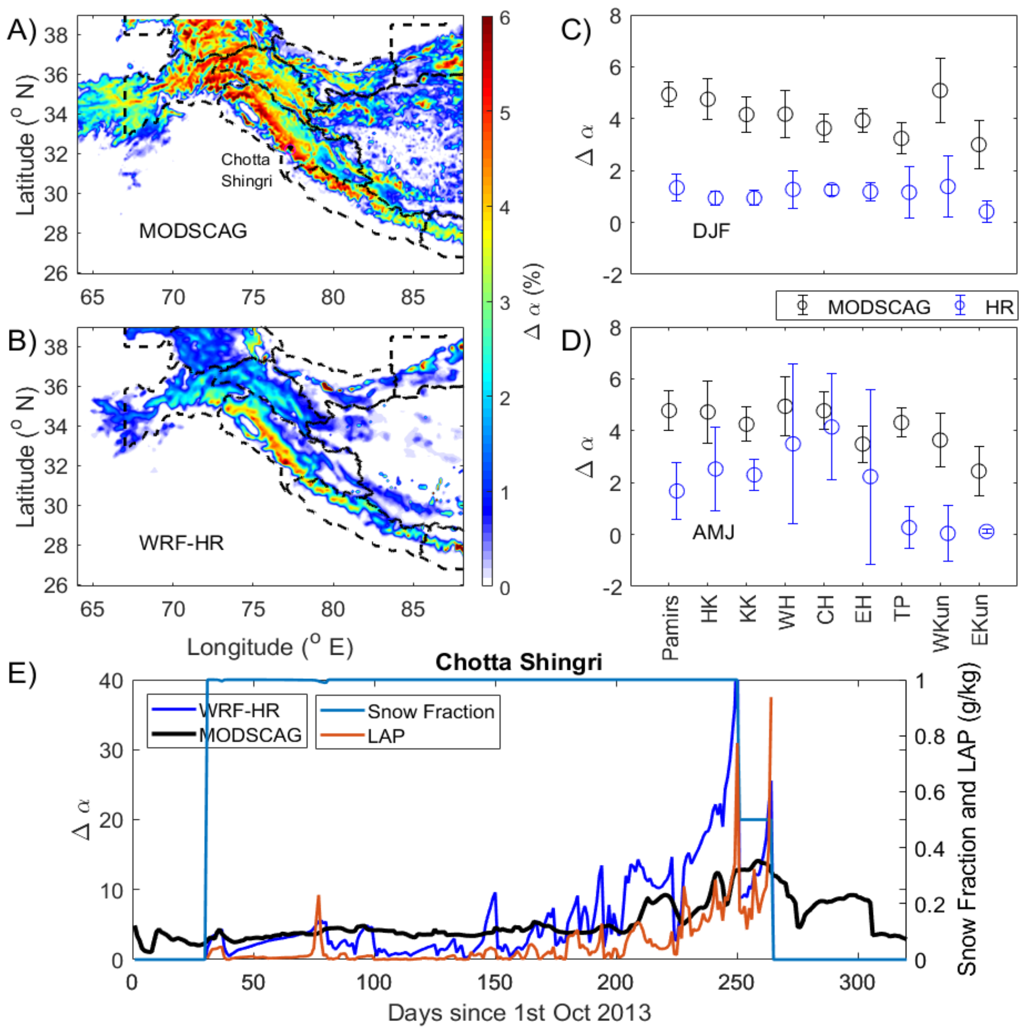


Figure 8: Longitudinally-averaged annual mean aerosol number concentration plotted in altitude-latitude space for two longitudinal traverses across the study domain, i.e. 78°N (Panels A and B) and 87°N (Panels C and D) for both WRF-HR (left column) and WRF-CR (right column). Corresponding terrain elevation is shown in solid black line. Corresponding to each latitude, the longitudinally-averaged annual mean snow depth is also presented in magenta color bars (using y-axis on the right).

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Atmospheric Chemistry and Physics
Discussions

673 **3.3 LAP-induced Snow darkening and radiative forcing**



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675 Figure 9: Spatial distribution of annual mean LAP-induced snow albedo darkening (Δα) during
676 midday (1000-1400 LT) for water year 2013-14 from A) MODSCAG retrievals and B) simulated
677 values from WRF-HR is shown. Panels C and D illustrate the distribution of midday mean Δα
678 over each sub region identified by glacier classification following Randolph Glacier Inventory.
679 The circle and vertical legs represent mean±standard deviation over each region for entire year.
680 Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West
681 Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, WKun and
682 EKun, respectively. Panel E shows time-series of daily midday Δα from STC-MODSCAG
683 (black) and WRF-HR (blue) over the grids located near the Chotta Shingri glacier (marked in
684 Figure A) of western Himalaya region. Also fractional snow cover and LAP concentrations from
685 WRF-HR over the same grids are included.

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
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Discussion started: 13 November 2018

STC-MODDRFS retrievals illustrate that locations in Hindu Kush and W. Himalayas have the highest annual mean LAP-induced reduction in snow albedo (Δα in %) followed by Karakoram, C.Himalayas and Pamir regions (Figure 9A). WRF-HR simulated the spatial variations in annual mean Δα reasonably well (Figure 9B), but, the magnitudes are underestimated by ~20-40 % throughout the domain. Note that the biases in annual mean values are lowest over grids in Himalayan ranges (where the underestimation is within 20%). Season wise and region wise distribution plots show that the WRF-HR biases are higher in winter months than the summer months (Figures 9C and 9D). While, WRF-HR simulated Δα values in the winter span between 1-3 %, the corresponding STC-MODDRFS estimates of Δα are larger with values ranging between 3-6% (thus no overlap with model values) over all the sub regions. In summer months, the distribution of modeled WRF-HR Δα values over Karakoram and Himalayan ranges are similar in magnitude, explaining the lower biases in annual mean values over Himalayas. This spatiotemporal variability in differences between STC-MODDRFS retrievals and simulated Δα values is consistent with the variability in biases of fractional snow cover seen in WRF-HR (Figures2E and 2F). Specifically, the large underestimation [and significant improvement relative to WRF-CR] in WRF-HR-Δα values (Figures 9C and 9D) in winter [summer] over Karakoram, Hindu Kush and Himalayas is in agreement with corresponding overestimation [improvement] of fSCA from WRF-HR over these regions (Figure 2).

For a closer look, the difference in daily midday mean Δα values from STC-MODDRFS (black) and WRF-HR (blue) are compared (in Figure 9E) over the grids of Chotta Shingri glacier (similar as Figure 4E). Corresponding, midday mean fSCA (light blue) and LAP (orange) from

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

WRF-HR are also plotted. Δα values from STC-MODDRFS are about 5% during winter months, but, increases in summer months until mid-June (peak value is 18%). Albedo reduction is closely associated with the temporal progression in midday LAP concentration in snow over this region at daily scale. In agreement, midday Δα values from WRF-HR are lower in winter months and higher in summer months. Except occasional peaks, with magnitudes of 3-4 %, Δα values from WRF-HR largely remained below 3 % till late February. A steep increase in Δα values from WRF-HR is seen in March (monthly mean ~ 4%), April (9%), May (13%) and June (18%). As already discussed, the simulated fSCA values in WRF-HR are greater than observed fSCA from STC-MODSCAG for most of the winter season (Figure4E). STC-MODDRFS estimated Δα is based on surface reflectance, while Δα calculated by model is for a surface layer of ~3 cm. The surface snow layer in SNICAR/CLM continuously evolves as fresh snowfall is added or with snow melting, so the LAP concentrations in surface layer depend on new snowfall, meltwater flushing, and layer combination/division (Flanner et al., 2007; Flanner et al., 2012; Oleson et al., 2010). Thus, more snow cover or thicker surface layers in winter results in lower values of annual mean LAP concentration and thus underestimates associated with LAP-induced snow darkening. In addition, the associated overestimation in modelled SGS during winter (Figure 4E) can also contribute to the lower WRF-HR-Δα values, because, bigger snow grains in WRF-HR lead to lower clean albedo and thus smaller reduction in albedo compared to STC-MODDRFS .

Another notable point is the large Δα values (> 20 %) from WRF-HR that occur towards the end of snow cover in June, which, is not seen in the STC-MODDRFS retrievals. The variations could be due to 2 reasons, either the snowpack is underestimated or the LAP concentration is overestimated by the model. Some factors which can contribute to these discrepancies in summer are 1) Larger LAP values may be simulated due to model uncertainties

Atmospheric Chemistry and Physics
Open Access
Discussions

in enhanced wet scavenging fluxes; 2) It is well known that with transport and deposition of aerosols from IGP to W.Himalayas increases afternoon with evolution of boundary layer over the IGP region (Dumka et al., 2015;Raatikainen et al., 2014). This feature is well simulated by the model (not shown). As STC-MODDRFS estimates are representative of 1000 LT, but, modelled values are representative of midday mean (1000-1400 LT), more aerosol deposition might be resulting in higher Δα values WRF-HR during summer months; 3) At the same time, the uncertainties associated with modelling aerosol-snow albedo microphysical feedbacks to snow melt may also be contributing to underestimation of snow packs in summer. However, more certainty on these modelled values require evaluation against in-situ measurements. It is worth mentioning here that no in-situ measurements are available for direct comparison of these high Δα values WRF-HR over W.Himalayas (Gertler et al., 2016). Nonetheless, the high values simulated during summer end are in near-range of previously reported values over other HMA regions. Kaspari et al., (2014) used the offline SNICAR model to report that BC concentrations in spring snow/ice samples at Mera Glacier were large enough to reduce albedo by 6-10% but that with the inclusion of dust, the reduction in albedo was 40-42% relative to clean snow. Recently, Zhang et al., 2018 has combined a large dataset of LAP measurements in surface snow with offline SNICAR model to illustrate that Δα can be >35% over Tibetan Plateau. Moreover, the composite effect of this discrepancy on seasonal/annual mean values is minimal as the snowpack is at its minimum at summer end. Similar high daily variability and huge LAPRF values (~200 W/m^2) in late summer as well as associated sudden decline in snow depth is also reported in sites over upper Colorado river basin (Skiles et al., 2015; Skiles and Painter, 2017).

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

Atmospheric Chemistry and Physics
Discussions

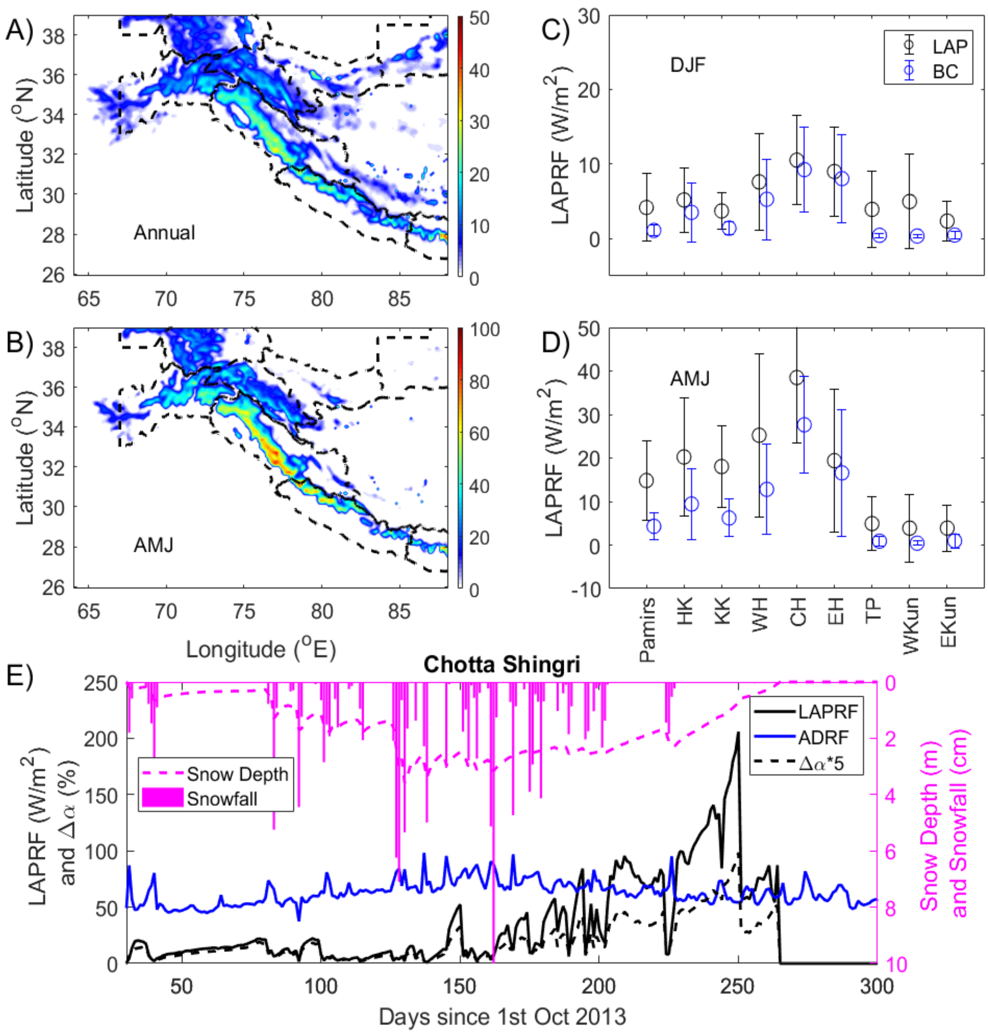

Figure 10: A) Spatial distribution of annual mean snow mediated LAP-induced radiative forcing (LAPRF) from WRF-HR and B) Spatial distribution of seasonal mean LAPRF over snow covered for summer season (AMJ) of water year 2013-14. Panels C and D illustrate the distribution of LAPRF (blue) and BC-only-LAPRF (black) over each sub region identified by glacier classification following Randolph Glacier Inventory for winter and summer, respectively. The circle and vertical legs represent mean±standard deviation over each region for entire year. Here, Hindu Kush, Karakoram, W.Himalayas, C.Himalayas, E.Himalayas, Tibetan Plateau, West Kunlun and East Kunlun regions are abbreviated as HK, KK, WH, CH, EH, TP, W.Kun and E.Kun, respectively. Panel E shows time-series of daily midday LAPRF (blue) and aerosol direct radiative forcing at surface (ADRF, blue) over the Chotta Shingri region (marked in Figure 9A) of western Himalaya region. The simulated LAP-induced change in snow albedo is also shown

in hashed black line. Also snow depth and snowfall from WRF-HR over the same grid is included.

In agreement with the spatial variation in Δα values, the annual mean snow-mediated LAP-induced radiative forcing (LAPRF) from WRF-HR over grids in W.Himalayas and C. Himalayas are highest followed by Karakoram, Hindu Kush, E.Himalayas and Pamir regions (Figure 10A). The spatial distribution of summer mean LAPRF values from WRF-HR (Figure 10B) is similar to that of the annual mean LAPRF values, but, the summer magnitudes are higher by an order of magnitude throughout most of the domain. This is mainly due to the large increase in Δα values in summer when LAPs aggregate on the surface compared with winter months when LAPs are continuously covered by new snow (Figure 9). Spatio-temporal variability in LAPRF is evident in the seasonal and regional distribution plots (Figure 10C and 10D). LAPRF values over the edges of HMA are greater than the highland TP region in both winter and summer months probably due to greater LAP deposition simulated over Hindu Kush, Karakoram and Himalayas regions (Figure S7) from the close proximity of dust sources. Also, it is clearly visible that the maximum LAPRF values within HMA region are present over grids in Himalayan ranges (during both winter and summer) with annual mean values > 50 W/m^2 (seen in Figures10B and 10D) and maximum instantaneous values higher than 150 W/m^2 (not shown). The time series of midday mean LAPRF values (Figure 10E) over the same grids in the Chotta Shingri Glacier region in western Himalaya is plotted to ascertain possible daily variability in LAPRF for midday over the region. Corresponding LAP-induced albedo reduction, snow depth and snowfall values are also plotted to show how LAPRF can affect local snow melting. The daily snow depth increases to 3.6 m in winter (30-150 days) followed by gradual reduction and snow cover melting in summer (150-270 days). The mean midday LAPRF value is ~10 W/m^2 in winter season, but, the magnitude increased gradually during March (18 W/m^2), April (44

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

W/m^2), May (86 W/m^2) and June (123.5 W/m^2), eventually terminating with a peak value of 202 W/m^2 in mid-June. The temporal evolution in $\Delta\alpha$ values closely followed the LAPRF values with a variation in range of 1-20%. The shortwave aerosol direct radiative forcing (ADRF) during midday at the surface over the same grid is also shown (as blue curve) in Figure10E. The momentarily high values of ADRF during the period indicate dust storms or sudden increase in aerosol loading over the grid. A closer look illustrate that the large daily variations in LAPRF and $\Delta\alpha$ are associated with the variability in ADRF and daily snowfall occurence over the grid. Fresh snowfall feedbacks result in subsequent reduction in LAPRF and enhancement in snow depth (Figure10E), while, higher aerosol loading over aged snow is followed by a clear increase in LAPRF. During melting, our model considers that only a fraction of LAP is washed away with meltwater, which, results in enhanced concentration of LAP during the end stages of the snowpack. This snow albedo feedback along with the momentary high aerosol loading (on 250th day) can explain the very high values of albedo reduction and LAPRF that were simulated during the last days of the snow cover over this grid. Higher LAPRF values indicate more energy being absorbed by the snow pack and thus more snow melting. Therefore, LAP-induced snow melting effect over Himalaya is very significant and is the largest relative to other areas within HMA.

Snow-mediated radiative forcing only due to BC- deposition is shown in Figures10C and 10D. Although, similar spatiotemporal pattern in LAPRF and BC-only-LAPRF is simulated, contrasting features in context to BC contribution to LAPRF are present in between the western part and eastern parts of HMA. During winter, the magnitude of BC-only LAPRF values is similar to that of total LAPRF over the western part of HMA region (i.e. Pamirs, Karakoram, Hindu Kush and W.Himalayas) suggesting that BC is a dominant contributor to net LAPRF (Figures10C). But, large differences between LAPRF and BC-only LAPRF is present during summer season over the western regions indicating that dust contribution is more or less equal to that of BC contribution over these regions (Figure10D). In contrast, the dust contribution to LAPRF values over the eastern domain (TP and Kunlun regions) is significant during winter season (Figures10C and 10D). The spatiotemporal variability in dust and BC contribution is reported previously and is mainly linked with the seasonal variability in meteorology and associated advection of South Asian emissions (Zhang et al., 2015; Wang et al., 2015a ; Niu et al., 2018). The ADRF-induced surface cooling effect may nullify the effects of LAPRF-induced warming effect on snowpack melting. But, the drastic increase in LAPRF values in April through June causes the magnitude of LAPRF to be twice that of ADRF over W. Himalayas during snow melting period, highlighting dominance of LAPRF (as also seen in Figure10E).

The simulated annual mean, summer mean and BC-only-LAPRF values from WRF-HR are in general, higher compared to previously reported estimates of LAPRF from model studies at coarser resolution. For example, Ménégoz et al., (2014) have reported annual mean LAPRF of ~ 1-3 W/m^2 over Himalaya using an online simulation at 50 km resolution grid. Similarly, Qian et al., (2011) used the Community Atmosphere Model version 3.1 at coarser spatial resolution to show that simulated aerosol-induced snow albedo perturbations can generate LAPRF values of 5-25 W/m^2 during spring over HMA. Also, coarsely resolved GEOS-Chem runs simulated BC-only-LAPRF can vary from 5 to 15 W/m^2 in the snow-covered regions over the TP (Kopacz et al., 2011). Recently, a decade long simulation using the RegCM model at 50 km spatial resolution also estimated maximum BC-only-LAPRF values of 5-6 W/m^2 over the Himalaya and southeastern TP averaged over non-monsoon season (Ji, 2016). However, the comparison of WRF-HR and WRF-CR simulations provided in this study clearly show that the magnitudes of snow macro- and micro-properties, aerosol loading and LAP-induced albedo darkening over

Himalayas improved significantly with finer spatial resolution. Thus, the global model simulated LAPRF values are likely underestimated. In agreement, recently, Zhang et al., (2018) and has estimated BC-only-LAPRF of 20-35 W/m^2 using offline SNICAR calculation forced with a greater coverage of measurements of surface snow content. He et al., (2018) have also reported similar high BC-only-LAPRF values after implementing a realistic snow grainsize parameterizations in offline SNICAR calculations over HMA.

4. Summary and Implications

In this study, we use the SNICAR model coupled with WRF-Chem regional model at high spatial resolution (WRF-HR; 12 km) to simulate the transport, deposition, and radiative forcing of light absorbing particles (LAPs; mainly Black Carbon and dust) over the high mountains of Asia (HMA) during water year 2013-14. The snow grain sizes and LAP-induced snow darkening was evaluated, for the first time, against comprehensive satellite retrievals (the MODSCAG and MODDRFS spatial and temporally complete retrieved satellite observations) over HMA. The atmospheric aerosol loading is evaluated against satellite and ground-based AOD measurements over HMA region. Results from another simulation which employ the same model configuration but a coarser spatial resolution (WRF-CR; 1 degree) are also compared with WRF-HR to illustrate the significance of a better representation of terrain on snow-pack and aerosol simulation over HMA. The main conclusions from our study are:

a) The simulated macro- and micro-physical properties and the duration of snow packs over HMA improve significantly due to the use of fine spatial resolution, especially over the southern slopes of Hindu Kush and Himalayan ranges.

b) Simulated aerosol loading over HMA is also more realistic in WRF-HR than in WRF-CR, which leads to a reduction in biases of annual mean LAP concentration in snow. This improvement is attributed to a more realistic simulation of wet deposition (due to a better simulation of snow pack) and dry deposition of LAPs (associated with a better representation of terrain) in WRF-HR.

c) WRF-HR captures the magnitude of LAP-induced snow albedo reduction (Δα) over Himalayas and Hindu Kush region relatively well compared to the STC-MODDRFS retrievals during summer. However, during winter, large biases in modelled Δα values are present. This is probably due to inherent uncertainties in model parameterizations and satellite retrievals associated with the cloud cover over HMA in winter period.

d) The glaciers and snow cover regions located in the Himalaya have the highest LAPRF within HMA i.e. annual mean LAPRF ~ 20 W/m^2 and summer mean LAPRF ~ 40 W/m^2. This is consistent with similar high values of Δα over Himalayan ranges i.e. annual mean Δα values ~ 2-4 % and summer mean Δα values ~ 4-8 %. The annual mean LAP concentration in snow values (200-300 mg/kg) are also high. Thus, the Himalaya (more specifically, western Himalayas) is most vulnerable to LAP-induced snow melting within HMA.

Ramanathan and G. Carmichael, (2008) suggest that atmospheric warming from LAPs (~20 W/m^2) may be just as important as greenhouse gases in the melting of snowpack and glaciers over Asia. In this context, the high magnitudes of LAPRF values in summer over HMA (~ 40 W/m^2) clearly shows that snow-mediated aerosol forcing on snow melting over HMA can be twofold the significance of the atmospheric forcing. Nonetheless, the differences in snow surface properties between WRF-HR and satellite observations indicate probable uncertainties in model parameterizations. At the same time, the STC-MODDRFS retrievals themselves may have an uncertainty of ~ 5% in instantaneous Δα measurements. Thus, efforts to further reducing the

Atmos. Chem. Phys. Discuss., https://doi.org/10.5194/acp-2018-979
Manuscript under review for journal Atmos. Chem. Phys.
Discussion started: 13 November 2018

LAPRF uncertainties in the model are warranted in the future by using in-situ observations (i.e. field campaigns), specifically over the most affected western Himalayas, where relevant measurements are largely absent (Gertler et al., 2016). Moreover, satellite retrievals will be markedly improved in the coming decade with the NASA Decadal Survey Surface Biology and Geology imaging spectrometer mission, which includes as a core measurement snow albedo and its controls (National Academies of Science, 2018). These visible through shortwave infrared imaging spectrometer retrievals have uncertainties an order of magnitude smaller (Painter et al., 2013) than those from multispectral sensors such as MODIS and will provide a more accurate constraint on the physically-based modeling pursued here.

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Atmospheric Chemistry and Physics
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EGU
Discussions

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